

**Quaternary glaciation history and
glaciology of Jakobshavn Isbræ and
the Disko Bugt region,
West Greenland: a review**

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Keywords

Jakobshavn Isbræ, Disko Bugt, Greenland, Quaternary, Holocene, glaciology, ice streams, H.J. Rink.

Cover

Mosaic of satellite images showing the Greenland ice sheet to the east (right), Jakobshavn Isbræ, the icefjord Kangia and the eastern part of Disko Bugt. The position of the Jakobshavn Isbræ ice front is from 27 June 2004; the ice front has receded dramatically since 2001 (see Figs 13, 45) although the rate of recession has decreased in the last few years. The image is based on Landsat and ASTER images. Landsat data are from the Landsat-7 satellite. The ASTER satellite data are distributed by the Land Processes Distribution Active Archive Center (LP DAAC), located at the U.S. Geological Survey Center for Earth Resources Observation and Science (<http://LPDAAC.usgs.gov>).

Frontispiece: facing page

Reproduction of part of H.J. Rink's map of the Disko Bugt region, published in 1853. The southernmost ice stream is Jakobshavn Isbræ, which drains into the icefjord Kangia; the width of the map illustrated corresponds to *c.* 290 km.

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GODHABN

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Abstract

Weidick, A. & Bennike, O. 2007: Quaternary glaciation history and glaciology of Jakobshavn Isbræ and the Disko Bugt region, West Greenland: a review. *Geological Survey of Denmark and Greenland Bulletin* 14, xx pp.

The Disko Bugt region in central West Greenland is characterised by permanent ice streams, of which Jakobshavn Isbræ is by far the most important. The first thorough studies on the glaciology of the region were conducted over 150 years ago by H.J. Rink, who introduced the terms 'ice streams' and 'Inland Ice'. Rink's work inspired new field work, which has continued to the present, and the long series of observations are unique for an Arctic region.

Cooling during the Cenozoic led to ice-sheet growth in Greenland. A number of interglacial occurrences have been reported from the Disko Bugt region, and during the penultimate glacial stage, the Greenland ice-sheet margin extended to the shelf break. During the last glacial maximum, the ice margin probably extended only to the inner part of the banks on the continental shelf, and large floating glaciers may have been present at this time. During the Younger Dryas cold period, the ice margin may have been located at a marked basalt escarpment west of Disko Bugt.

Disko Bugt was deglaciated rapidly in the early Holocene, around 10 500 – 10 000 years before present (10.5–10 ka B.P.), but when the ice margin reached the eastern shore of the bay, recession paused, and major moraine systems were formed. With renewed recession, the present ice-margin position was attained around 8–6 ka B.P., and by *c.* 5 ka B.P. the ice margin was located east of its present position. The subsequent Neoglacial readvance generally reached a maximum during the Little Ice Age, around AD 1850. This was followed by recession that has continued to the present day.

The relative sea-level history shows a rapid sea-level fall in the early Holocene, and a slow rise in the late Holocene. This development mainly reflects a direct isostatic response to the ice-margin history.

Jakobshavn Isbræ is the main outlet from the Greenland ice sheet. It drains *c.* 6.5% of the present Inland Ice, and produces *c.* 35–50 km³ of icebergs per year, corresponding to more than 10% of the total output of icebergs from the Inland Ice. The velocity of the central part of the ice stream at the front has been around 7 km/year since records began, but has nearly doubled in recent years. Other calf-ice producing glacier outlets in Disko Bugt produce *c.* 18 km³ per year. The large calf-ice production of Jakobshavn Isbræ may have been initiated at about 8 ka B.P. when the glacier front receded from the iceberg bank (Isfjeldsbanken) near Ilulissat. Ice streams in inner and outer Egedesminde Dyb may have been active during the early Holocene and during the last glacial maximum.

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Fig. 1. Map of Greenland showing the localities mentioned in the text.

Introduction

Jakobshavn Isbræ (Sermeq Kujalleq) in Disko Bugt, West Greenland, has been recognised as the king of Greenland glaciers amongst scientists and travellers in the Arctic for many decades. It is generally assumed that none of the other fast-moving outlets of the Inland Ice produce comparable quantities of icebergs. In December 2000, the Greenland Home Rule authority decided to nominate the icefjord in front of Jakobshavn Isbræ together with the surrounding areas for inclusion in the World Heritage List of UNESCO (United Nations Educational, Scientific and Cultural Organisation). The nomination report was submitted in 2003 and 'Ilulissat Icefjord' was included on the World Heritage List at the annual meeting of the World Heritage Committee in June 2004. This volume presents a comprehensive description of the region around Jakobshavn Isbræ, including the 'Ilulissat Icefjord' World Heritage site, and emphasises the importance of the region for glaciological investigations in Greenland.

Jakobshavn Isbræ and Disko Bugt

Preparation of the Ilulissat Icefjord nomination report (Mikkelsen & Ingerslev 2002) involved perusal of the large number of scientific papers and descriptions of Jakobshavn Isbræ, its ice production and the glaciological and Quaternary history of the region. However, the nomination document was a technical report published in a limited number, and following inclusion in the World Heritage List a profusely illustrated book designed for a wider audience was produced and published in separate Danish, English and Greenlandic editions (Bennike *et al.* 2004).

The present volume documents the scientific background for the description and conclusions provided by Bennike *et al.* (2004) in the above book. The historical introduction is followed by sections that focus on the geological history, the special peculiarities of the ice cover, and the present-day status of the glacier. The increasing number of recent publications, and the wide spectrum of scientific investigations of the glacier and its environment, have necessitated an updating of the descriptive section. The opportunity is taken here to discuss and present wider conclusions on the geological history of the ice sheet and its surroundings.

In this volume, Greenlandic place names are used according to the spelling that was introduced in 1973. However, the pre-1973 spelling is retained for established stratigraphic

and other terms introduced prior to this year. The Danish name of the main glacier in Disko Bugt used in most published descriptions is 'Jakobshavn Isbræ' (older version: Jakobshavns Isbræ), derived from the former name for the town on the north side of the fjord (Jakobshavn, now Ilulissat). The authorised Greenlandic name for the glacier is Sermeq Kujalleq ('the southern glacier'), but this place name is also used for several other large outlet glaciers that drain from the Inland Ice into Disko Bugt (Qeqertarsuup Tunua) and Uummanaq Fjord (Uummanap Kangerlua). To avoid confusion with other glaciers in the area with the same name, the former name 'Jakobshavn Isbræ' is retained here for the glacier, a usage that accords with most published descriptions.

The formal present-day name of the icefjord in front of the glacier is Kangia (= 'its eastern part', i.e. east of the town of Ilulissat), but other names are also used such as 'Ilulissat Isfjord', 'Jakobshavn Isfjord' and 'Jakobshavn Icefjord'. The World Heritage List uses the name 'Ilulissat Icefjord' for the entire World Heritage site, which includes land areas and parts of the Inland Ice adjacent to the icefjord.

A few other Danish place names are used in order to avoid confusion or for historical reasons. Greenland place names used are listed in the locality index (Appendix 1) at the end of this treatise; the locations of place names appear on Figs 1, 2.

Setting

The Disko Bugt region including Jakobshavn Isbræ is situated in the central part of West Greenland, with the position of the present front of Jakobshavn Isbræ located at *c.* 69°10'N, 50°W. As the glacier is fed by an extensive sector of the ice sheet, a summary of the general morphological and glaciological features of Greenland (Fig. 3) is provided as a background for the following description. Comprehensive descriptions of the present physiography of Greenland and its relation to the complex geological history are provided by Escher & Watt (1976), Funder (1989) and Henriksen *et al.* (2000). A comprehensive account of the dynamic and climatic history of the Inland Ice is given by Reeh (1989).

The importance of the major ice stream of Jakobshavn Isbræ can best be illustrated by the size of the sector of the Inland Ice that feeds it (Fig. 3). This was estimated at



Fig. 2. Map of the Disko Bugt region showing the localities named in the text. The stippled line indicates the position of Isfeldsbanken – the shoal at the mouth of Kangia.

between 3.7 and 5.8% of the Inland Ice, corresponding to an area of 63 000 – 99 000 km², by Bindshadler (1984). More recent estimates have increased this figure to 6.5% of the ice sheet, i.e. 110 000 km² (Echelmeyer *et al.* 1991). This extensive catchment area accounts for the greater part of the actual ice flow to the Disko Bugt region, based on Zwally & Giovinetto (2001).

Areas and volumes

The total area of Greenland is *c.* 2.2 million km² of which the ice sheet (the Inland Ice) constitutes *c.* 1.7 million km²

(Weng 1995). This latter figure also includes some minor marginal ice caps that, while contiguous with the Inland Ice proper, have their own ice dynamics. These ice caps are situated on highlands (especially in East Greenland) and usually have a thickness of a few hundred metres (Weidick & Morris 1998); their combined area only amounts to a few per cent of the total area of the Inland Ice. The Inland Ice is an approximately lens-shaped body, with a maximum thickness of 3.4 km, and with the highest elevation at Summit of 3238 m a.s.l. (Fig. 3). The Inland Ice rests in a bowl-shaped depression (Fig. 4), which in the central parts is below present sea level due to the depression of the Earth's

crust caused by the load of the ice. Estimates of the volume of the Inland Ice vary from *c.* 2.6 million km³ (Holtzschcher & Bauer 1954) to 2.9 million km³ (Bamber *et al.* 2001; Layberry & Bamber 2001), corresponding to *c.* 7% of the world's fresh water (Reeh 1989).

Mass balance of the Inland Ice

Snow accumulation in the central parts of the ice sheet and loss in the marginal parts govern the mass balance of the Inland Ice. The accumulation is estimated to be 500–600 km³ ice per year, which until 2000 was assumed to approximately match the loss. About half of the loss was ascribed to melting, and the other half to calving. Bottom melting of floating glaciers may reduce the calf-ice production of North and North-East Greenland outlets (Reeh 1989, 1994, 1999).

Although calving glaciers are widespread along the coasts of Greenland, the main annual loss of calf ice from the Inland Ice is concentrated at a rather small number of outlets along its *c.* 6000 km long perimeter. Many of the important calving glaciers are found along the west coast of Greenland, with approximately 84 km³ of the calf-ice production originating from five outlets (Table 1, Fig. 5). Four of these occur along a 300 km stretch of coast in the Disko Bugt – Uummannaq Fjord region.

The calf-ice production of Jakobshavn Isbræ is of particular importance for the mass balance of the ice sheet (Fig. 5). The production has generally been estimated to be about 35 km³ ice per year, corresponding to more than 10% of the estimated total output of icebergs from the Inland Ice, but more recent estimates are higher, around 50 km³ per year in 2003, according to Joughin *et al.* (2004). Jakobshavn Isbræ is considered the most active glacier in Greenland by Legarsky & Huang (2006). The reason for the large ice production is that a major part of the Inland Ice drains towards the central part of West Greenland, especially the relatively low uplands in the interior of Disko Bugt (Fig. 3).

Table 1. Calf-ice production from the five largest outlets in West Greenland

Glacier	Latitude	Production (km ³ /year)	Velocity (km/year)
Jakobshavn Isbræ (Sermeq Kujalleq)	69°11'N	<i>c.</i> 35	5–7
Sermeq Kujalleq in Torsukattak	70°00'N	8–10	2.6–3.5
Sermeq Kujalleq (Store Gletscher)	70°20'N	14–18	4.2–4.9
Rink Isbræ (Kangilliup Sermia)	71°45'N	11–17	3.7–4.5
Gade Gletscher	76°20'N	<i>c.</i> 10	

Sources: Bauer *et al.* (1968a); Carbonell & Bauer (1968); Weidick (1995).

With regard to the dynamics of glaciers, determinations of movement and calf-ice production over long time spans are rare. For Jakobshavn Isbræ and Sermeq Avannarleq in Torsukattak velocity measurements go back to 1875. It appears from scattered velocity records for the Disko Bugt

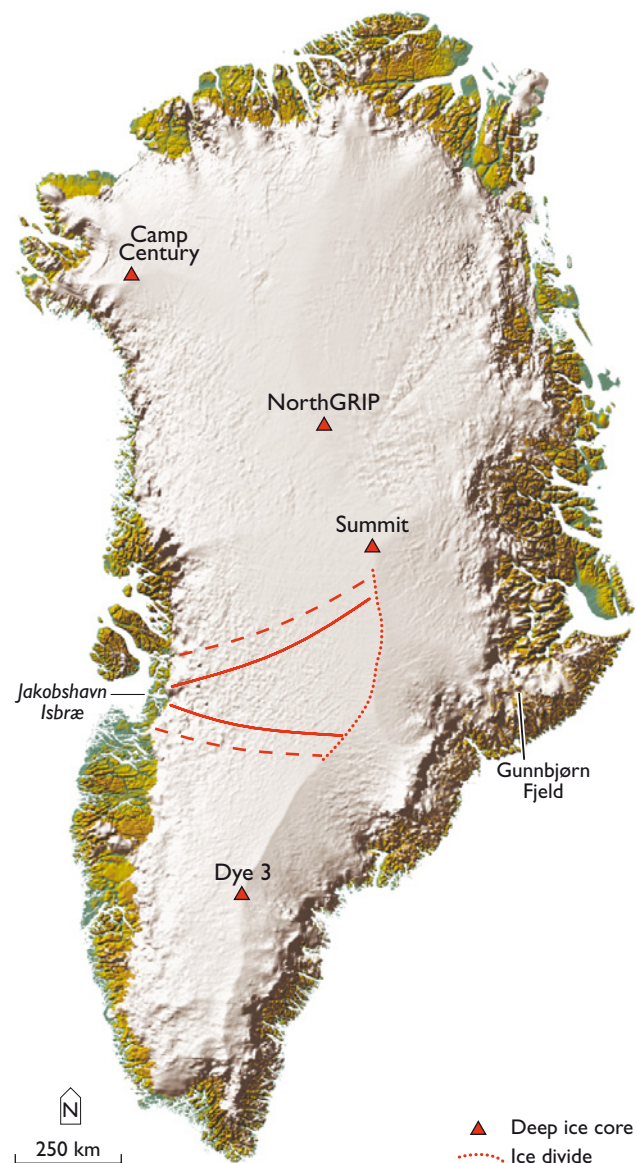


Fig. 3. Map of Greenland showing the location of Jakobshavn Isbræ and the deep cores on the Inland Ice. The approximate ice drainage area to Jakobshavn Isbræ is shown with the solid red line, and the ice drainage area to the entire Disko Bugt region with the dashed line; these are based on the flow-line map of Zwally & Giovinetto (2001). The dotted red line shows the trend of the ice divide. Summit (3238 m a.s.l.) is the highest point on the ice sheet, and Gunnbjørn Fjeld (3693 m a.s.l.) is the highest mountain in Greenland. Original map base by Simon Ekholm, reproduced with the permission of Kort & Matrikelstyrelsen (KMS) [National Survey and Cadastre], Copenhagen.

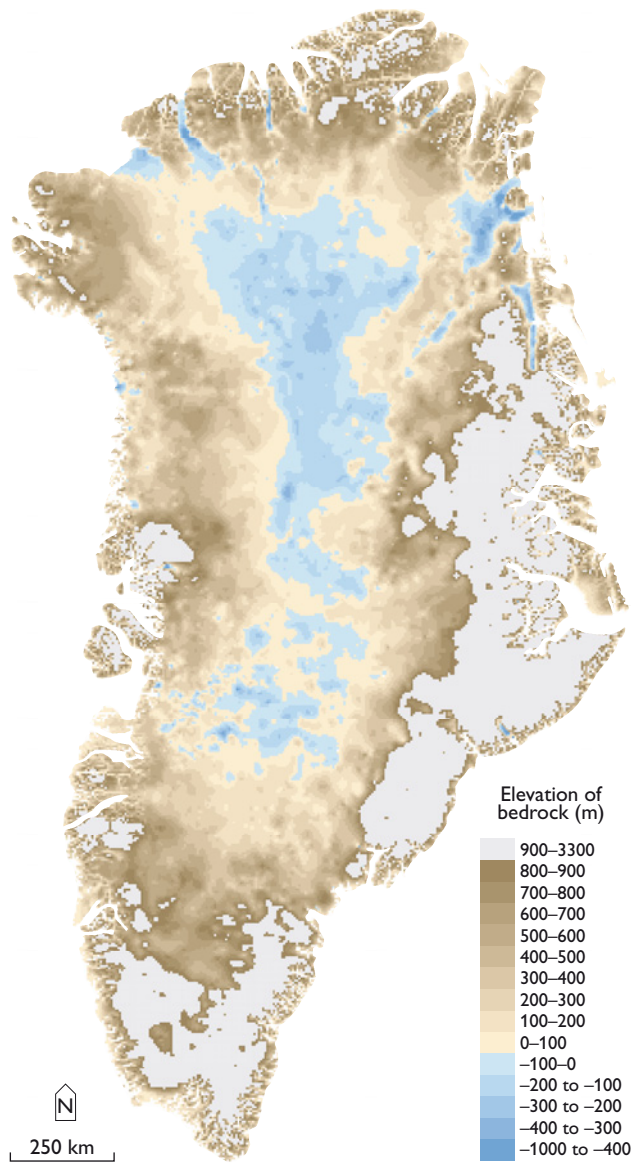


Fig. 4. Bedrock topography below the Inland Ice, from Bamber *et al.* (2001) and Layberry & Bamber (2001); data provided by the National Snow and Ice Data Center DAAC, University of Colorado, Boulder, USA. Elevations are not corrected for the present load of the glacier ice (cf. Fig. 17). The map shows the 'channel' connecting the subglacial central basin to Jakobshavn Isbræ and the Disko Bugt region.

region from the last part of the 19th and the 20th century that the glaciers have maintained a rather permanent rate of flow.

In other parts of Greenland, outlets from the ice sheet show pulsating or surging behaviour, such as Harald Moltke Bræ (Ullip Sermia) in North-West Greenland (Mock 1966),

Storstrømmen in North-East Greenland (Bøggild *et al.* 1994) and Eqalorutsit Killiit Sermiat in South Greenland (Weidick 1984). More recent examples are provided by Rignot & Kanagaratnam (2006).

Over the past decades, aircrafts and satellites are increasingly being used to monitor the Inland Ice. This has led to more detailed observations on changes of velocity, calf-ice production, ice elevation and frontal positions. This development coincided with dramatic changes in the marginal parts of the ice sheet. Marked thinning and recession have been reported for many outlets in Greenland (Rignot & Kanagaratnam 2006). The velocity of many glaciers has increased, and the ice-sheet mass deficit changed from 90 to 220 km³ per year between 1996 and 2005. In East Greenland, Kangerlussuaq Gletscher (Fig. 1) accelerated 210% from 2000 to 2005, and its front receded 10 km. With its velocity of 13–14 km/year it is now the fastest glacier in Greenland. Helheimgletscher farther south accelerated 60% and receded 5 km. In 2001, the velocity of this glacier was measured to be *c.* 8 km/year (Thomas *et al.* 2001). In West Greenland, Narsap Sermia accelerated by 150% from 2000 to 2005 while Jakobshavn Isbræ accelerated by 95% and receded *c.* 10 km. The velocity of Jakobshavn Isbræ was 12.6 km/year in 2003 (Joughin *et al.* 2004).

The marked flow-velocity increase is considered to be related to global warming, which leads to increased melting and sliding in the marginal parts of the ice sheet (Krabbill *et al.* 2000, 2004). A thinning of around 2 m per year is reported for marginal parts of the Inland Ice, which is scarcely matched by a snow accumulation increase of 5–6 cm per year in the central parts of the ice sheet (Alley *et al.* 2005; Johannesen *et al.* 2005; Dowdeswell 2006; Rignot & Kanagaratnam 2006). However, whereas Jakobshavn Isbræ has maintained the high frontal velocity for several years, the velocity increase of other glaciers (Kangerlussuaq Gletscher and Helheimgletscher in eastern Greenland) seems to have been a short-lived event as the velocity and discharge have decreased since 2006 (Howat *et al.* 2007; Truffer & Fahnestock 2007).

The net loss of ice from the Greenland ice cover plays an important role in global sea-level rise, and therefore more detailed investigations of the causes for the marked changes in Greenland are required to assess and model ongoing and future changes. The recently observed changes may lead to a new stable situation, but if the changes continue they may eventually lead to the disappearance of the Inland Ice (Alley *et al.* 2005). However, we note that the Inland Ice did not disappear during the Eemian, even though temperatures were around 5°C higher than at the present.

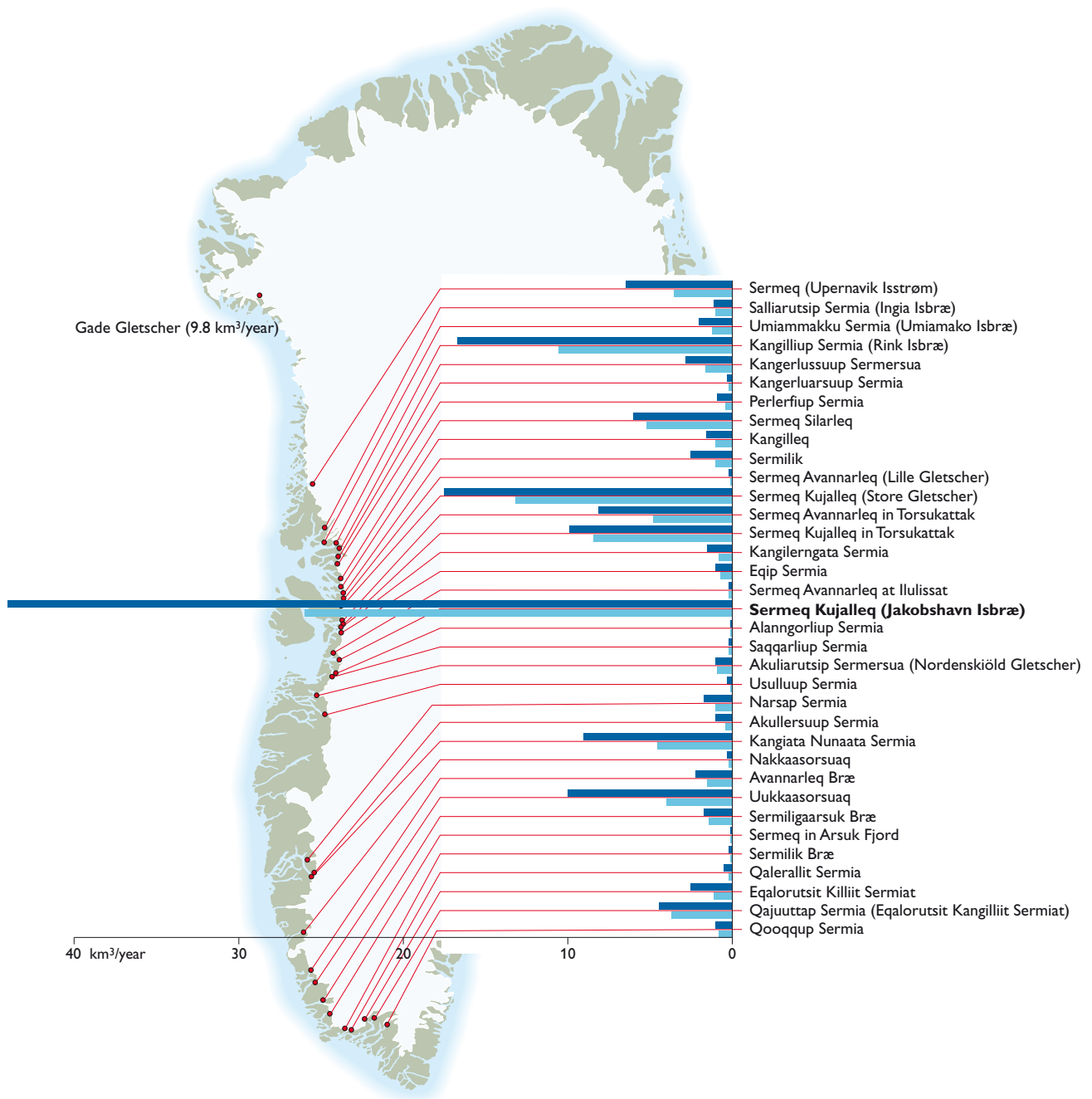


Fig. 5. Estimated maximum (dark blue) and minimum (pale blue) calving ice production from glaciers along the west coast of Greenland. Jakobshavn Isbræ (Sermeq Kujalleq) is clearly in a class of its own (from Weidick *et al.* 1992). The data are based on Bauer *et al.* (1968a) and Carbonell & Bauer (1968), and hence do not consider the dramatic change in calving ice production after *c.* A.D. 2000.

Gross features of the coastland and major drainage of the ice sheet

At the present day, the Inland Ice is separated from the sea by a coastal strip of more or less ice-free land, which is up to 300 km wide. The outlets from the Inland Ice to the sea are, for the most part, restricted to fjords that cut through the coastal strips of ice-free land. Exceptions are found in

North-West Greenland, in Melville Bugt (Qimusseriarsuaq) and in the Kane Basin, where large parts of the ice sheet margin reach the sea. The high alpine mountains of eastern Greenland, with peaks up to 3693 m (Gunnbjørn Fjeld at 68°55'N, 29°53'W, Fig. 3) form a topographic barrier that is in contrast to the widespread hilly uplands of western Greenland. The general topographic fall from this high

mountainous barrier is to the west, and thus the main ice drainage is towards the west.

Offshore Greenland, continental shelves and slopes of variable width border the deep sea. The width of the shelf varies from 50 to 300 km, with the greatest extent off Jøkelbugten in North-East Greenland, off Kangerlussuaq in South-East Greenland, and off Disko Bugt in West Greenland. The water depth over the shelf is mainly between 100 and 300 m, except off South-East Greenland where large shelves areas are found at 300–400 m below sea level. The total area of the continental shelf is estimated to be *c.* 825 000 km² (Henriksen *et al.* 2000). Sinuous troughs, 600–1000 m deep, cut through the shelf at intervals and extend from fjord systems or depressions to the continental slope. Their origin is connected with drowned valleys of Neogene age (Pelletier 1964) that were over-deepened and transformed by glacial erosion during the ice ages. Quaternary marine sediments and till cover large areas of the shelf. Ridges and areas of coarse sediments on the banks are interpreted as marginal deposits that were formed at advanced positions of the ice-sheet margin during the ice ages.

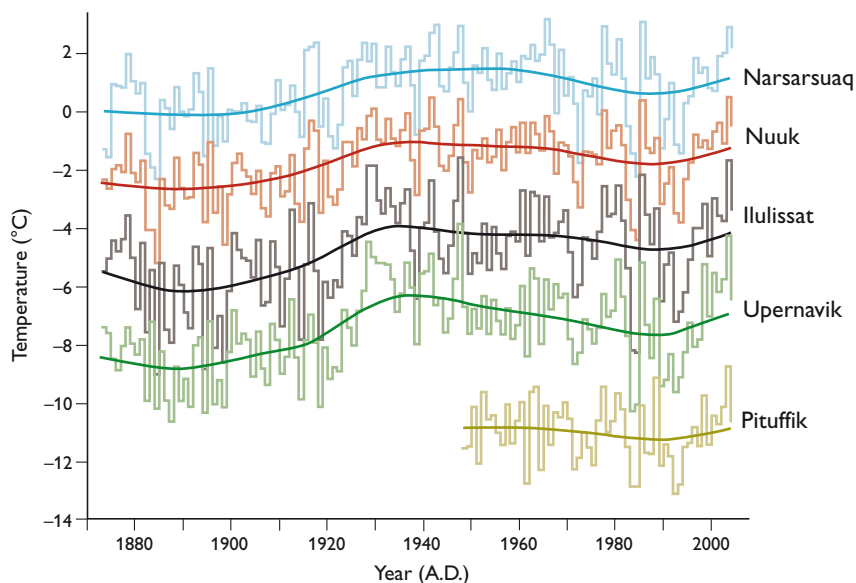
Present climate

A high pressure system usually prevails over the Greenland Inland Ice at the present day. The high landmass of Greenland tends to split eastward-moving low pressure systems approaching from the south-west into two separate

depressions, one travelling north along West Greenland and one travelling up along the east coast of Greenland. In addition, a number of depressions also approach West Greenland from the Hudson Bay region; most precipitation in West Greenland is related to the passage of low pressure systems (Hansen 1999). The climate along the outer coast in eastern and southern Greenland is affected by the East Greenland Current that transports large amounts of sea ice from the Arctic Ocean down along the coast of East Greenland, around Kap Farvel, and northwards up to the Paamiut area at *c.* 62°N on the west coast. Due to the relatively warm current washing certain stretches of West Greenland, parts of the coastal region are open for ship traffic, even in winter. The Disko Bugt region and areas to the north are usually covered by sea ice during the winter. In the summer, western Greenland up as far as the Thule area is nearly ice-free, with the exception of calf ice from the Inland Ice outlets (Buch 2002).

Precipitation in Greenland shows a marked general decrease from south to north. More than 2500 mm/year is recorded in southern Greenland, decreasing to less than 150 mm/year in the interior of Peary Land in northern Greenland (Reeh 1989). In all parts of Greenland, a decrease in precipitation is apparent from the outer coast to areas near the ice-sheet margin. The latter areas are characterised by a continental type of climate; mean annual precipitation at Kangerlussuaq airport in West Greenland at 67°N is around 150 mm, and at Ilulissat at 69°13'N it is around 270 mm (Cappelen *et al.* 2001). On the western slope of the Inland Ice at around 70°46'N, observations indicate

Fig. 6. Annual mean temperatures between 1873 and 2004 for five stations in West Greenland, including Ilulissat (redrawn from Cappelen 2005). The smooth curves were created using a Gauss filter; locations of the stations shown on Fig. 1.



a mean annual precipitation of *c.* 600 mm that is thought to be related to the relatively easy access of humid air masses passing through Disko Bugt (Reeh 1989); this results in lowering of the glaciation level in this region (Humlum 1985, 1986). The average accumulation for the entire ice sheet is around 31 cm water equivalent annually (Ohmura & Reeh 1991), with a general south to north decrease.

In West Greenland, the mean annual air temperature decreases from +0.6°C at Qaqortoq (61°N) to -3.9°C at Ilulissat (69°N) and -11.1°C at Thule Air Base at 77°N (Cappelen *et al.* 2001; Box 2002). These values are based on data from meteorological stations situated close to sea level. Historical data from these and other meteorological stations show a rise in temperature between *c.* 1880 and *c.* 1950, followed by cooling of 1–2°C between *c.* 1950 and *c.* 1990 and a subsequent marked warming (Fig. 6). Data from a network of automatic weather stations indicate that

the mean annual air temperature for the central parts of the ice sheet was around *c.* 2°C higher during the time period 1995–1999, as compared to the period 1951–1960. At an elevation of 1000–2000 m, the temperature increase was only *c.* 1°C (Steffen & Box 2001).

On the ice sheet itself, the mean annual temperature at the surface varies between 0°C and -32°C, according to elevation and geographical location. The temperature increases with depth in the ice body and can reach the pressure-melting point (-2.6°C for 3000 m thick ice) due to geothermal heat and heating caused by ice deformation. It is generally believed that most glacial erosion takes place near the margins of the ice sheet. However, cold bed conditions are found at high, elevated areas with thin and almost stagnant ice. In such areas, the ice cover will protect the subsurface, rather than erode it (Reeh 1989).

History and exploration

Archaeology

In the Disko Bugt region, the three waves of Palaeo-Eskimo and Neo-Eskimo cultures in West Greenland (Fig. 7) are richly represented by numerous archaeological sites (Larsen & Meldgaard 1958; J. Meldgaard 1983; M. Meldgaard 2004; Gulløv *et al.* 2004). Remains from Palaeo-Eskimo and Neo-Eskimo cultures can be found as far inland as Qajaa, which was situated only a few kilometres from the front of Jakobshavn Isbræ during its maximum extent at around 1850.

The oldest culture, the Saqqaq culture, is dated to the period from *c.* 2500 B.C. to about 800 B.C. At Sermermiut, the deposits of this culture are separated from those of the subsequent Dorset culture by a sterile layer. During the time of the Greenlandic Dorset Culture (also referred to

as the Early Dorset culture, *c.* 800 B.C. to 0 B.C.), a smaller and more scattered population lived in the region around Disko Bugt. At about A.D. 1100, the third (Thule) culture arrived in Greenland and settled in the Disko Bugt region about 100 years later (Gulløv *et al.* 2004). The uninhabited periods between the three cultures have been related to climatic deterioration, but other factors such as over-exploitation of the natural resources may also have played a role. During some periods, the Disko Bugt region was densely populated, judging from the large number of settlement sites – at least compared to other regions in the Arctic.

The well-known former settlement, the Sermermiut prehistoric ‘town’ situated close to present-day Ilulissat, was abandoned in the middle of the 1800s. The establishment of the trade centre at Jakobshavn (Ilulissat) by Jakob Severinsen in 1741 was probably the main factor behind the abandonment of Sermermiut.

Although there was certainly some contact between the Thule Neo-Eskimo settlers that migrated from the north and the Norse people that came from the south, the surviving Icelandic sagas provide no identifiable description of the region around Ilulissat.

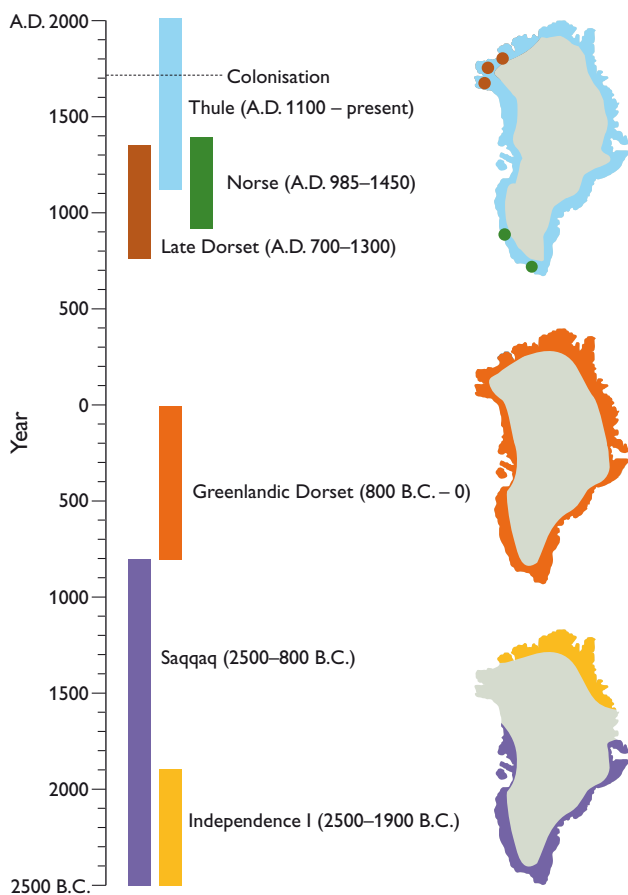


Fig. 7. Chronology of the different archaeological cultures that have colonised Greenland according to Gulløv *et al.* (2004). Modified from Bennike *et al.* (2004).

Discovery, rediscovery, early mapping and descriptions up to *c.* 1845

The first description of the ice cover of Greenland comes from the Norse people, who settled in southern West Greenland after the arrival of Erik the Red in A.D. 986 (Fig. 7; Gad 1967, p. 43). In ‘Kongespejlet’ (‘The King’s Mirror’, here cited from the English translation by Larson 1917, p. 143–144), a description written about 1260 records: “In reply to your question whether the land thaws or remains icebound like the sea, I can state definitely that only a small part of the land thaws out, while all the rest remains under the ice. But nobody knows whether the land is large or small, because all the mountain ranges and all the valleys are covered with ice, and no opening has been found anywhere”. Although this description concerns southern Greenland, it is generally valid for the whole of Greenland.

The Norse settlements were abandoned in the 1400s (Gad 1967; Arneborg 1996). The disappearance of the Norse population is often linked with the onset of the Little Ice Age although other factors, such as exhaustion of natural resources, rising sea level, the political situation in



Faefimile von Paul Egede's Karte von Grönland 1788.

Fig. 8. Poul Egede's map of Greenland from 1788, showing the fictitious strait extending across Greenland from Disko Bugt to the coast of East Greenland (from Nordenskiöld 1886, p. 212). The size of the original map is 294 × 379 mm.

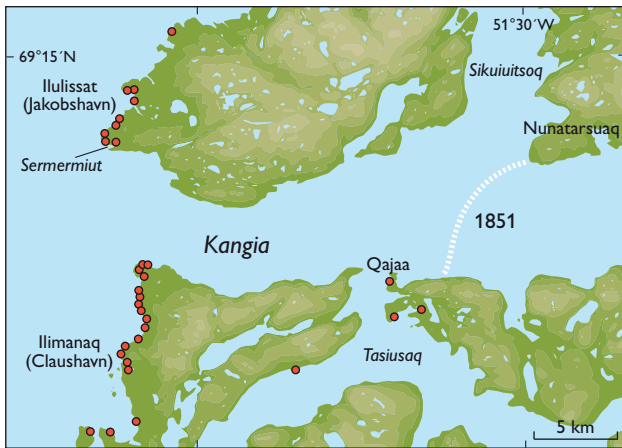


Fig. 9. Map of Kangia and the surrounding region, showing the location of former Inuit settlements (indicated by the red dots) and the approximate position of the front of Jakobshavn Isbræ (Sermeq Kujalleq) in A.D. 1851 (white line). Compiled by the National Museum and Archives of Greenland; modified from Bennike *et al.* (2004).

Scandinavia, the black death (plague), or attacks by Eskimos or Biscay pirates, may also have played a role (Gad 1967).

Apart from sporadic observations by English and Dutch whalers and explorers, there are no descriptions of the physiography of Greenland or its ice cover during the 1500s and 1600s. For Disko Bugt in particular, the survival of numerous place names of Dutch origin (Rodebay, Claushavn, Vaigat etc.) is linked to Dutch whaling activity in the 1600s and the beginning of the 1700s. An early description of the icefjord was published by the Dutch whaler, Feykes Haan, in a navigation guide: “Half a mile north of Sant-Bay is a fjord that is always full of ice, with frightfully tall icebergs, but from where they come is unknown. This fjord is called Ys-Fioert [Icefjord].” (Haan 1719, quoted from Bobé 1916, p. 47; authors’ translation). Short as it is, this description reveals that the conditions of the iceberg bank and the icefjord in the early 1700s were much the same as can be seen today.

Up to the beginning of the 1700s, knowledge of the West Greenland coastal region was poor. In 1721, however, the priest Hans Egede settled in Greenland near Godthåb (now Nuuk, at *c.* 64°10’N), and in subsequent years permanent trading stations and missions spread rapidly. By the end of the 1700s, a network of Danish settlements covered West Greenland from Nanortalik in the south (at 60°N, established 1797) to Upernavik in the north (at 72°N, established 1772). In the Disko Bugt region, Qasiqiannguit (Christianshåb) was established in 1734, Ilulissat (Jakobshavn) in 1741, Asiaiat (Egedesminde) in 1763 and Qeqertarsuaq (Godhavn) in 1773.

With colonisation came the first attempts to undertake a systematic mapping of Greenland (in particular West Greenland) and also an increasing number of local descriptions were made. The development of early mapping is described and illustrated by Dupont (2000). Poul Egede’s map from 1788 illustrates the status of mapping at that time (Fig. 8). Apart from the lack of detail, it should be noted that a channel is depicted connecting Disko Bugt in West Greenland with Kangerlussuaq in East Greenland. This so-called ‘Frobisher Strait’ is an error copied from older maps. Egede commented on the map: “It is said that the strong current that flows continuously from the ice-dome comes from Ollum Længri Fjord” (i.e. from East Greenland; authors’ translation). Egede also stated in another comment to the map: “The entire land is concealed under ice and snow, from Staten Huk to the extreme north” (authors’ translation). ‘Staten Huk’ was the Dutch whalers’ name for Kap Farvel at the southern tip of Greenland. The way this is expressed perhaps indicates some doubt about the alleged channel through Greenland. Egede’s comments about the icefjord, with respect to the strong currents from the glacier may be based on observation, as may his depiction of the icefjord with a length similar to the present.

Historical descriptions of the conditions in Kangia and its surroundings were collated by Larsen & Meldgaard (1958), supplemented by Georgi (1960a, b). Both stressed the records that the icefjord was more ice-free in the early 1700s than in subsequent times. This conclusion is essentially based on a letter from the manager of Jakobshavn, Hans Rosing, from 1831. It states in translation: “An old woman still living here [at Jakobshavn] knew, when she was young, another old woman who told her that, when she was a young girl, there were practically no icebergs in the fjord, but the water was so open that the Dutchmen with their large vessels went into the fjord. Up along the coast of the fjord the Greenlanders lived in their tents, and in the winter there were houses of which ruins still can be seen at least one ‘miil’ [approx. 10 km] from the mouth” (Larsen & Meldgaard 1958, p. 24–25). Historically these events can be related to a time just before 1740. This evidence can be combined with descriptions from a visit to the ice margin on 16 January 1747 by P.O. Walløe (Bobé 1927). Walløe described the advancing ice margin in so much detail that there can be little doubt of the validity of his observations, which were probably made close to Qajaa (Fig. 9). Qajaa was abandoned in the middle of the 1700s, presumably because of the advance of Jakobshavn Isbræ in the 18th and first half of the 19th century. Rink (1875, p. 15) provides further confirmation: “...proof that Jakobshavn-Fjorden was earlier accessible further in, is given by the remains of an older dwelling site in a location that can-

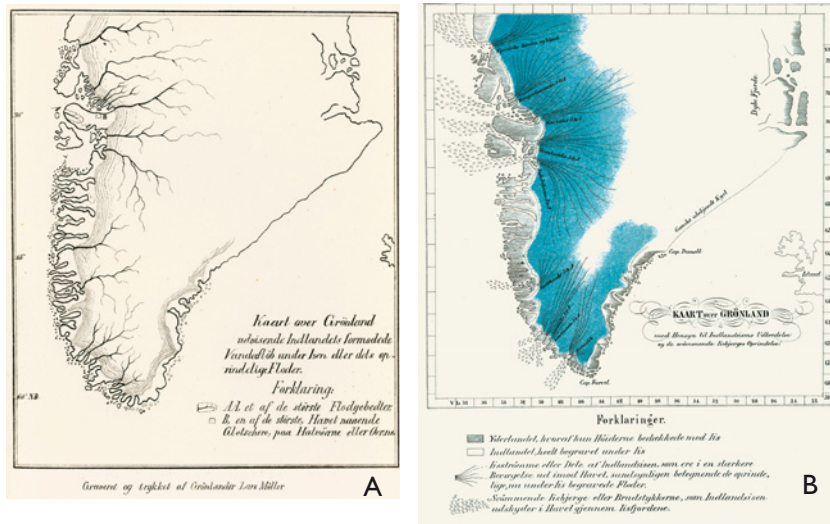


Fig. 10. A: Rink's concept of the drainage of the interior of Greenland by rivers (Rink 1862). The outline of East Greenland was then not well known, and the depiction of the area north of Ilulissat in West Greenland is also imprecise. B: Rink's map showing the extent of the Inland Ice, and the large ice streams that drain into Disko Bugt and Umannaq Fjord (Rink 1857). Smaller glaciers in southern Greenland are also indicated. C: Segment of Rink's original map of the interior part of Disko Bugt showing the positions of the ice streams draining into Kangia and Torsukattak (Rink 1853).

not be reached now due to ice” (authors’ translation). Together these observations lead to the conclusion that the glacier could have been in a rather retracted position in 1747, but perhaps in an initial advancing state. The glacier may well have been just as productive then as it is now, as implied by Haan’s description of icebergs on the iceberg bank in 1719. An initial advance in the beginning of the 1700s may only have led to a slight reduction in calf-ice production.

Unfortunately, the observations on the advancing ice margin around Jakobshavn Isbræ up to *c.* 1850, backed up by records of vegetated soil or ruins buried by the advancing ice, generally lack information as to the exact location at which they were made.

These observations from the 18th and the beginning of the 19th century, however, form valuable contributions to the growing interest in the changes of ice cover. The observations on the Greenland ice cover from the 18th century were compared by Crazz (1770) with the status of European and American glaciers, which were then also advancing. This provided impetus for more detailed mapping and descriptions of the Greenland ice cover in the following century, when it was first appreciated that glaciers could be regarded as a kind of ‘climatoscope’, with their recessions and advances reflecting alternating warm and cold periods. In Greenland, systematic observations and descriptions of the ice cover were initiated by Hinrich Johannes Rink and recorded in a series of outstanding publications.

Hinrich Johannes Rink (1819–1893)

Originally educated in chemistry and physics (Oldendow 1955), H.J. Rink carried out geological investigations in parts of northern West Greenland between 1848 and 1852. From 1853 to 1858 he was trade manager in Qaqortoq (Julianehåb) and Nuuk (Godthåb), and from 1858 to 1868 he was inspector of the Royal Greenland Trade in South Greenland. He travelled over large parts of West Greenland. One of his ambitions was to produce a pioneer map of West Greenland, incorporating his own mapping with that of older sea charts, the mapping of the missionary Samuel Kleinschmidt (1814–1886; see Wilhjelm 2001, p. 144–152), and map sketches made by local hunters. He devoted much of his own mapping efforts to the interior, eastern ice-free fjords and land areas that were largely unknown at the time. During his travels he visited, described and mapped extensive areas of the icefjords and their glaciers. Rink stayed in Ilulissat (Jakobshavn) over the winter of 1850–1851, and sailed to the Inland Ice margin at the fjord Paakitsoq, north of Ilulissat, in October 1850. He also

travelled by dog sledge to the southern side of the front of Jakobshavn Isbræ in April 1851 (Rink 1857; vol. 1) and visited the north side of the glacier in May the same year (von Drygalski 1897, p. 129). His approximate determination of the frontal position of Jakobshavn Isbræ is shown in Fig. 9, as depicted in later reviews. The original version of Rink’s map is shown in Fig. 10.

Rink’s observations of the frontal position of Jakobshavn Isbræ in 1850–51 were the first in a series of observations that have now extended over more than 150 years. The long series of observations of the frontal changes and the velocity of Jakobshavn Isbræ (from 1875), and the long series of continuous meteorological records at Ilulissat (since 1873), are unique for an Arctic area.

Rink was the first scientist to appreciate the immense extent and special form of the ‘ice plain’, covering the entire interior region of Greenland. Rink called it ‘Indlandsisen’ [the Inland Ice] following a suggestion by the Danish scientist Japetus Steenstrup. It was clear to Rink that this large body of ice was quite different from the local glaciers hitherto described from other parts of the world. In Europe a new idea was emerging at this time, namely that extensive ice sheets had covered large parts of northern Europe in the past. Rink’s demonstration of an extant, immense ice sheet in Greenland was sensational news, and provided critical support for arguments that such an ice cover had once existed in Europe. While it is true that earlier mapping and descriptions, such as Poul Egede’s map (Fig. 8), had indicated the presence of extensive ice in Greenland, it was Rink’s detailed observations and descriptions that provided documentary evidence, and provided the basic background for the numerous subsequent glaciological investigations in Greenland.

In his attempts to understand the origin and dynamics of the Inland Ice, Rink also ventured into many of the problems that concern the mechanics of calf-ice production. Rink’s main thesis was that just as precipitation over land areas is drained by rivers, the ice streams deriving from the interior of Greenland drain the snow and ice. Ice streams act as ‘rivers’ in a surrounding area of ‘quiet’, dynamically less active ice (Fig. 10; Rink 1862).

Rink estimated the production of calf ice on the basis of glacier size and the quantity of ice in the fjords. On these criteria, Rink (1857, 1862) recognised five ice streams (‘isstrømme’) of ‘the first order’, namely Jakobshavn (69°10’N), Tossukatek (69°5’N), Den større Kariak (70°25’N), Den større Kangerdlursoak (71°25’N) and Upernivik (73°N; Fig. 10); the spelling used by Rink is retained here. Rink estimated the catchment area for each of the five ice streams to be at least *c.* 50 000 km² (*c.* 1000 Danish square miles) (Rink 1875, p. 15), a clear indica-

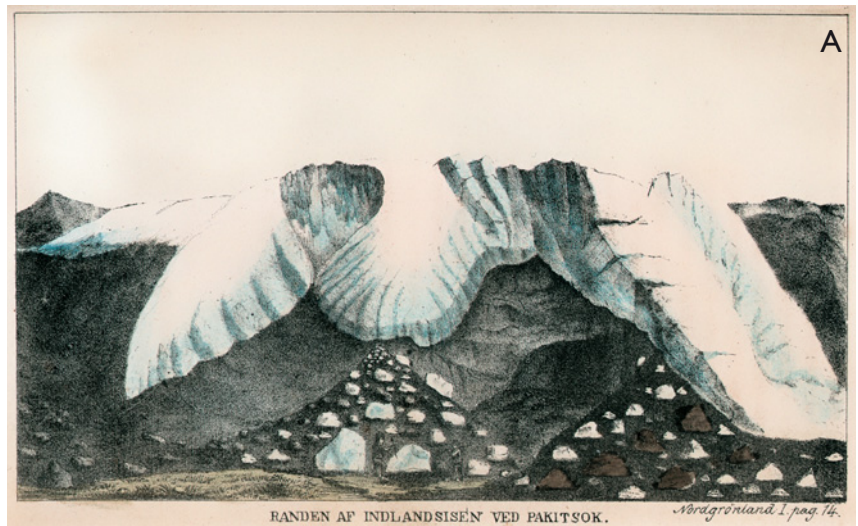


Fig. 11. The margin of the Inland Ice at Paakitsoq north of Kangia; the central hill is about 80 m high. The three images (A: lithograph, H.J. Rink, 1850. B: photograph, A. Weidick, 1961. C: photograph, H.H. Thomsen, 1987) illustrate changes in the ice margin at this site. In 1850, the glacier front was advancing, reaching a maximum around 1880. Since then a gradual thinning has taken place as can be seen from the photographs from 1961 and 1987.



Fig. 12. The lobe of the Inland Ice margin at Paakitsoq, seen from the south-west; the central hilltop is about 150 m high. A: R.R. Hammer's drawing from 1883, which marks the maximum historical extent of the glacier lobe. The photographs from 1961 (B: A. Weidick) and 1987 (C: H.H. Thomsen) illustrate the progressive thinning of the ice margin.

tion that he recognised the extraordinary size of the Greenland ice sheet. However, while recognising the large calf-ice production from these major outlets, Rink did not appreciate the unique status of Jakobshavn Isbræ at Ilulissat. This was first recognised in the second half of the 19th century, during investigations on the rate of movement. Several illustrations in Rink's papers provide details of the extent of the glacier cover in the middle of the 1800s, of which Figs 11 and 12 are examples showing Paakitsoq, north of Ilulissat.

Observations and mapping of the ice margin around Disko Bugt

As mentioned above, interest in Greenland glaciers increased during the second half of the 19th century. An increasing number of scientists visited the ice margin around Disko Bugt, and most of these provided descriptions of the front or the Inland Ice margin around Jakobshavn Isbræ. A chronological list of the most significant visits after Rink is given below.

1867. The British mountaineer Edward Whymper made a visit to the area around Qajaa in Kangia. He noted that the glacier ice was too crevassed to allow passage to the Inland Ice (Whymper 1873; Nordenskiöld 1886, p. 121).

1870. A.E. Nordenskiöld visited West Greenland, and described the icefjord and the front of Jakobshavn Isbræ. He could not determine the boundary between the front of the glacier and the calf ice in the fjord (Nordenskiöld 1871; Engell 1904). His visit to Jakobshavn Isbræ was undertaken in connection with one of the early attempts to visit the interior of Greenland, and Nordenskiöld also visited Nordenskiöld Gletscher (Akuliarutsip Sermerssua), *c.* 90 km south of Jakobshavn Isbræ. Here, from the head of Arfersiorfik fjord, he led a reconnaissance expedition that reached 56 km into the ice sheet at an altitude of 670 m a.s.l. In 1883, from the same starting point, another party led by Nordenskiöld reached *c.* 350 km into the Inland Ice to an altitude of 1947 m. Surface features of the ice, such as cryoconite holes (meltholes) and glacier spouts, were described.

1875. A. Helland visited Jakobshavn Isbræ in July 1875 and conducted the first measurements of the rate of movement, which were published together with a description of the position of the front of Jakobshavn Isbræ (Helland 1876). He also measured the movement of Sermeq Avannarleq in Torsukattak icefjord, *c.* 100 km north of Jakobshavn Isbræ, and described the surface of the ice margin at Paakitsoq.

1879. K.J.V. Steenstrup (1883a) also visited Sermeq Avannarleq in Torsukattak icefjord, and measured the rate of movement of this glacier in May 1879 and 1880.

1879. R.R.J. Hammer mapped the entire fjord system around Kangia, including the glaciers at the heads of the tributaries of Sikuiuitsoq and Tasiusaq (Hammer 1883). Hammer also determined the position of the front of Jakobshavn Isbræ in September 1879.

1880. R.R.J. Hammer repeated his visit to Jakobshavn Isbræ in March and August 1880. A winter advance of *c.* 1 km and a subsequent summer recession of *c.* 2 km were recorded. The rate of movement was determined from the southern edge of the glacier front to its central part. Hammer found that the movement was not uniform, and could find no relationship between air temperature and glacier velocity. The position of the front was observed to be very variable, and Hammer concluded that large icebergs were released by fracturing of the ice front due to the buoyancy of the floating part of the glacier. Hammer also tried to evaluate the hydrographic conditions of the icefjord.

1883. R.R.J. Hammer mapped the eastern parts of Disko Bugt between *c.* 68°30' and 70°N (Hammer 1889), and made a sketch of Jakobshavn Isbræ while attempting to determine the glacier recession since 1851.

1883. A.E. Nordenskiöld revisited Nordenskiöld Gletscher (see 1870 above).

1886. Robert E. Peary made an attempt to reach the interior of the Inland Ice from a starting point at Paakitsoq, 40 km north of Kangia. Accompanied by C. Majgaard he reached a point *c.* 185 km into the ice sheet at an altitude of almost 2300 m a.s.l. Peary also described the landscape of the ice margin (Peary 1898).

1888. S. Hansen made a sketch map of the front of Jakobshavn Isbræ for the Royal Danish Sea Chart Archive (Engell 1904). According to Engell, a photograph from this visit showed the frontal position of Jakobshavn Isbræ to lie farther to the east than indicated on Hansen's 1888 map.

1891. In June 1891 the German polar explorer Erich von Drygalski visited the north side of Kangia.

1893. von Drygalski visited the area to the south of the front of Jakobshavn Isbræ in February 1893, and plotted the frontal position on a map, although it was difficult to determine the position precisely (von Drygalski 1897).

1902. M.C. Engell measured the frontal position of Jakobshavn Isbræ in July 1902, and showed that the recession had continued (Fig. 13). Velocity determinations were similar to those previously measured (Helland 1876; Hammer 1883; Engell 1904). In addition, Engell made an extensive description of the whole fjord region, and produced detailed maps of the front of Jakobshavn Isbræ and

the ice margin at the head of Orpissooq fjord in the south-eastern corner of Disko Bugt.

1903. M.C. Engell again visited Jakobshavn Isbræ in July 1903, but could only determine the frontal position with some uncertainty; it appeared to be *c.* 350 m farther west than in 1902. Engell observed the release of a large iceberg (Engell 1910).

1904. M.C. Engell made his third visit to the front of Jakobshavn Isbræ in the summer of 1904 although, the position of the glacier front was not determined (Engell 1910). He mapped the interior of Disko Bugt from *c.* 68° 45' N (head of Tasiusaq fjord) to *c.* 70° 05' N (Torsukattak icefjord).

1912. Alfred de Quervain and Paul-Louis Mercanton made a west-to-east crossing of the Inland Ice from Eqip Sermia in Disko Bugt to Ammasalik in East Greenland. A description of the ice margin around Eqip Sermia, *c.* 70 km north of Ilulissat, was later published (de Quervain & Mercanton 1925).

1913. Johan P. Koch and Alfred Wegener made an east-to-west crossing of the Inland Ice during their 1912–1913 expedition. In August 1913 the two scientists visited Jakobshavn Isbræ, and their determination of the frontal position indicated a significant recession since 1902 (Koch & Wegener 1930).

1929. During the 1929 preparations for the German Alfred Wegener Expedition to the Inland Ice in 1930–1931, reconnaissance for an alternative route to the ice sheet was made in Disko Bugt. During May and June, the equipment was tested and ablation measured on a route that extended 150 km from Eqip Sermia north-eastwards into the Inland Ice reaching an altitude of 2090 m. Jakobshavn Isbræ was visited in September 1929, and the front position and the velocity determined. The frontal positions of Eqip Sermia and the glaciers in Torsukattak icefjord were also described (Georgi 1930; Wegener *et al.* 1930).

1931/32. The position of the glacier front of Jakobshavn Isbræ was depicted on the first 1:250 000 scale map sheet of the Jakobshavn area, issued by the Geodetic Institute, Copenhagen.

1934. Martin Lindsay started a traverse of the Inland Ice from Eqip Sermia, from the same starting point that de Quervain and Mercanton used in 1912. The main objective of the three-man expedition was to explore the mountainous region south-west of Scoresby Sund in East Greenland (Fristrup 1966).

1936. Eigil Knuth and Paul-Emile Victor participated in a French/Swiss/Danish expedition that crossed the Inland Ice from the ice margin *c.* 80 km south of Ilulissat to Ammasalik in East Greenland. A description of the ice-margin landscape was published by Knuth (1937).

After 1936. During and after the World War II, there were rapid developments in aerial photography techniques. The increase of information can be illustrated by the archive of aerial photographs of the region available at the National Survey and Cadastre in Copenhagen, dating from the years 1942, 1946, 1948, 1953, 1957, 1958, 1959, 1964 and 1985.

Large-scale glaciological projects after World War II

After World War II, a series of major investigations of the Inland Ice was carried out by military and scientific organisations. A detailed account of this work is outside the scope of this bulletin, and the investigations are only briefly mentioned here as a background for the research around Disko Bugt itself. Reviews of the history of exploration of the Greenland ice sheet and the significant results achieved are given by Fristrup (1966), Reeh (1989) and Dansgaard (2004).

The new era of scientific expeditions was initiated by the Expéditions Polaires Françaises (EPF 1948–1953), which continued the work of the German Alfred Wegener Expedition of 1929–1931. EPF worked along an east–west profile from coast to coast over the central part of Greenland, and in addition to geodetic and geophysical work carried out mass-balance measurements along the route. Detailed mapping of the ice sheet margin and descriptions of the area around Eqip Sermia were also made.

The successor of EPF was the Expédition Glaciologique Internationale au Groenlande (EGIG 1957–1960), an international collaboration between Austria, Denmark, France, Germany and Switzerland. EGIG included in its programme a project that aimed at photogrammetric determinations of the rate of movement and estimates of calving production of all outlets from the Inland Ice that drain into Disko Bugt and the Uummannaq Fjord. Vertical aerial photographs were used for this work. In 1957, flights were repeated at intervals of four to five days (Bauer *et al.* 1968a), and in 1964 for the same glaciers at intervals of about two weeks (Carbonell & Bauer 1968). The project confirmed the paramount role of Jakobshavn Isbræ with respect to calving production, compared to other productive glaciers in Greenland (Fig. 5). Variations in the frontal position of Jakobshavn Isbræ were also determined (Fig. 13).

The EGIG work continued after 1960 with follow-up projects. One important task was to determine the thickness of the Inland Ice, and hence also the elevation and topography of the landscape below the ice, by means of airborne radar. Prior to this initiative, the thickness had only

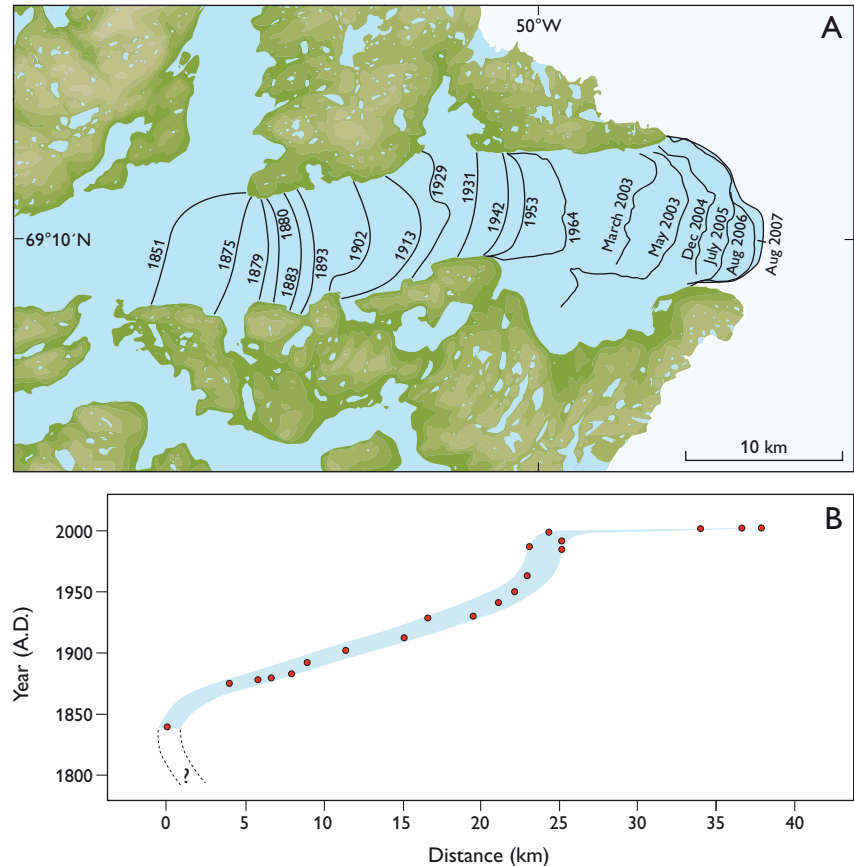
been measured along a few profiles traversing the Inland Ice (Holtzschler & Bauer 1954), using seismic, gravimetry or radar techniques. The Technical University of Denmark (DTU) in collaboration with the US National Science Foundation and the Scott Polar Research Institute in England modified the radar technique for use in aircraft. During six seasons between 1968 and 1976, more than 60 000 km of profiles were flown, and for the first time a large-scale map of the landscape under the Inland Ice became available (Gudmandsen & Jakobsen 1976). At that time, this subsurface map was a major breakthrough with significant implications for the understanding of ice dynamics, Quaternary geology and geomorphology. However, details such as the continuation of the deep fjords under the present Inland Ice margin could not be resolved, because radar waves could not penetrate the chaotic, crevassed ice of the ice streams. This problem was solved at Jakobshavn Isbræ by applying seismic methods (Clarke & Echelmeyer 1996).

American military groups developed ways of travelling on the ice sheet, erected and maintained Inland Ice stations, and were also directly involved in scientific operations. The latter included the first deep drilling through the ice

sheet at Camp Century in 1963–1966 (Langway 1970; Langway *et al.* 1985). The first systematic mapping of snow accumulation over the entire Inland Ice was carried out by Cold Regions Research and Engineering Laboratory, Corps of Engineers, US Army (CRREL) in 1952–1955 and 1959–1960, and led to a division of the Inland Ice surface into dry snow, percolation facies and wet-snow facies according to altitude (Benson 1959, 1962, 1994, 1996; Ragle & Davis 1962). More recently, the National Aeronautics and Space Administration (NASA) has been responsible for the development and maintenance of the present satellite and airborne monitoring system (Sohn *et al.* 1997a; Williams & Hall 1998; Thomas *et al.* 2001). This system provides detailed information on current changes of surface elevation, volume and rate of movement of the Greenland ice sheet.

The development of the global positioning system (GPS) and airborne and satellite-based altimetry has resulted in a vast expansion of data. The main focus has been on assessing the impact of climate change on the mass balance of the ice cover, especially with respect to volume changes and surface movements of the major outlets of the ice sheet (Garvin & Williams 1993). In 1994, NASA conceived a

Fig. 13. A: Frontal positions of Jakobshavn Isbræ from 1850 to the present. Modified from Bauer *et al.* (1968a) and Weidick *et al.* (2004a). B: Recessional curve of the Jakobshavn Isbræ glacier front, up to 1964 based on the positions given by Bauer *et al.* (1968a). The younger parts of the curve are based on satellite information by Stove *et al.* (1983), Sohn *et al.* (1997a, b), later Landsat, ASTER and MODIS images (Alley *et al.* 2005; Cindy Starr, NASA (personal communication, 2007)) and data from a reconnaissance in 2005 (F. Nielsen, personal communication 2006). The width of the curve depicts the range of seasonal variations in the positions of the glacier front. Note the rapid break-up and recession from 2002.



Program for Arctic Regional Assessment (PARCA), which for a decade has collected mass balance data covering the entire ice sheet, with a back-up of ‘ground-truth’ stations (Abdalati 2001; Abdalati *et al.* 2001). The programme includes the collection of ice-thickness data around Jakobshavn Isbræ (Gogineni *et al.* 2001), where one of the ‘ground-truth’ stations is located. This is the ‘Swiss Station’ established at the equilibrium line near Jakobshavn Isbræ by the Eidgenössische Technische Hochschule Zürich, Switzerland (ETH), and later operated by the University of Colorado (Steffen & Box 2001; Zwally *et al.* 2002).

With respect to subsurface mapping around Jakobshavn Isbræ, some details were added during investigations of the hydropower potential for the towns of Ilulissat and Qasigiannuit 1982–1992 (Thomsen *et al.* 1989; Braithwaite 1993). These investigations collected data for the ‘quiet’ parts of the Inland Ice margin (the areas between the ice streams), using a radar device carried by helicopters. The device was developed by DTU and modified by the Geological Survey of Greenland (Thorning *et al.* 1986; Thorning & Hansen 1987). The most recent radar thickness measurements of the ice sheet (with particularly detailed coverage around Jakobshavn Isbræ) were published by Bamber *et al.* (2001), Gogineni *et al.* (2001) and Layberry & Bamber (2001).

Deep cores from the ice sheet

The constant deposition of snow over the interior parts of the ice sheet results in the compaction of the underlying snow and conversion to glacier ice, which in the course of time flows downwards and outwards towards the margins of the Inland Ice. The discovery of the capability of the ice to preserve information about the climate at the time of snow deposition has given remarkable results with respect to climate change and related geological and atmospheric changes. A number of intermediate and deep ice cores have been recovered from the Greenland ice sheet (Reeh 1989). Of the five deep cores (Fig. 3), the first was made at Camp Century (77°11′N, 61°08′W) in 1964–1966, and had a length of 1390 m (Dansgaard *et al.* 1969; Langway 1970; Langway *et al.* 1985; Reeh 1989). Subsequently, a 2037 m long core was retrieved at Dye 3, where drilling was completed in 1981 (65°11′N, 43°49′W; Langway *et al.* 1985; Reeh 1989). The success of these cores was followed up by two >3000 m deep cores at Summit, on the highest point of the ice sheet, namely the GRIP core (72°34′N, 37°37′W; Johnsen *et al.* 1997) and the GISP2 core (72°35′N, 38°29′W; Grootes *et al.* 1993). The fifth deep core was completed at the NorthGRIP site (75°06′N, 42°19′W) in 2001, and reached a depth of 3001 m (Dahl-Jensen *et al.*

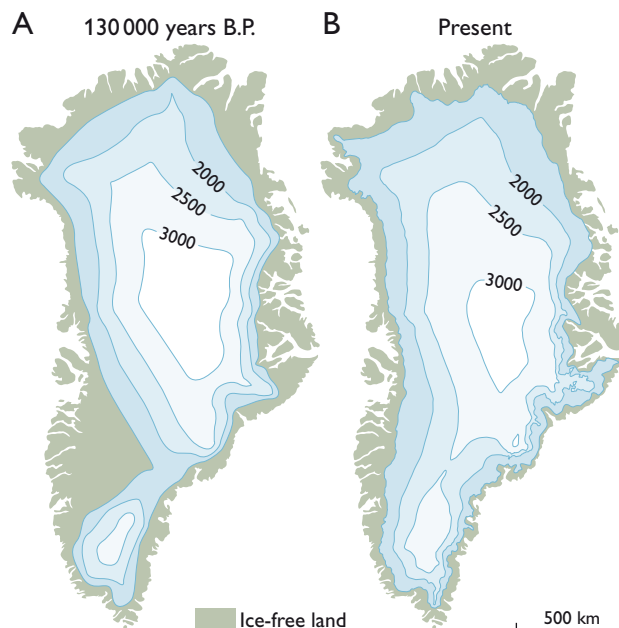


Fig. 14. A: Model of the surface elevation of the Greenland ice sheet during the last interglacial (Sangamonian/Eemian) according to Letréguilly *et al.* (1991a, b). B: Modern-day surface elevation according to Escher & Pulvertaft (1995).

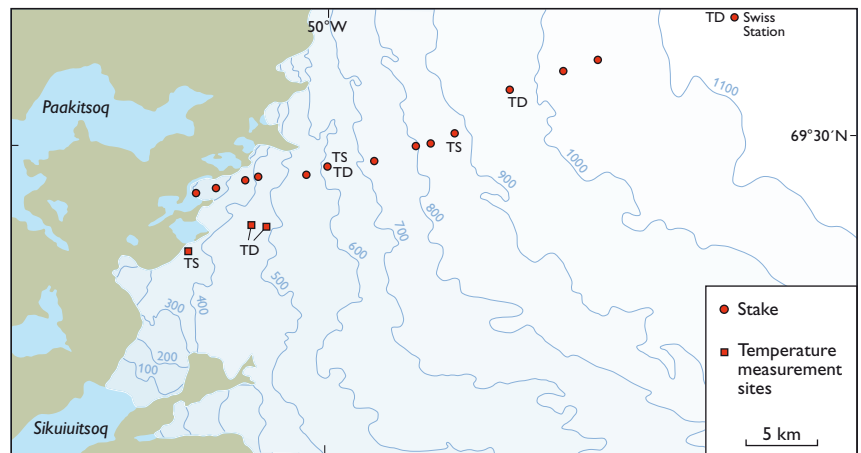
2002). All cores have provided detailed climatic information about the last ice age, and the cores from the central parts of the Inland Ice provide information on the climate up to *c.* 250 000 years back in time.

The data obtained from the ice cores, together with detailed information about the subsurface and surface of the ice sheet, have been incorporated into models of changes of the ice sheet with time, as well as scenarios for the future development of the Inland Ice (Fig. 14; Letréguilly *et al.* 1991a, b; Weis *et al.* 1996; Ritz *et al.* 1997; Huybrechts 2002; Alley *et al.* 2005).

Hydropower and climatic change

The energy crisis in 1973 led to a focus in Greenland on the utilisation of local energy sources, as an alternative to imported oil. In Greenland, hydropower was the obvious potential source of power, but an evaluation required systematic collection of hydrological and glaciological data. As part of a project carried out by the Geological Survey of Greenland (GGU), a series of stations was erected along the Inland Ice margin in West Greenland between *c.* 60° and *c.* 70°N. Data were collected for a hydropower project to serve the town of Ilulissat, and also a project 45 km further south to serve the town of Qasigiannuit. The aim

Fig. 15. The drainage basin at Paakitsoq, north of Ilulissat, where the Sermeq Avannarleq glacier enters Sikuiuitsoq (Fig. 2), showing the positions of stakes used for mass balance calculations. TS and TD mark the positions of shallow and deep thermistor-strings, respectively, that were used for measuring temperatures in the ice. The Swiss Station marks the highest point of the glacier hydrological survey. From Thomsen *et al.* (1991).



of both projects was to determine the amount of meltwater runoff from the ice-sheet margin.

The northernmost station at Paakitsoq, *c.* 40 km north-east of Ilulissat, was started in 1982 (Thomsen *et al.* 1988; Olesen 1989). The studies covered mass-balance measurements along a line of stakes extending from the ice-sheet margin at *c.* 230 m a.s.l. to *c.* 1050 m, close to the equilibrium line (Fig. 15). The studies included drilling of ice holes for measuring ice temperatures and subglacial melt. A detailed map of the subsurface of the ice margin around Jakobshavn Isbræ was also produced (Thomsen *et al.* 1988; Olesen 1989; Weidick 1990).

The data collected from Paakitsoq have not yet resulted in a decision to exploit hydropower for Ilulissat. The annual potential is around 72 GWh for Ilulissat and *c.* 11 GWh for Qasigiannugit (Nukissiorfiit 1995). However, a hydropower plant is now in operation near Nuuk in West Greenland (Kangerluarsunnguaq power plant, production 185 GWh/year), and two other power plants at Qaqortoq/Narsaq in South Greenland and at Ammassalik in South-East Greenland have been established.

The debate on climatic change due to the increasing greenhouse effect and the increased melting of glaciers has stressed the need for data on the actual melting at the ice margin of the Inland Ice. The mass-balance measurements by GGU that started in 1982 in relation to hydropower, were therefore continued in 1990 in collaboration with teams from the Alfred Wegener Institute for Polar and

Marine Research, Germany (AWI) and ETH (Thomsen *et al.* 1991). Energy-balance measurements at the equilibrium line were made from a permanent field station on the ice, and a programme for ice-temperature measurements was also set up in a collaboration between GGU and ETH. The original stake line measured from 1982 was extended from 1100 to 1600 m a.s.l. The ETH programme to study climate, energy balance and the thermal regime covered the years 1990 and 1991 (Ohmura *et al.* 1991), with participants from the Institute of Arctic and Alpine Research (INSTAAR, University of Colorado, Boulder, USA). INSTAAR runs a long-term project that aims to investigate the effects of refreezing of meltwater runoff from the Inland Ice, by studies of snow and ice hydrology, heat transfer, and using modelling (Pfeffer *et al.* 1991).

Satellite and airborne radar and laser altimetry with large-scale coverage is now an important tool for monitoring changes in the geometry and dynamics of the ice sheet (Garvin & Williams 1993; Thomas *et al.* 2001). However, 'ground-truth' data are still required, and the Swiss Station mentioned above is now part of the network of Automatic Weather Stations that cover the Inland Ice (Steffen & Box 2001).

With respect to the dynamics of Jakobshavn Isbræ and the adjacent margin of the ice sheet, a number of American and Swiss projects have been carried out along the margin (see also the glaciology section).

Geology

Bedrock geology

Precambrian crystalline rocks such as orthogneisses, granites and metavolcanic and metasedimentary rocks dominate West Greenland and represent the roots of ancient fold belts. Cretaceous sedimentary rocks are found on Disko and western Nuussuaq, where they are overlain by Palaeogene basalts (Fig. 16).

The crystalline bedrock east of Disko Bugt is dominated by Archaean (*c.* 2800 million years old (2.8 Ga)) grey orthogneisses with intercalations of mica schist, amphibolite and minor ultrabasic rocks (Garde & Steenfelt 1999; Kalsbeek & Taylor 1999). In Palaeoproterozoic time, a sedimentary cover (the Anap nunâ Group) was deposited unconformably on this Archaean basement. Both basement and cover were subsequently deformed and metamorphosed

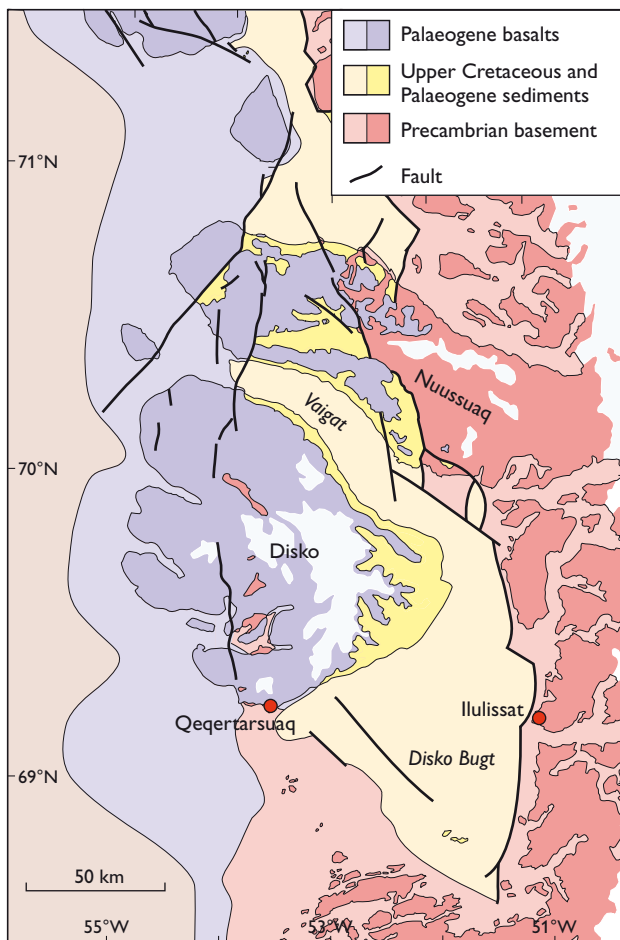


Fig. 16. Onshore (full colours) and offshore (pastel shades) bedrock geology of the Disko–Nuussuaq area. Modified from Chalmers *et al.* (1999) and Larsen & Pulvertaft (2000).

during the *c.* 1.85 Ga old Nagssugtoqidian/Rinkian orogenesis. The Anap nunâ Group is now exposed south of the Torsukattak fjord, forming a curved belt of sandstone, siltstone and minor calcareous rocks (now in greenschist facies) within the reworked Archaean rocks. The Palaeo-

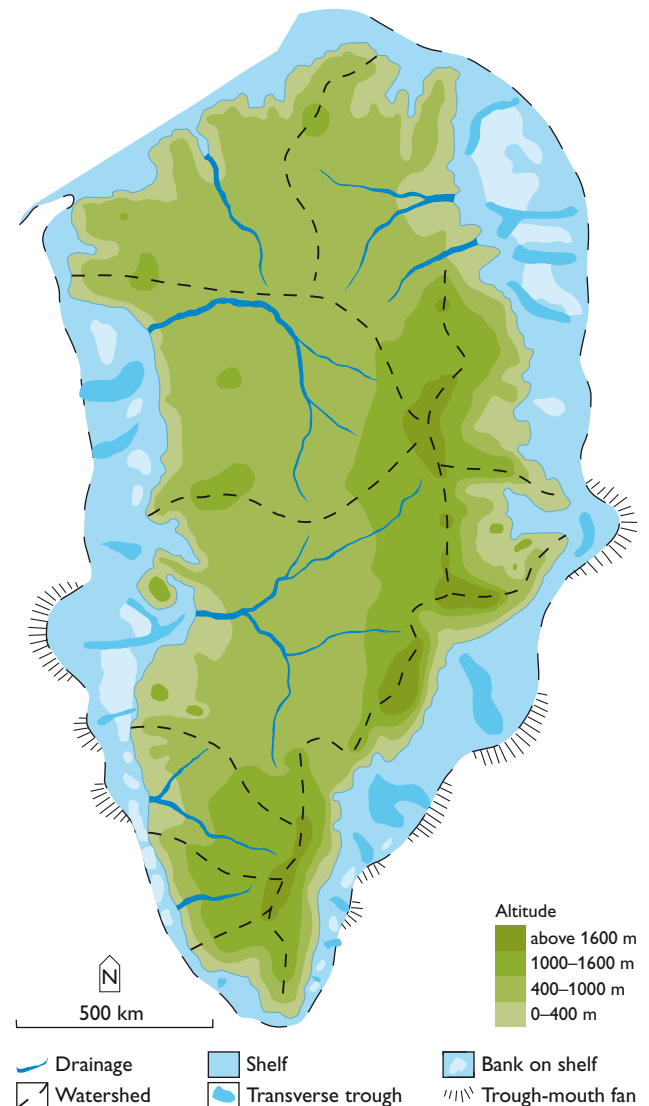


Fig. 17. Map of Greenland without the present ice cover, with altitudes corrected for the present load of the Inland Ice (cf. Fig. 4). In this reconstruction, large parts of central Greenland are shown to be drained by river systems, of which one major system flows into Disko Bugt. The trends of the pre-glacial river systems have been reconstructed from the topography. The offshore continuation of the drainage of the interior of Greenland is reflected by prominent troughs on the shelves. Elevations on land according to Letréguilly *et al.* (1991b), with offshore morphology from Funder (1989) and Escher & Pulvertaft (1995).

proterozoic orogenesis in central West Greenland can be correlated with a similar event in eastern Canada, and was caused by N–S collision of two Archaean continents within a large Canadian–Greenlandic plate-tectonic system (Connelly *et al.* 2005). Several kilometres of Archaean and Palaeoproterozoic overburden, which formed the upper levels of the Palaeoproterozoic orogen, have since been eroded away.

During the Mesozoic, the Precambrian shield began to fragment as a consequence of plate tectonic movements. In West Greenland, down-faulting and rifting parallel to the present coast took place. This led to the formation of a sedimentary basin – the predecessor of the offshore shelf area – and later to the formation of the present Davis Strait.

Sedimentation began in the middle Cretaceous (Pedersen & Pulvertaft 1992; Henriksen *et al.* 2000), and Cretaceous and Paleocene sediments are preserved between Disko (Qeqertarsuaq) and Svartenhuk Halvø (Sigguup Nunaa) (69–72°N). The sediments may originally have extended both east and south of their present area of outcrop (Chalmers *et al.* 1999; Henriksen *et al.* 2000). They are overlain by a thick cover of basalts of Paleocene and Eocene age (see Fig. 16) related to sea-floor spreading and the formation of the Davis Strait (Clarke & Pedersen 1976). Another consequence of sea-floor spreading was the movement of Greenland towards higher latitudes. Thus the Disko Bugt

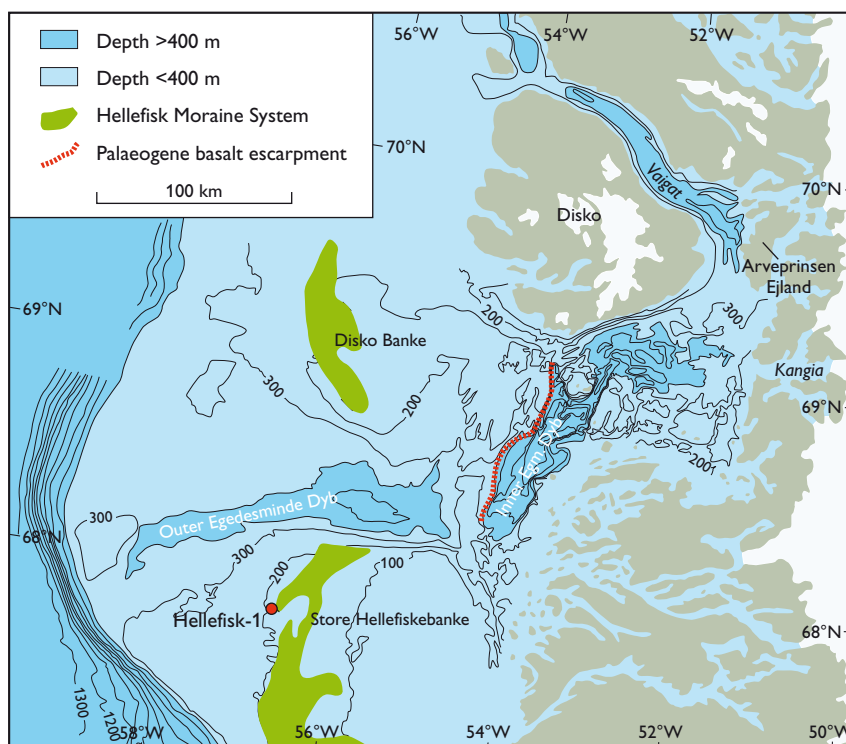
region drifted from *c.* 59°N to its present-day position (*c.* 69°N) during the Cenozoic (Hurley *et al.* 1981).

The Cretaceous and Palaeogene sediments found offshore and onshore show that the Disko Bugt region was a depositional centre for rivers draining large parts of the interior of Greenland. The subsequent eruption of basalts during Late Paleocene and Eocene times blocked and diverted the original drainage southwards, south of Disko.

In general, West Greenland experienced considerable epeirogenic uplift during the Cenozoic (Henderson *et al.* 1981; Japsen & Chalmers 2000; Bonow 2004; Japsen *et al.* 2005). This uplift shaped the present-day morphology of Greenland through several cycles of uplift and erosion, which in West Greenland left a coastland of peneplaned Precambrian bedrock bordered by marine-shelf areas with thick sequences of Cenozoic sediments.

Characteristic features of the shelf are the large offshore, trough-mouth fans at the shelf break that developed off major fjord systems (Figs 17, 18). They seem to have developed during the Neogene as an extension of the drainage of the interior of Greenland. The largest submarine fan of this kind is found off Disko Bugt, and extends so far westwards as to reach Canadian offshore territory. This sediment fan appears to have a long history of formation, with the source area comprising large parts of the interior of Greenland, draining westwards into the Disko Bugt region. The fan

Fig. 18. The continental shelf offshore Disko and Disko Bugt. Bathymetry of the southern area (including Egedesminde Dyb) from Brett & Zarudzki (1979), and of the northern area (including Vaigat) from Chalmers *et al.* (1999). The Hellefisk Moraine System, comprising Disko Banke and Store Hellefiskebanke, is bisected by the outer Egedesminde Dyb. The location of the well ‘Hellefisk-1’ is plotted from Risum *et al.* (1980). Inner Egm Dyb, Inner Egedesminde Dyb.



developed in the Miocene and continued to form during the Pliocene, when a base level 300 m lower than at present has been recorded (Sommerhoff 1975).

As noted above, Rink (1862) had initiated ideas of drainage of the interior of Greenland. These were subsequently elaborated by Cailleux (1952), on the basis of the first systematic mapping of the subsurface of the ice sheet around 1950 (Holtzschcher & Bauer 1954). The subsequent more detailed mapping of the subsurface confirms the importance of the major drainage of central parts of Greenland towards the Disko Bugt area. The drainage patterns can be traced in the transverse troughs cutting the offshore banks, the deep fjords that cut through the coastland, and the present-day ice streams that drain the Inland Ice. Geophysical techniques are unable to localise the actual drainage channels beneath the Inland Ice, and the depiction of the former drainage patterns of the interior is thus provisionally constructed as a best fit of the contours of the subsurface (Fig. 17).

History of glaciations and interglacials

Cooling during the Cenozoic led to increasing intensity of glaciations. The primary cause of the Pliocene–Pleistocene glaciations is ascribed to the periodic and quasi-periodic changes in the Earth's orbital parameters of eccentricity, obliquity and precession, leading to changes in the distribution and amount of solar energy (Fig. 19; Zachos *et al.* 2001). Local climatic developments were, however, modified by variations in the topography and the trends of oceanic currents.

Evidence from deep-sea sediment cores indicates a marked cooling in the Oligocene, with the first formation of the Antarctic ice sheet more than 30 million years (30 Ma) ago (Fig. 19). After a relatively warm period during the Middle Miocene, with a temperature maximum about 15 Ma ago, global cooling continued, leading to a gradual, step-wise glaciation of both the southern and later also the northern hemisphere. This is demonstrated by the occurrence of ice-rafted debris (IRD) in deep-sea sediments in the northern North Atlantic (Thiede *et al.* 1998). The gradual expansion of continental ice can be traced by the oldest IRD occurrence from the Fram Strait between North-East Greenland and Svalbard at 14 Ma, and around Iceland and South Greenland about 10 Ma ago (Thiede *et al.* 1998). An IRD occurrence off South Greenland at 7 Ma was taken as the first indication of full-scale glaciation of southern Greenland (Larsen *et al.* 1994), although this did not involve a permanent glaciation of Greenland. A Pliocene/Pleistocene strengthening of IRD pulses since *c.* 4 Ma has been observed

in the Labrador Sea, just south of Baffin Bay (Thiede *et al.* 1998). Deposition of IRD in deep-sea sediments must be taken as important evidence of contemporaneous glaciations of adjacent coastal areas. However, there is still debate concerning the timing of the IRD pulses and their relationship to the extent of ice cover. It has been speculated as to whether IRD maxima relate to the onset/advance, or to recession/thinning/disintegration, of ice cover. Discussion has also concerned the relationships between IRD deposition and the type or mode of iceberg calving (Reeh *et al.* 1999; Reeh 2004).

Recent data from the Arctic Ocean (Moran *et al.* 2006; Sluis *et al.* 2006), however, may have implications for the glacial record summarised above. Firstly, these data indicate that the Eocene thermal maximum was warmer than hitherto believed. Secondly, a small gneiss pebble, within sediments *c.* 45 Ma old, is interpreted as having been ice-rafted, thus suggesting that the onset of glaciations was synchronous in the Arctic and Antarctica.

Onshore evidence of the first major glaciation of Greenland is provided by the Kap København Formation in northern Greenland. This formation includes deposits from a pre-Tiglian ice age (*c.* 2.5 Ma) that contain shell fragments from the previous warmer period (Reuverian); this is succeeded by sediments referred to the subsequent warm period, the Tiglian, *c.* 2.2–2.4 Ma. The occurrence of forest tundra at this locality at this time indicates a Greenland without an extensive central ice sheet (Bennike 1990; Funder *et al.* 2001).

The subsequent glaciations (ice ages) grew in intensity and it is possible that the present Inland Ice first formed during the Middle and Late Pleistocene (as late as *c.* 0.8 Ma). This required sufficient cooling during the ice ages to form an ice cover that was large enough to survive melting during the intervening interglacials.

In the vicinity of Disko Bugt, the oldest known Quaternary sediments are the deposits at Pattorfik (Fig. 20). They were observed by K.L. Giesecke and H.J. Rink in the first half of the 1800s, and described and investigated in detail by Símonarson (1981) and Funder & Símonarson (1984). Amino acid analyses of shells from these marine deposits indicate that they belong to Early Pleistocene (1 Ma or more). Although situated on the southern shore of Uummannaq Fjord, the deposits have been protected from subsequent glacial erosion during the ice ages by a layer of lithified talus breccia.

Other interglacial and interstadial deposits have been discovered subsequently. These pre-Holocene occurrences are nearly all located at the extreme western coastal parts of central West Greenland (Fig. 20). They occur from western Disko, over the tip of the Nuussuaq peninsula to Svartenhuk

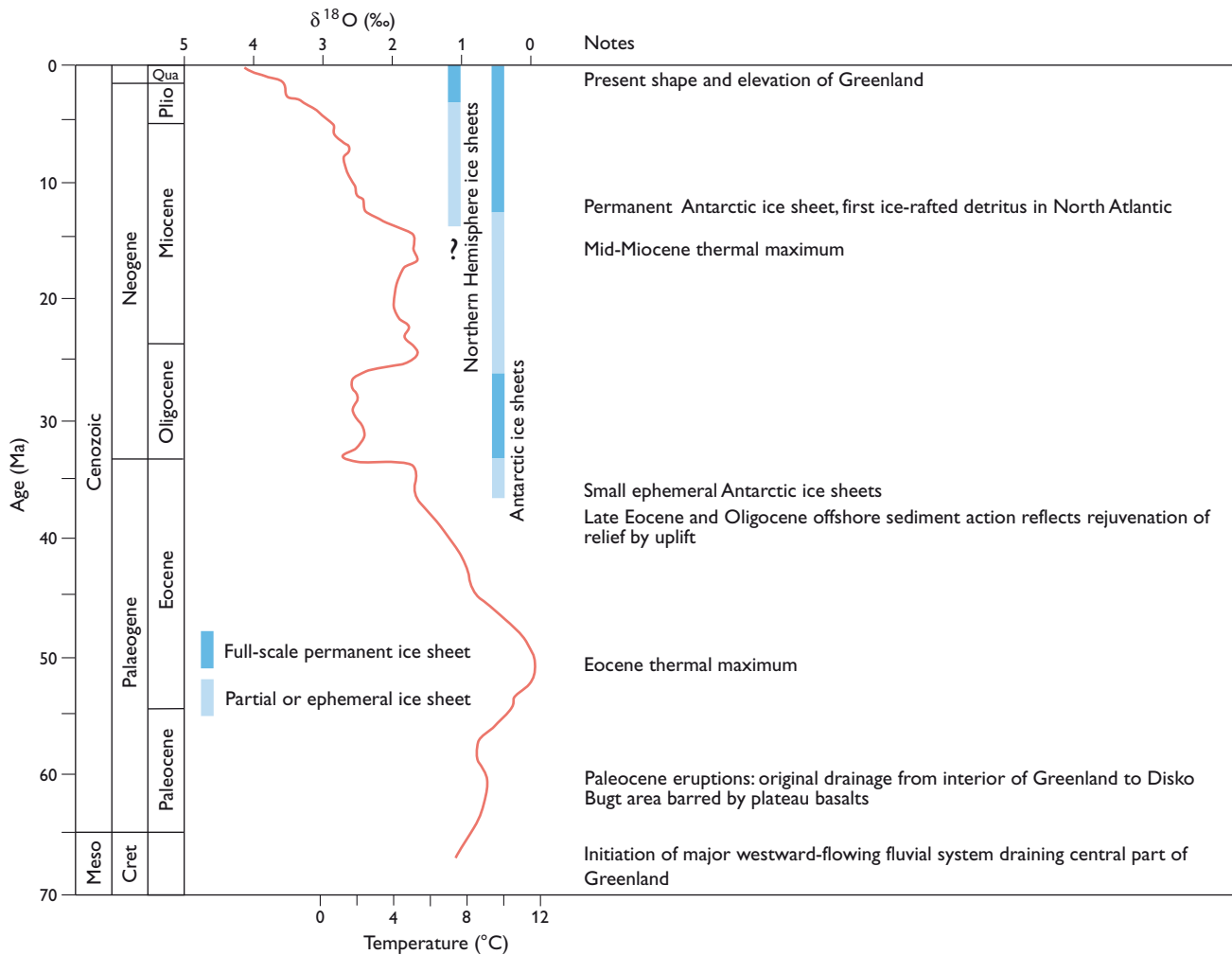


Fig. 19. Global cooling during the Cenozoic is illustrated by a temperature curve based on deep-sea oxygen isotope records (from Zachos *et al.* 2001); the temperature evolution is compared with global ice-sheet development and selected local events in Greenland. Cret, Cretaceous; Meso, Mesozoic; Plio, Pliocene; Qua, Quaternary.

Halvø, and are referred to a border zone of the ice age glaciations. At present, about 25 occurrences are known between Disko Bugt and southern Svartehuk Halvø. On the basis of faunal compositions and amino acid analyses, the deposits have been referred to four pre-Holocene marine events: the interglacial Ivnaarssuit marine event (Early–Middle Pleistocene), the interglacial Nordre Laksebugt marine event (mid-Middle Pleistocene), the interstadial Laksebugt marine event (Middle–Late Pleistocene), and the last interglacial Svartehuk marine event (Eemian/Sangamonian; 130 000 – 115 000 years before present (130–115 ka B.P.); Bennike *et al.* 1994). One of the interglacial sites is located on eastern Disko island (Fig. 20), and consists of reworked shells in moraine (Funder & Símonarson 1984).

During the intervening ice ages, the ice sheet expanded out to the offshore banks, implying that there was little coastal lowland on which deposits older than the Holocene could be preserved. Based on weathering differences, it has been established that an extensive glaciation (Hellefisk glacial event) resulted in the expansion of the ice out to the shelf break south of *c.* 68°N (Kelly 1985); this event has been tentatively dated to the Illinoian/Saalian (*c.* 380 to 130 ka B.P., Gibbard *et al.* 2005, or *c.* 300 to 130 ka B.P., Geyh & Müller 2005). The moraine system referred to the Hellefisk glacial event is depicted in Fig. 18 after Brett & Zarudzki (1979) and Kelly (1985). The moraines are situated near the shelf break on the southern Store Hellefiskebanke, and on the central parts of Disko Banke, presumably because of the greater depth of this bank (Fig. 18).

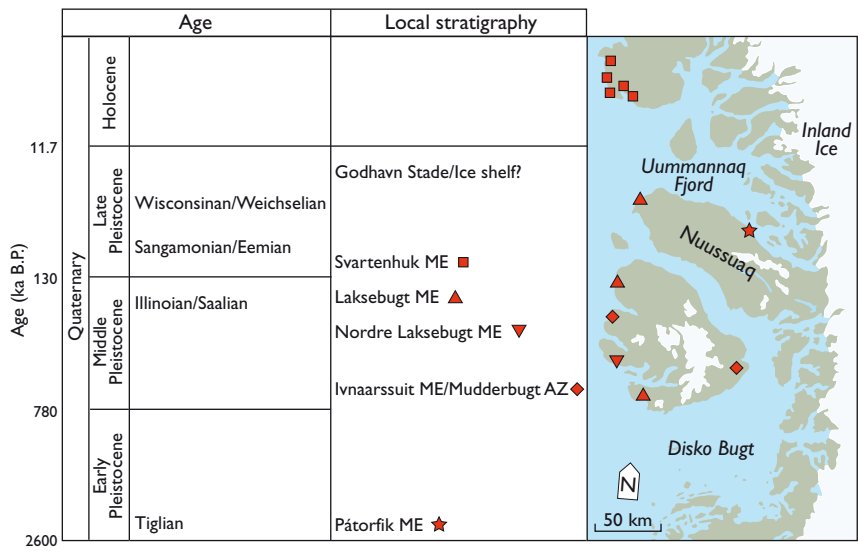


Fig. 20. Stratigraphic summary of interglacial and interstadial deposits between Disko Bugt and southern Svartenhuk Halvø, based on Kelly (1986) and Bennike *et al.* (1994). AZ, aminozone; ME, marine event. The accompanying map shows the location of these deposits; the symbols represent one or more localities.

There is little information to constrain the extent of the ice-age ice sheets of West Greenland. As mentioned above, it is possible that the Illinoian/Saalian ice cover extended over Store Hellefiskebanke out to the shelf break in the shelf areas south of Disko island. The till on Store Hellefiskebanke covers a thick suite of Eocene and younger deposits (Whittaker 1996; Chalmers *et al.* 1999). In contrast, on the eastern parts of Disko Banke, a ‘rough sea floor type’ seems to indicate that volcanic rocks underlie the Quaternary sediments (Brett & Zarudzki 1979).

Early investigations reported erratic gneiss boulders on the basalt terrain of Disko, and on the outer parts of Nuussuaq and Svartenhuk Halvø farther to the north (Steenstrup 1883b). On Qilertinnguit on the north coast of Nuussuaq (Fig. 1), Steenstrup reported erratic occurrences up to 1200 m a.s.l., and in Nordfjord (Kangersooq) and around Disko Fjord (Kangerluk) on Disko he noted occurrences up to 920 m and 628 m a.s.l. respectively (for location see Fig. 2). All three localities are situated close to the outer coast and could be related to the Illinoian glaciation, or even older ice ages. If so, the outer Egedesminde Dyb (Fig. 18) must have repeatedly served as a southern drainage outlet from Disko Bugt for a high-arctic ice stream, probably moderate in size and with a depth of only 600 m. The calf-ice production was presumably also moderate, as can be seen from the relationship between the mode of calving of ice sheet outlets and temperature conditions (Reeh 1994; Reeh *et al.* 1999).

The last interglacial (Sangamonian/Eemian) and the last ice age (Wisconsinan/Weichselian)

It was noted above that the Hellefisk glacial event may have been followed by the Svartenhuk marine event, which is referred to the last interglacial stage. The Sangamonian/Eemian interglacial has been described as more humid and warmer than the present interglacial, with a mean summer temperature up to 5°C higher than in the Holocene (Bennike & Böcher 1994). Based on the climate record from ice cores and the altitude of the surface and subsurface of the present ice sheet, a detailed picture of the changes in the extent of the ice sheet throughout the last 130 000 years has been modelled (Letréguy *et al.* 1991a, b; Weis *et al.* 1996). These models indicate that ice-sheet recession during the Eemian interglacial was so extensive that the Inland Ice almost split into a large northern ice sheet and a minor southern one, and left large parts of south-western Greenland ice-free (Fig. 14). The stronger recession in the south-west is ascribed to the generally low elevation of this region, which together with strong ablation resulted in a rapid recession. Loss by calving through ice streams may at this time have been confined to outlets north of Disko Bugt. An extensive reduction of the Inland Ice is also indicated by the character of the basal ice in the Camp Century and Dye 3 ice cores (Koerner 1989). The basal silty ice from the Dye 3 core has tentatively been dated at 450–800 ka B.P., which suggests that the Dye 3 region was not deglaciated during the last interglacial (Willerslev *et al.* 2007).

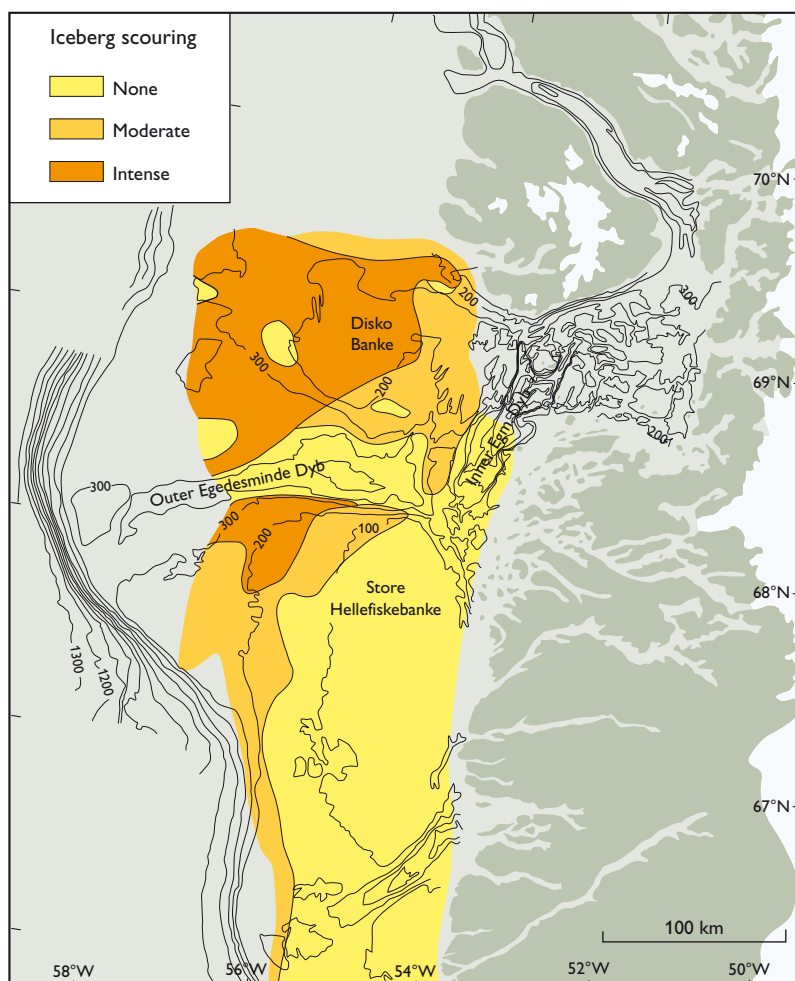
The subsequent build-up of the Inland Ice has been modelled by Letréguilly *et al.* (1991b). The general trend shows a slow temperature decrease from the end of the Eemian through the early and middle parts of the Wisconsinan, followed by large oscillations as shown by the ice-core records (Dansgaard *et al.* 1993). During short-lived cold periods at about 70 ka B.P. and 25–22 ka B.P., the temperature fell to around 23–25°C below the present (Johnsen *et al.* 1995; Dahl-Jensen *et al.* 1998; Dansgaard 2004). Throughout the Wisconsinan, the ice-core records show abrupt temperature changes of 10–12°C that seem to be related to abrupt changes in the course and intensity of ocean currents (Broecker *et al.* 1985). The last glacial temperature minimum has been dated to around 21.5 ka B.P. (Johnsen *et al.* 1995).

Around Disko Bugt, as elsewhere in Greenland, the last glacial maximum (LGM) can be related to offshore features and deposits, although the exact position of the ice margin during the LGM is still a matter of debate. The ice margin at the LGM is usually related to offshore morpho-

logical features of the banks and the intervening troughs ('dyb', the transverse channels of Holtedahl 1970). In central West Greenland, the most prominent of these troughs is the c. 350 km long Egedesminde Dyb – where an ice-age predecessor of Jakobshavn Isbræ may have been located (Figs 18, 21).

The south side of Egedesminde Dyb is limited by Store Hellefiskebanke, one of the largest of the offshore banks in West Greenland. At the most shallow locality, the water depth is only 8 m. An exploration well (Hellefisk-1; Fig. 18) drilled on the western slope of the bank revealed a 200 m thick cover of probable glacial till, overlying c. 3 km of Cenozoic sediments resting on basaltic lava flows (Risum *et al.* 1980). Geophysical profiling has also revealed thick deposits of Cenozoic sediments (Henriksen *et al.* 2000). The surface morphology indicates significant marginal moraines, of which a western system near the shelf break (as noted above) could be referred to the Illinoian. Another system of moraines is found around the eastern slopes of Store Hellefiskebanke, and on the banks farther south. These

Fig. 21. Distribution and intensity of iceberg scouring on the shelf offshore West Greenland, from Brett & Zarudzki (1979). Inner Egm Dyb, inner Egedesminde Dyb. Bathymetric contour intervals are 100 m.



moraines continue as marginal and terminal moraine systems surrounding the troughs between the banks, and their lobate nature could indicate that the moraines were originally deposited on land (Funder 1989). During the Wisconsinan ice age, the global sea level was about 130 m below the present (Lambeck & Chappell 2001). However, the increased glacier load in Greenland during this period would have reduced this figure, and abrasion terraces and beach ridges down to 70 m below present sea level in West Greenland imply that the Wisconsinan ice margin rested on dry land to a large degree (Sommerhoff 1975). The 'inner moraines' of the offshore areas have been correlated with the oldest and highest situated moraines of the coastal high mountains (distinguished by their degree of weathering) and with the period described as the 'Sisimiut glaciation' (Kelly 1985).

Disko Banke, located south-west of Disko, is bounded to the south-east and south by the troughs referred to as the inner and outer Egedesminde Dyb (Fig. 18). The shallowest part of Disko Banke has a water depth of *c.* 150 m, with the surface only partly till-covered. Eocene and younger sediments dominate the western part of Disko Banke (Chalmers *et al.* 1999), whereas the surface of the eastern part comprises volcanic rock ridges with a thin, discontinuous layer of till (Zarudzki 1980). The basalts dip gently westward, with steep scarps to the east (Brett & Zarudzki

1979). A pronounced east-facing escarpment connects Disko Banke in the north with the easternmost part of Store Hellefiskebanke in the south. This escarpment forms a threshold with depths of 200–300 m below present sea level. It divides Egedesminde Dyb into an eastern, deep and narrow channel with maximum depths over 1000 m (inner Egedesminde Dyb) and a shallower western part with depths up to around 600 m (outer Egedesminde Dyb; Fig. 18).

If the moraine system on Store Hellefiskebanke does mark the outer limit of an extensive Illinoian glaciation, then the 'reduced' extent of a Wisconsinan glaciation implies that the western shores of Disko were only characterised by a shelf glaciation (Bennike *et al.* 1994). Along the coast of southern Disko, near the town of Godhavn, are major moraines that were formed during the Godhavn stade (Fig. 22; Ingólfsson *et al.* 1990). These moraines are taken as evidence for the maximum extent (LGM) of the Greenland ice sheet during the Wisconsinan. The main moraine, 'Pjetursson's moraine', is *c.* 1.5 km long and reaches an elevation up to 220 m a.s.l. east of Godhavn. Smaller moraines are found up to 110 m a.s.l. about 15 km farther to the west. Although closely related to the marine limit of the area, it cannot be excluded that the Godhavn stade marks an early re-advance, or halt, during recession from the LGM.

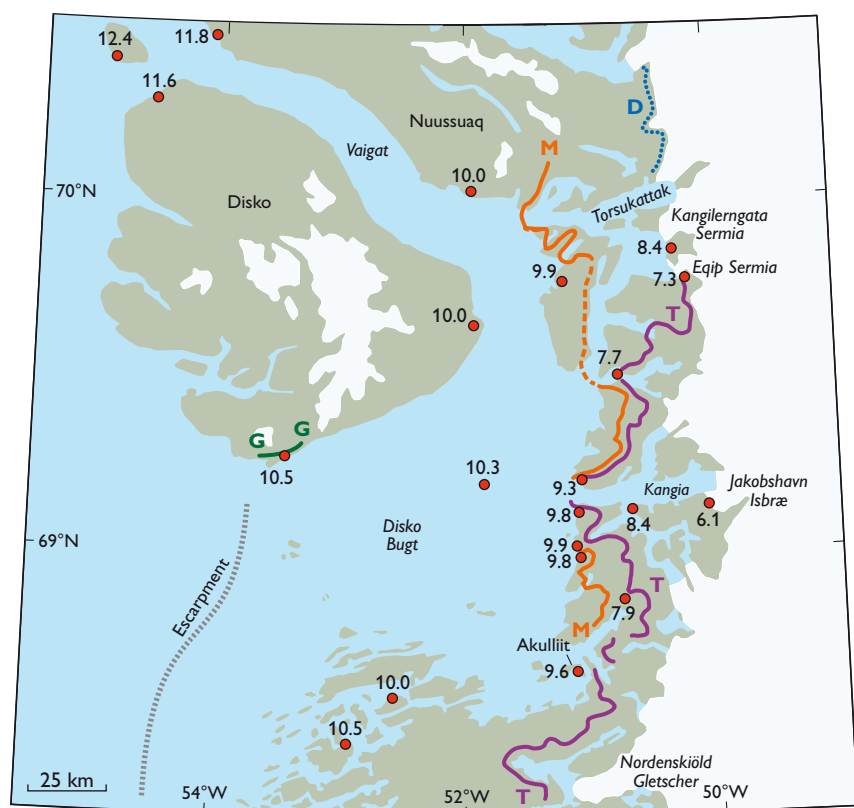


Fig. 22. Radiocarbon dates pertaining to the last deglaciation of the ice-free areas around Disko Bugt. Dates are given in calibrated thousand years before present (cal. ka B.P.), using the INTCAL04 data set for calibration (Reimer *et al.* 2004). Based on dates published by Ingólfsson *et al.* (1990), Rasch (1997), Bennike *et al.* (1994), Bennike & Björck (2002), Long & Roberts (2003), Long *et al.* (2006) and this study (Appendix 2); further details are given in Tables 2 and 3. The map also shows the approximate trend of the basalt escarpment west of the mouth of Disko Bugt, the Mairait moraine system (M), the Tasiussaq moraine system (T), the Drygalski moraines (D) and moraines of the Godhavn stade (G).

Table 2. Radiocarbon age determinations relating to the last deglaciation (see also Table 3)

Locality	Latitude N	Longitude W	Laboratory number*	Material	¹⁴ C age† (± 2σ) years B.P.	Rcorr ¹⁴ C age‡ years B.P.	Calib. age‡ years B.P.	Mean age ka B.P.	Reference
Hareøen	70°23′	54°57′	Ua-1789	Shells	10870 ± 130	10470	12800–11950	12.4	Bennike <i>et al.</i> (1994)
W Nuussuaq	70°28′	54°02′	AAR-3496	Bryophytes	10160 ± 75		12100–11400	11.8	Bennike (2000)
NW Disko	70°16′	54°37′	I-16393	Shells	9920 ± 150	9920	12050–11050	11.6	Bennike <i>et al.</i> (1994)
E Nuussuaq	70°04′	52°06′	K-994	Shells	8940 ± 170	8940	10500–9550	10.0	Weidick (1968)
E Disko	69°40′	52°01′	K-3667	Gyttja	8950 ± 125		10400–9600	10.0	Ingólfsson <i>et al.</i> (1990)
E Disko	69°40′	52°00′	K-3660	Shells	8700 ± 120	8700	10200–9500	9.9	Ingólfsson <i>et al.</i> (1990)
Arveprinsen Eiland	69°46′	51°15′	Beta-107879	Gyttja	8820 ± 100		10200–9550	9.9	Long <i>et al.</i> (1999)
Godhavn	69°17′	53°28′	AAR-5	Shells	9650 ± 250	9250	11250–9750	10.5	Ingólfsson <i>et al.</i> (1990)
Central Disko Bugt	69°11′	51°49′	AA-37711	Foraminifers	9483 ± 65	9083	10500–10150	10.3	Lloyd <i>et al.</i> (2005)
Egedesminde	68°36′	52°34′	Hel-362	Shells	8970 ± 170	8970	10500–9550	10.0	Donner & Jungner (1975)
S Egedesminde	68°26′	52°57′	AA-38842	Gyttja	9330 ± 99		10800–10200	10.5	Long & Roberts (2003)
SE Disko Bugt	68°40′	51°07′	AA-39659	Gyttja	8585 ± 86		9790–9430	9.6	Long & Roberts (2002)

* Sources of age data: Ua: Ångström Laboratory, Uppsala. AAR: Aarhus AMS C14 Dating Centre. I: Teledyne Isotopes. K: The former radiocarbon laboratory in Copenhagen. Beta: Beta Analytic. AA: The NSF-Arizona Facility. Hel: The Dating Laboratory at University of Helsinki.

† The radiocarbon age determinations from Ua and AAR have been corrected for isotopic composition by normalising to –25‰ on the PDB scale, and those from K by normalising to 0‰. The date from I has not been normalised.

‡ Rcorr: Reservoir-corrected. The age determinations on marine material from Ua and AAR have been seawater reservoir corrected by subtracting 400 years. The dates from K and I have not been corrected (Bennike 1997).

§ Calibrated using the INTCAL04 data set (Reimer *et al.* 2004) and the OxCal. v.3.10 software program (Bronk Ramsey 2001).

The outer Egedesminde Dyb was described as a valley by Zarudzki (1980), who also noted that “well-preserved flank moraines, found at gradually lower elevations in the valley, testify to a recent withdrawal of the ice” (Zarudzki 1980, p. 60). It may be suggested that the outer Egedesminde Dyb drained a high-arctic type ice stream during the LGM, as well as during earlier more extensive glaciations, although it was probably restricted in size and also in calf-ice production.

The northern drainage route for Wisconsinan ice, the Vaigat (Sullorsuaq) strait, is a typical glaciated fjord with depths over 600 m in its south-eastern part and around 200–300 m near Hareøen (Qeqertarsuaq) at its western mouth. Quaternary deposits with a thickness of several hundred metres are found in the strait (Denham 1974).

There is general agreement on a ‘reduced’ extent of the Wisconsinan ice sheet during the LGM. The margin of the ice sheet may have been situated on the proximal parts of the offshore banks south of Disko Bugt, with outlets in the intervening transverse troughs reaching terminal moraines (or thresholds) near the shelf break. The northern conduit of ice from Disko Bugt through Vaigat only filled the strait as far as a position near Hareøen at its mouth. This is in accordance with the concept of a LGM limit at the mouth of the fjords farther north in West Greenland (Funder 1989). The acceptance of the Godhavn stade at or near the LGM on southern Disko near Egedesminde Dyb, would leave the whole of western Disko unglaciated by the ice sheet, but presumably with local glaciers of high-arctic type contributing to shelf ice in Baffin Bay. The nature of this shelf may have been comparable to

that of the present Ellesmere Island ice shelf in Canada, which has a thickness of up to 100 m, and shows a gradual transition between an ice shelf evolved from sea ice and a shelf of glacier ice (Jackson 1997; Jeffries 2002), or to the ice shelves in northern Greenland (Koch 1928; Higgins 1989). Ice shelves are known to be particularly sensitive to climatic changes. This is well documented for ice shelves in Antarctica, such as the Ross Shelf (Bindschadler & Bentley 2002), and also for the ice shelves of northern Ellesmere Island adjacent to the Arctic Ocean (Jeffries 2002). A comparable great variability in ice-shelf extent, related to former climatic oscillations, would be expected during ice ages.

A record of extensive iceberg scouring of the seabed in the eastern part of Disko Banke and the northern slope of Store Hellefiskebanke (Fig. 21; Brett & Zarudzki 1979), points to the close proximity of a calving glacier margin at Disko Banke; the higher, northern part of Store Hellefiskebanke may have been dry land. Iceberg scouring is recorded to a depth of 340 m, and it is assumed that the 300 m deep bedrock threshold across the outer and inner part of Egedesminde Dyb did not allow icebergs with a greater draft to pass. It is presumed that the deeper scours were created by icebergs calved in outer Egedesminde Dyb when the ice extended beyond the threshold. The relative sea level at the end of the LGM may also have been higher than at present.

The escarpment at Egedesminde Dyb south of Disko may have had a braking effect on the glacier ice, but the width of the glacier (*c.* 130 km) may have compensated for the reduced ice thickness. The subsequent concentration of ice

Table 3. Selected early Holocene radiocarbon age determinations (see also Table 2, Appendix 2)

Locality	Latitude N	Longitude W	Altitude m a.s.l.	Laboratory number*	Material [‡]	¹⁴ C age [†] (± 2σ) years B.P.	Rcorr ¹⁴ C age [§] years B.P.	Calib. age [‡] ka B.P.	δ ¹³ C ‰	Reference
Ilulissat and Qasigiannuit areas										
Sermermiut	69°12′	51°04′	1–2	Ua-1086	Mc shell	8795 ± 130	8395	9.3	0 [¶]	This study
'Qassortoq'	69°06′	51°04′	5–10	K-1818	Shells	8630 ± 130	8630	9.8		Weidick (1972b)
Pinguarssuit	69°05′	51°08′	21	I-6243	Shells	6835 ± 125	6835	7.7		Weidick (1973)
'Sandbugten'	69°03′	51°08′	?	K-2022	Shells	7690 ± 120	7690	8.6		Weidick (1974a)
Lersletten	69°02′	51°01′	36	K-992	Shells	7110 ± 140	7110	7.9		Weidick (1968)
Narsarsuaq	69°02′	51°01′	33	K-987	Gyttja	7850 ± 190		8.8		Tauber (1968)
Tasiusaq	69°02′	50°56′	10–15	Ua-4575	Pa shell	8140 ± 95	7740	8.7	–14.49	This study
'Lerbugten'	69°01′	51°08′	?	K-2023	Shells	8680 ± 135	8680	9.8		Weidick (1974a)
Marraq, w. part	69°00′	51°07′	25	Ua-4574	Shell	9180 ± 75	8780	9.9	–0.32	This study
Eqaluit	68°56′	50°58′	25	K-993	Shells	7650 ± 140	7650	8.5		Weidick (1968)
Eqaluit	68°56′	50°58′	20?	Ua-4573	Mt shell	8215 ± 80	7815	8.7	1.21	This study
Serfarsuit	68°50.5′	50°47′	15	Ua-4572	Mt shell	7500 ± 75	7100	7.9	1.33	This study
Other areas										
Ussuit	67°51′	50°16′	42	K-1556	Shells	6760 ± 130	6760	7.6		Kelly (1973)
Eqip Sermia	69°46′	50°13′	0.8–2.3	K-6373	Shells	6420 ± 110	6420	7.3	1.4	Rasch (1997)
Qapiarfuit	69°52′	50°19′	2–3	K-3663	Shells	7600 ± 110	7600	8.4	–0.1	Ingólfsson <i>et al.</i> (1990)

* Sources of age data: Ua: Ångström Laboratory, Uppsala. I: Teledyne Isotopes. K: The former radiocarbon laboratory in Copenhagen.

[‡] Mc: *Macoma calcarea*, Pa: *Portlandia arctica*, Mt: *Mya truncata*.

[†] The radiocarbon age determinations from Ua have been corrected for isotopic composition by normalising to –25‰ on the PDB scale, and those from K by normalising to 0‰. The date from I has not been normalised.

[§] Rcorr: Reservoir-corrected. The age determinations on marine material from Ua have been seawater reservoir corrected by subtracting 400 years. The age data from K and I have not been corrected (Bennike 1997).

[‡] Calibrated using the INTCAL04 dataset (Reimer *et al.* 2004) and the OxCal. v.3.10 software program (Bronk Ramsey 2001).

[¶] Assumed value.

masses through the funnel-like shallow conduits in the bedrock of the eastern part of Disko Banke might then have led to a concentration of ice in the outer Egedesminde Dyb. The high-arctic conditions of the environment and the ice cover at that time (low precipitation, low temperature) suggest that this outer Egedesminde Dyb ice stream was less important than the present Jakobshavn Isbræ; it may have been similar to present glaciers in northern Greenland (Reeh 2004) or Antarctica (Thomas 1979; Swithinbank 1988; Jacobs *et al.* 1992). These glaciers, or ice streams, are characterised by greater width and lower velocities than present ice streams in the southern part of Greenland (Rignot & Kanagaratnam 2006). In addition, they have long, floating tongues with a high rate of bottom melting. Calf-ice production, precipitation and ablation are more limited than in present-day ice streams in the southern parts of Greenland. The dynamic differences between Arctic and Antarctic ice streams are considered by Truffer & Echelmeyer (2003).

The oldest radiocarbon age determination from the mouth of Disko Bugt is 10.5 ka B.P. (Table 2; Ingólfsson *et al.* 1990), which post-dates the Godhavn stade mentioned above (Fig. 22). For the north-western entrance to the Vaigat strait, the information is sparse, but a shell has been dated to 12.4 ka B.P. (Bennike *et al.* 1994; Bennike & Björck 2002); an estimate of the extent of the ice cover before that time can only be speculative.

To summarise the events from the LGM (22 ka B.P.) to the glacial situation at 13–10 ka B.P., it is suggested that onset of a climatic initial warmth around 20 ka B.P. was followed at *c.* 19 ka B.P. by a rise in sea level, with a break-up of the outer parts of marine ice shelves and margins at around 18 ka B.P. At 16–14.5 ka B.P., a further rise in sea level coupled with the Allerød/Bølling warm period (14.7–12.6 ka B.P.; Lambeck & Chappell 2001) accelerated the process of thinning of the ice margin and break-up of the ice shelf. During the Younger Dryas (12.6–11.7 ka B.P.) the ice margin may have receded to a position near the basalt escarpment between Disko Banke in the north and Store Hellefiskebanke in the south. This escarpment must have formed a barrier that retained and hampered the glacier ice flow to the sea. Thus little recession of the ice margin took place. The Vaigat lobe of the ice sheet north of Disko probably reached the outer part of the Vaigat strait.

From the extent of the ice margin during the Godhavn stade, and from comparisons with the present-day surface profile of the ice sheet, it is estimated that the ice cover over Disko Bugt had a thickness of 1000–1500 m. This estimate is in agreement with the elevation of the nunatak moraines on the outer coastal highland farther south around 67°N, the 'Taserqat stade' of Weidick (1972a).

The collapse of the ice cover in Disko Bugt

The seabed of the southern and central parts of Disko Bugt is mainly 200–400 m below sea level, and characterised by a rugged bedrock terrain (Fig. 18; Brett & Zarudzki 1979). Seismic data reveal Quaternary deposits of 100 m or more in some places (Denham 1974; Chalmers *et al.* 1999). On the bathymetric map (Fig. 18), a submarine divide connects eastern Disko and Arveprinsen Ejland (Alluttoq). From this area, depressions ('drowned glacial valleys') lead either north-westwards to the Vaigat strait or south-westwards towards Egedesminde Dyb. Off Kangia, two E–W-trending channels can be seen as a continuation of the fjord (Long & Roberts 2003), although they are only about 400–500 m deep.

Minimum ages for the chronology of the deglaciation of Disko Bugt are provided by dates of shells from raised marine deposits, from marine sediment cores or from dating of basal gyttja in lakes (Fig. 22).

Sediment cores have been retrieved from the inner Egedesminde Dyb, 20–30 km east of the threshold, but only mid- to late Holocene sediments were penetrated (Kuijpers *et al.* 2001; Jensen 2003). Other cores have been collected in eastern Disko Bugt (Lloyd *et al.* 2005), and a minimum date for the deglaciation here is 10.3 ka B.P. The onset of a branch of the West Greenland Current into Disko Bugt has been dated at *c.* 9.2 ka B.P. – at a time when Jakobshavn Isbræ terminated at Isfjeldsbanke; it receded from this position at 7.9 ka B.P., an event indicated by a reduced sedimentation rate seen in the cores (Lloyd *et al.* 2005).

Recent detailed investigations of isolation basins, relative sea-level changes and deglaciation history have been carried out by Long *et al.* (1999, 2003) and Long & Roberts (2002, 2003). As mentioned above, a minimum date for the deglaciation of the Godhavn area is recorded by a date of *c.* 10.5 ka B.P. A date of *c.* 9.6 ka B.P. for basal gyttja from a lake above the marine limit provides a minimum age for deglaciation of the south-eastern corner of Disko Bugt (Table 2; Long & Roberts 2002). From the area between Qasigianniguit and Ilulissat, several dates have been obtained on shells from basal marine deposits, described by Laursen (1944, 1950) and Laursen (in: Weidick 1974a); selected dates are presented in Table 3. The ages, between 9.9 and 9.3 ka B.P. are presumed to be related to a marine limit of *c.* 70 m a.s.l. From Arveprinsen Ejland, farther north in Disko Bugt, basal gyttja from isolation basins has yielded ages up to *c.* 9.9 ka B.P. (Long *et al.* 1999), also related to a marine limit of about the same altitude. A dating of shells from easternmost Disko also yielded an age of 10.0 ka B.P. (Table 2; Ingólfsson *et al.* 1990).

At about the same time, the Vaigat strait became progressively ice-free. An age determination of 12.4 ka B.P. provides a minimum age for the deglaciation of the mouth of Vaigat (Bennike *et al.* 1994). The outer part of Vaigat was deglaciated before 11.8 ka B.P. (Bennike 2000) and the inner part before 10.0 ka B.P. (Table 2; Weidick 1968).

For both routes from Disko Bugt to Davis Strait, minimum dates for the last recession are thus available, but a few details about the processes during the recession are known. The broad nature of the mouth of the bay south of Disko may have led to fast recession, reinforced by a temporary ice stream in the inner Egedesminde Dyb (Long & Roberts 2003). However, the threshold between Disko and Store Hellefiskebanke may have acted as an iceberg bank, in the same way as the present Isfjeldsbanke at Ilulissat restrains the icebergs coming from the present Jakobshavn Isbræ.

Halts in the recession at 'pinning points' at the mouth as well as in the bay (Long & Roberts 2003) may have influenced the rate of recession, but these halts may have been relatively short stops of decades during a fast recession. The shallow-water belt between Godhavn and Aasiaat, including the islands in the mouth of Disko Bugt, may have caused a short halt in recession.

Calculations of the calf-ice production of the former ice streams in Disko Bugt, based on an empirical correlation between calf-ice production and contemporaneous suggested water depth at calving glacier fronts (Pelto & Warren 1991), and combined with the 'Jakobshavn effect', are provided by Long & Roberts (2003). This is an interesting approach to understanding the life of ice streams. The 'Jakobshavn effect' is caused by rising temperatures, which lead to increased surface melting at the ice margin (Fastook & Hughes 1994). The meltwater drains into crevasses and moulins, warming the ice and lubricating the bed, and leading to higher velocities and increased crevassing of the surface, which again leads to increased heat transport from the surface to the bottom of the glacier. However, this is just one element in the complex interplay of mass-balance changes and glacier response to climatic change through the dynamics of the ice margin. Further elements need to be included to explain the onset and demise of the individual ice streams, which are actors in the break-up of such a large segment of the ice-sheet margin as the former marine outlet covering Disko Bugt.

Another climatic element may explain the 'Disko stade' of Disko, during which local glaciers filled most of the broad valleys on eastern Disko. It has been dated to about 10.7 ka B.P. (Ingólfsson *et al.* 1990). The local readvance at this relatively late and warm time has been explained by changes in the prevailing wind systems causing heavier

snowdrift from the west to the east. Moraines referred to the Disko stade are also found on other parts of Disko, with correlation made by determination of the depression of the glaciation level. However, the violent expansion of the local glaciers during the Godhavn stade in eastern Disko may perhaps be related to surging behaviour. At the present day, surging glaciers are common and widespread on Disko and in central East Greenland, where the bedrock

is dominated by basalt (Weidick 1988). In recent studies, it was found that 75 out of 247 local glaciers on Disko could be classified as surge-type glaciers, and that the quiescent phase could be as long as 100 years or more (Yde & Knudsen 2005, in press). In the broad valleys on eastern Disko, large areas of relict glacier ice are seen, which can be explained by glaciers that have surged in the past. The relict ice on Disko has not been dated, but Neoglacial relict

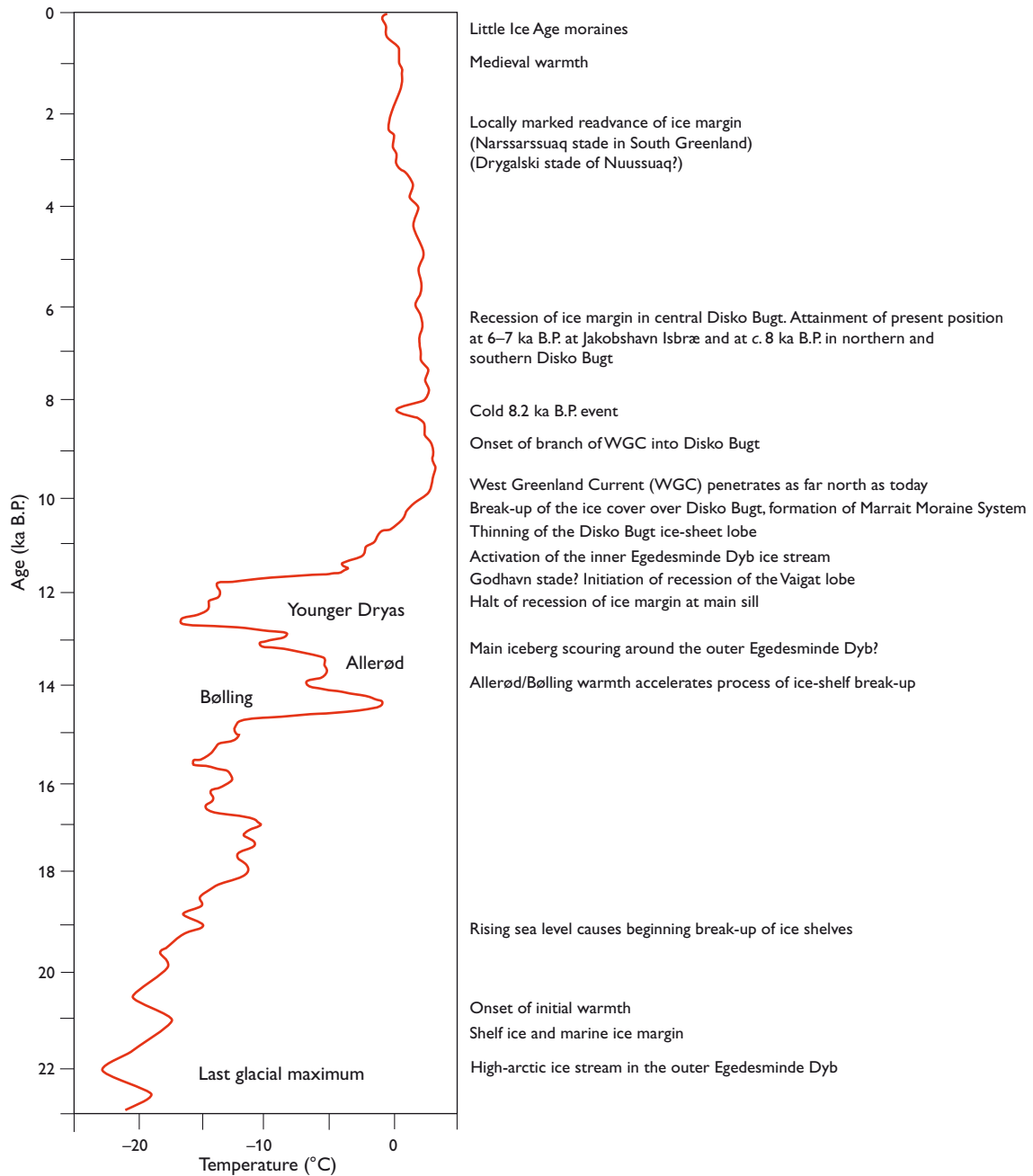


Fig. 23. Trend of the temperature development since the last glacial maximum, compiled from ice-core records (Dansgaard *et al.* 1984; Dahl-Jensen *et al.* 1998; Dansgaard 2004) and events related to the recession of the ice-sheet margin around Disko Bugt.

ice in Sanddalen, Jökelbugten, North-East Greenland, shows similarities to the relict ice on Disko (Bennike & Weidick 2001).

The abrupt temperature rise at the transition from the Younger Dryas to the Holocene at 11.7 ka B.P. (Fig. 23) must have led to intense melting and thinning of the ice margin, especially in a lowland or marine environment such as Disko Bugt. The opening of the ice-filled nearby Davis Strait would have led to increasing humidity and ablation. Heat transfer from the ocean to the over 100 km long glacier front in western Disko Bugt must have been appreciable, as the West Greenland Current at 10.2 ka B.P. seems to have penetrated just as far north then as it does at the present day (Funder 1990, 1994). An 'ice stream' occupying the inner Egedesminde Dyb may have favoured the whole process of fast disintegration of the glacier lobe in Disko Bugt. With its NE–SW orientation and its form, the inner Egedesminde Dyb cannot be classified as a typical U-valley, but its large depth could be due to glacial erosion. Perhaps the inner Egedesminde Dyb is related to the N–S-oriented marginal channels, which separate the gneiss terrain near the coast from the offshore banks (Holtedahll 1970). The marginal channels were formed by faulting followed by fluvial and glacial erosion.

The role of the inner Egedesminde Dyb was gradually reduced as the ice margin receded to the north and east beyond the inner Egedesminde Dyb. The fast recession came to a halt when the front of Jakobshavn Isbræ became anchored at Isfjeldsbanen near Ilulissat at 9.9 ka B.P. The frontal area was, however, strongly reduced, scarcely allowing for more than a fraction of present day calf-ice production (Weidick 1994a, b; Long & Roberts 2003).

The marked increase in temperature at the transition from the Younger Dryas to the early Holocene, at *c.* 11.7–10.0 ka B.P., was followed by a marked rise of global sea level. A relative sea level of *c.* 70 m above the present was reached at *c.* 10 ka B.P. in eastern Disko Bugt. In the Disko Bugt region, the recession of the glacier lobe from the escarpment began in this period, so that the outer parts of Disko Bugt quickly became free of glacier ice.

The main thinning of the Disko Bugt piedmont lobe took place in the millennium after the end of the Younger Dryas, and at around 10.5 ka B.P. the Godhavn area and areas near Aasiaat were ice-free (Dønner & Jungner 1975; Ingólfsson *et al.* 1990; Long *et al.* 2003). Deglaciation dates show that the outer parts of Disko Bugt were already ice-free at around 10.3 ka B.P. (Fig. 22). The final break-up of the ice cover in the bay may have taken place over a few centuries or less.

During the recession, at around 10.0–9.5 ka B.P., a major change in the condition of the ice margin occurred.

Large parts of the ice margin were now resting on land. The recession (or break-up of the marine ice) probably continued through the Torsukattak fjord system in the north-eastern part of Disko Bugt, or was brought to a halt in the eastern parts of Disko Bugt. Palaeoceanographic investigations show that a strengthening of the West Greenland Current took place at 9.2 ka B.P. in Disko Bugt (Lloyd *et al.* 2005). At this time, the front of Jakobshavn Isbræ had receded somewhat into Kangia, as shown by the occurrence of shells at Sermermiut that yield an age of 9.3 ka B.P. (Table 3).

The 'Fjord stage' and the attainment of the present ice-margin position

After the recession from the marine-based ice margin in Disko Bugt to the uplands of the islands and peninsulas in the eastern interior part of Disko Bugt, the morphological environment of the ice margin changed drastically. Calving became restricted to a few fjords, and large parts of the ice margin became land based where ice loss was mainly by superficial melting in the ablation zone (Weidick 1985).

A group of moraines, kame terraces and other ice contact features were identified during reconnaissance mapping of West Greenland Quaternary deposits in the late 1960s and 1970s (Figs 22, 24). These generally N–S-trending ice margin deposits occur locally throughout the inner part of the Greenland coastland from 64° to 70°N. It was already clear from the first description (Weidick 1968), that the deposits of this former ice-margin zone were formed over a longer period, and distinction was made between an older 'Marrait moraine system' and a younger 'Tasiussaqaq moraine system' (Kelly 1985). The older system formed contemporaneously with a local marine limit of *c.* 75 m, whereas the younger system formed when the relative sea level was *c.* 40 m a.s.l.

A group of dates in the south-eastern corner of Disko Bugt (Fig. 22) have been obtained from basal gyttja deposits, mainly from isolation basins, but including one lake situated above the marine limit (Long & Roberts 2002). It appears that the island of Akulliit was ice-free before *c.* 9.6 ka B.P., whereas a moraine on the Nuuk peninsula, 5 km further to the east-north-east, was formed around 8 ka B.P., perhaps related to the 8.2 ka B.P. cold event (Fig. 23).

The radiocarbon age of 9.6 ka B.P. from SE Disko Bugt provides a minimum date for the local deglaciation, but this age is not related to ice-margin deposits (Table 2). However, *c.* 30 km farther to the north, four ages of 9.9–9.3 ka B.P. have been obtained on shells from cliffs in basal marine silt along the shores between the Narsarsuaq plain south of



Fig. 24. Upland landscape at *c.* 600 m a.s.l. between Kangia and Paakitsoq fjord, east of central Disko Bugt, looking towards the north-west. The person is standing on the central part of an interlobate moraine ('The Fjord stage'). The moraine is a part of a system of ice-margin features extending from the iceberg bank at the mouth of Kangia north-eastwards to the mouth of Paakitsoq. The present margin of the Inland Ice is visible in the right background. This area was investigated for a potential hydro-power plant in the 1980s. R.E. Peary and C. Majgaard visited the ice margin here in 1886. Photograph by A. Weidick in 1963.

Ilimanaq (Claushavn) and Ilulissat (Figs 22, 25; Table 3). This area can be characterised as an upland with altitudes up to 400–600 m a.s.l.; the mountain ridges are partly till-covered and in the intervening valleys, marine silt is overlain by glacio-fluvial deposits related to the subsequent Tasiussaq moraine system. A straightforward correlation of the numerous moraine remnants cannot be made. Outwash deposits and beach ridges up to 60–70 m a.s.l. related to the most westerly of these moraine remnants can now be dated to around 10–9 ka B.P. according to the emergence curves of the area (Long *et al.* 1999). The deposition of the basal silt must also be related to this period. The three dated deposits south of Kangia of 9.9 (Ua-4574) and 9.8 (K-2023 and K-1818) ka B.P. are all situated in west-facing coves (Fig. 25). The northernmost dated sample from near Ilulissat of 9.3 ka B.P. (Ua-1086) was taken from marine silt underlying the archaeological site at Sermermiut (Fig. 26). The archaeology and palaeobotany of this site have been described by Larsen & Meldgaard (1958) and Fredskild (1967).

Farther north, there is clear evidence that the western coast of Arveprinsen Ejland (Alluttoq) was ice-free before 9.7–9.9 ka B.P. (Long *et al.* 1999), which implies that the position of the Marrait moraine system could be at the mouth of the Torsukattak fjord (Fig. 22).

The younger Tasiussaq moraine system in Disko Bugt can be followed from the Nuuk peninsula, mentioned above and described by Long & Roberts (2002), northwards along most of the bay. Morphologically, it is characterised by marginal moraines and wide alluvial plains.

The dated sites are from south to north (Fig. 25):

1. The Kangersuneq fjord, 22 km north of the Nuuk peninsula, where a shell sample gave a date of 7.9 ka B.P. (Ua-4572, Table 3). The shell material was collected in a cliff 15–20 m a.s.l. near Serfarsuit at the head of the Kangersuneq fjord. The date gives a minimum age of the deglaciation of this site.

2. Two samples from a locality 15 km farther to the north in Eqaqut (Laksebugt) gave dates of 8.7 and 8.5 ka B.P. (Ua-4573 and K-993, Table 3; Fig. 25). Both are from marine silt covered by gravel of an alluvial plain at altitudes of 38–40 m a.s.l. according to recent detailed mapping of the area by GEUS. The elevation of the sampling sites, determined by altimeter, was stated to be 50–55 m by Weidick (1968). The alluvial plain is related to the ice margin features of the Tasiussaq moraine system.
3. From south of Claushavn (Ilimanaq) and north of Ilulissat, dates on marine shells have been obtained by different authors (Table 3; Fig. 25). In addition to the group of dates between 9.9 and 9.3 ka B.P. mentioned above, a younger group of shell samples, dated to 8.6–7.7 ka B.P., appear to be related to a marine level around 40 m. The marine sediments at this elevation were laid-down before the alluvial plains of the Tasiussaq moraine system. This is most markedly seen at the Narsarsuaq ('Lersletten') plain (Table 3), where details of braided rivers and dead ice holes have been recognised (Weidick 1968).

Two exceptions to the younger dates should be noted: (1) A basal gyttja from a lake sediment core in an oxbow lake of the alluvial plain of Narsarsuaq gave an age of 8.8 ka B.P. (K-987; Kelly in: Tauber 1968), which is older than the underlying marine sediments (K-992, 7.9 ka B.P.). The date of the lake sediment is thus presumed to be too old, possibly due to hard-water effects or reworked older carbon. (2) A shell sample dated to 7.7 ka B.P. that was collected at an elevation of 30 m a.s.l. (I-6243, Table 3); this sample is presumed to relate to a marine level at or below this height.

Near Qajaa, a minimum age for the deglaciation is provided by gyttja dated to 8.8–8.0 ka B.P., and the Tasiussaq moraine system must be older than this. Farther north, there is evidence that Paakitsoq was ice-free before 7.7 ka B.P. (Fig. 22; Long *et al.* 2006). The marine limit at these sites was found to be around 40 m a.s.l.

Little is known about the age of the moraine systems in the northern parts of Disko Bugt. Two radiocarbon dates of marine shells from the inner parts of the fjords in this region are available. One was obtained from the coast near the southern flank of the outlet Eqip Sermia (Fig. 22),

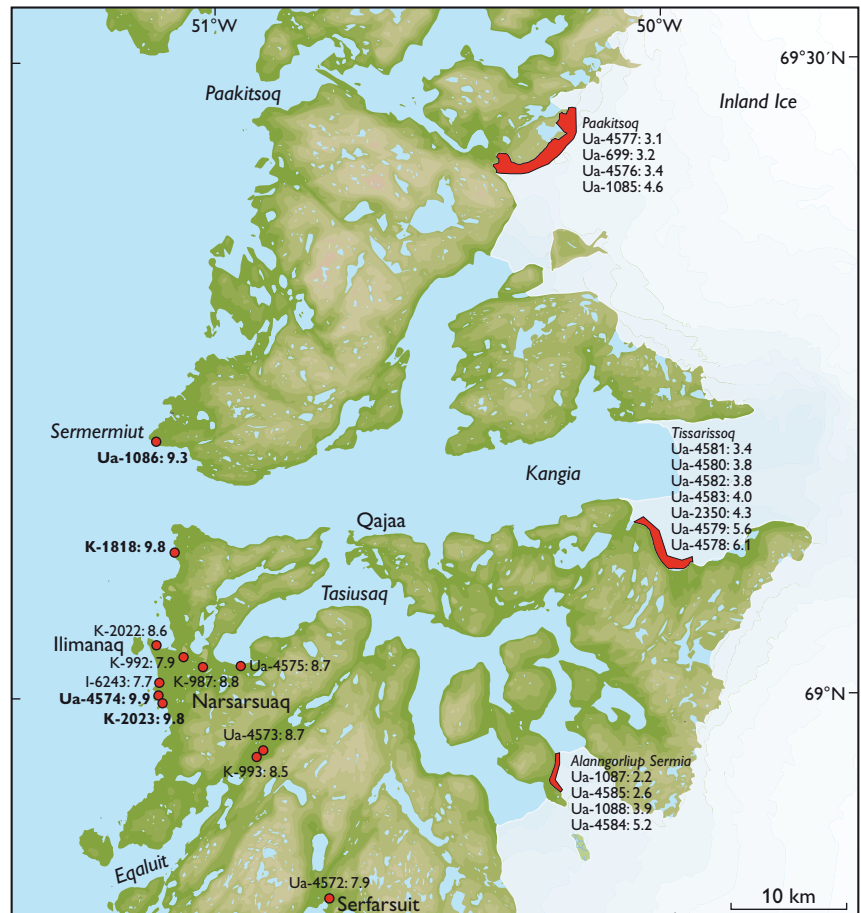


Fig. 25. Radiocarbon dates related to Holocene ice-margin deposits in the Kangia region. The older group (dates in bold: 9.9–9.3 ka B.P.) is related to the Mairait moraine system whereas younger dates (8.7–7.7 ka B.P.) are related to the Tasiussaq moraine system. The 6.1–2.2 ka B.P. dates along the ice margin relate to recession during the Holocene thermal maximum. For details, see Tables 3 and 4. Note that the map is based on 1994 ice-margin data (cf. Fig. 13).

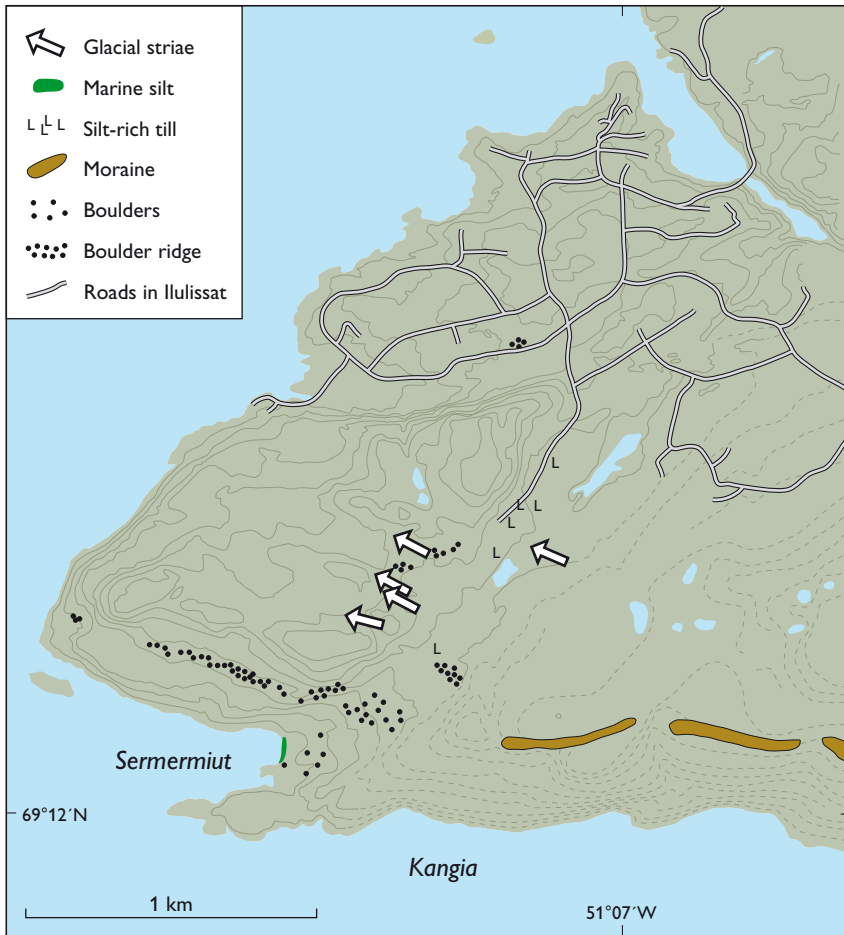


Fig. 26. Sketch-map of the area around Ilulissat (Jakobshavn) and Sermermiut showing Quaternary features. The map is based on the Geodetic Institute (now part of National Survey and Cadastre, Copenhagen) map sheets 1:2000, Jakobshavn and 1:8000, Sermermiut. Contour interval is 10 m. Redrafted from Weidick (1969).

where shells collected 2–3 m a.s.l. yielded an age of 7.3 ka B.P. (K-6373; Rasch 1997), while another shell collection from Qapiarfitt on the south side of the outlet Kangilerngata Sermia (Fig. 22) at the same elevation gave an age of 8.4 ka B.P. (K-3663; Ingólfsson *et al.* 1990). It appears, therefore, that the attainment of the present ice-margin position in the fjords in this region here was reached as early as about 8 ka B.P. This implies that the zone of moraine deposits in this region diverges, so that the northern correlates of the Tasiussaq moraine system are found in the interior parts of the fjords, close to the present ice margin.

A high concentration of ice-contact features is usually interpreted as indicative of deposition at a 'stable' ice margin. The dates of the 'Fjord stage' cover a time of *c.* 2 ka (*c.* 9.9–7.9 ka B.P.), with a net recession of the ice margin in the central and southern parts of Disko Bugt of only a few kilometres. The two-fold division of the Fjord stage into the older Marrait moraine system and a younger Tasiussaq moraine system might reflect two phases of development, with the older primarily due to decreased ablation and calf-ice production following the reduced contact between the

sea and the ice margin. Another factor is the complexity of the response of the ice margin to climate change. The response of a continental ice sheet and its marginal positions depend on long-term changes in ice flow (on centennial to millennial time scales) due to sustained changes in accumulation and surface temperatures. However, short-term annual to decadal changes in mass-balance elements, such as accumulation, run-off and iceberg calving, also play important roles (Reeh 1999; Dahl-Jensen 2000). All these elements are present in the case of the early Holocene change of the ice cover in Disko Bugt. Thus the abrupt increase in snow accumulation over Greenland at the end of the Younger Dryas, as documented by Alley *et al.* (1993), may have had a positive effect on the mass balance and response of the ice margin, which could counter the effect of the subsequent temperature increase.

The abrupt temperature rise of 10–15°C from around 11.7 to *c.* 10 ka B.P., that caused the break-up of the ice cover over Disko Bugt at *c.* 10 ka B.P., must have led to a change of the ice-margin profile. Increased ablation, especially at lower levels of the ice, would lead to a steeper slope

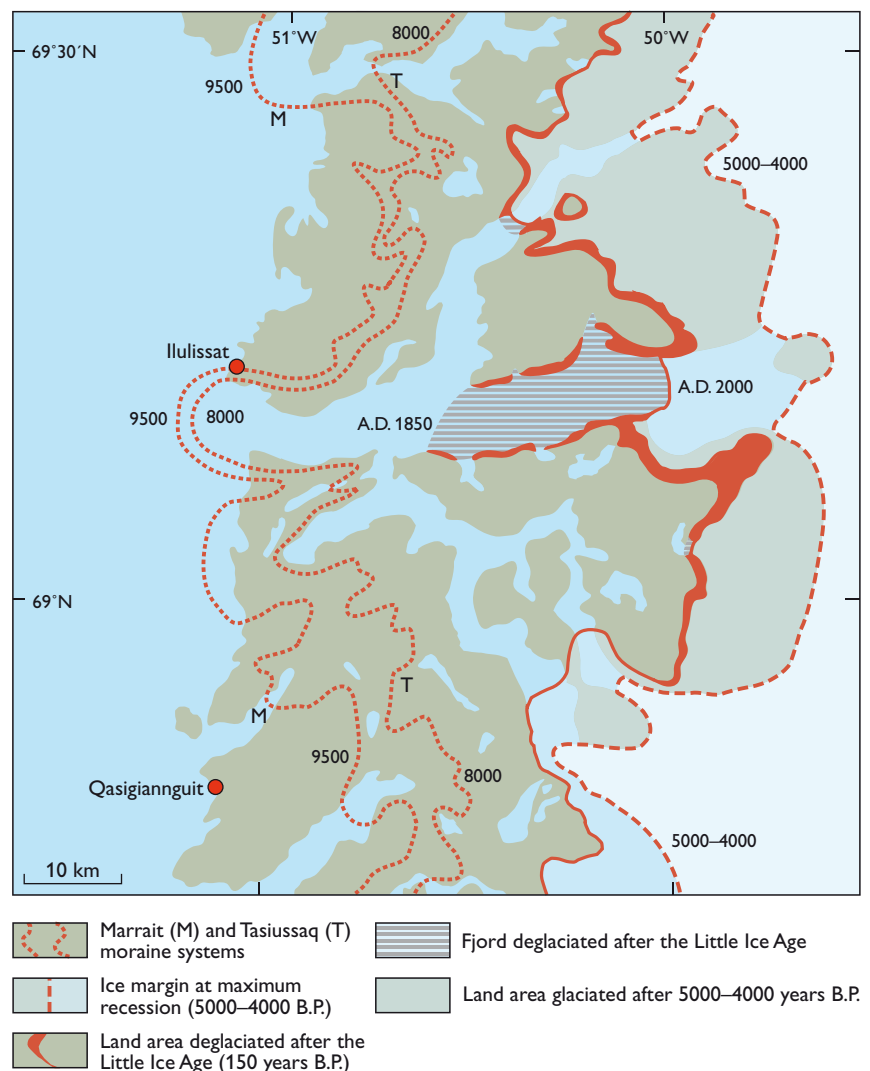
of the ice margin and also to a reduction of the ablation zone, until changes in the ice dynamics led to a new quasi-equilibrium. Studies of the effects of the present centennial temperature rise on the ice margin indicate that such changes of geometry are currently taking place (Thomas *et al.* 2001; Bøggild *et al.* 2004; Hughes 2004). During the build-up of both the Mairait and the Tasiussaq moraine systems (c. 9.9–7.9 ka B.P.), the front of Jakobshavn Isbræ was resting on or at Isfjeldsbanken at the mouth of Kangia. Iceberg production, estimated from the frontal area determined by the depth of the iceberg bank (200–300 m) and the trend of the moraines, must have been reduced during this period (Weidick 1994a, b). During the subsequent recession, the ice margin may have reached the position of the present location at or before 6–7 ka B.P.

Palaeoceanographic investigations show that large volumes of meltwater produced by Jakobshavn Isbræ deflected

the warmer waters of the West Greenland Current away from the eastern coastal areas of Disko Bugt until 7.9 ka B.P., when the glacier finally receded from Isfjeldsbanken to the interior of the present Kangia fjord system. This recession allowed the warmer West Greenland Current to penetrate to the eastern part of Disko Bugt, as can be seen from changes in the fauna (Lloyd *et al.* 2005). The relationship between the recession of the ice margin and the warming of the bay must be rather complex. After the recession from the Tasiussaq moraine system, where numerous drainage channels led to the formation of alluvial plains with the main drainage from the ice-sheet margin directly to Disko Bugt, the drainage became concentrated into a few channels that drained into the deep fjords of Kangia and Torsukattak.

The first slow-down of the ice recession is marked by the Mairait moraine system (around 9.9–9.4 ka B.P.), which probably developed in response to the change of environ-

Fig. 27. Provisional reconstructions of the position of the ice margin in the Ilulissat area at c. 9500, 8000, 5000–4000 and 150 years B.P. The change of the ice margin between the Little Ice Age maximum (150 years B.P.) and the present day (the trimline zone) is shown in red for land areas and horizontal shading for floating glaciers. Note that the zone without vegetation (the trimline zone) becomes narrow south of Jakobshavn Isbræ and almost absent farther south.



ment and a much reduced output of calving ice in the central and southern parts of Disko Bugt, and possibly also a changed profile of the ice margin. The halt or slow-down may have been prolonged during the deposition of the Tasiussaq moraine system (around 8.8–7.9 ka B.P.), influenced by the 8.2 ka B.P. cold event (Fig. 23; O'Brien *et al.* 1995; Dansgaard 2004). This event was the most marked cold episode during the early Holocene, and in Greenland it was probably associated with the formation of the moraine on the Nuuk peninsula in the south-east corner of Disko Bugt (Long & Roberts 2002). The 8.2 ka B.P. event was probably related to the release of large amounts of cold meltwater from the receding Laurentide ice sheet.

The subsequent recession of the ice-sheet margin from the Tasiussaq moraine system to its present position is not known in detail. The recession began at around 8 ka B.P., and the attainment of the present ice-margin position is estimated to have occurred at or before 6–7 ka B.P. for the Jakobshavn Isbræ area (Weidick *et al.* 1990). This is based on the minimum age of the Tasiussaq moraine system of about 8 ka B.P. (Fig. 27), and on dates of marine material (mainly shells) transported westwards by the ice to the present margin of the ice; the age range of these dates is shown in Fig. 25. It can be seen that the dated samples cover a time span from 6.1 to 2.2 ka B.P., during which time the ice margin was east of its present position.

When attempting to correlate the individual stages in the recession and the subsequent Neoglacial advance of the ice margin, the scant evidence from the regions north and south of Disko Bugt should also be considered. In the Uummannaq Fjord complex to the north, it is known that the ice margin was situated well into the fjord complex at about 10.7–10.5 ka B.P. (Símonarson 1981; Bennike 2000; Bennike & Björck 2002). The high relief of the fjord landscape, the deep fjords, and the many still productive calving glaciers (Fig. 5) might well explain an early recession, although the time at which the glaciers reached their present position is unknown.

South of Disko Bugt, the lowlands east of Aasiaat provide little information about the recession history of the ice margin. The southern shores of Disko Bugt were deglaciated prior to 10.5 ka B.P. (Donner & Jungner 1975; Long *et al.* 2003). A moraine system is found *c.* 25 km west of the present front of Nordenskiöld Gletscher (Fig. 2), and seems to be related to the large alluvial plain of Naternaq (Lersletten). The alluvial plain overlies marine deposits (e.g. Harder *et al.* 1949; Laursen 1950), but little systematic work has been carried out here. The altitudes of the lakes on this plain are 37–54 m a.s.l. (GEUS 2004), and it is presumed that the Tasiussaq moraine system relates to a sea level about 50–40 m above the present day. A mor-

phological extension of these moraines to the south might correspond to the locality near Ussuit (67°51'N, 50°16'W; Fig. 1), from where Kelly (1973) obtained a date of 7.6 ka B.P. on marine shells. These shells appear to be related to a marine level of *c.* 42 m a.s.l., and provide a minimum age for the adjacent moraine system.

As the onset of deglaciation of the outer coast south of Disko Bugt took place well before *c.* 10.5 ka B.P., the subsequent net recession of the ice margin probably occurred at approximately the same rate as in the northern parts of Disko Bugt. However, local temporal variations in the recession must have characterised the region. During the recession over the lowlands between Arfersiorfik fjord in the south and Disko Bugt in the north, the relative sea level was 50–100 m above the present. Although an extended contact between the ice margin and the sea can be envisaged, the region was a shallow area compared to the depths in Disko Bugt. In the area south of Disko Bugt, attainment of the present ice-margin position has been estimated to have occurred at about 8 ka B.P., with the ice margin subsequently receding eastwards beyond its present position (Letréguy *et al.* 1991b). The calculation for the Holocene thermal maximum at around 5 ka B.P. gives a recession of the ice margin in this region of the same order of magnitude as that recorded at Jakobshavn Isbræ.

The reason for the high sensitivity of the ice margin to climate change may be that this region is characterised by extensive coastal areas of low elevation and low accumulation. The position of the ice margin is thus essentially controlled by ablation, resulting in a rapid response to any early Holocene warming. In fjords or bays (such as Disko Bugt), as well as in lowland areas such as those south of Disko Bugt, the rapid Holocene recession is mainly related to increased ablation. Other factors, as noted above, may be important, and can explain local temporary deviations from the general case; these must be considered when constructing a more detailed history of ice-margin changes. Roberts & Long (2005) have suggested a convergent ice-flow drainage pattern through the shallow troughs in the southern part of Disko Bugt, and through the shallow fjords south of Disko Bugt, related to a large drainage outlet through the outer Egedesminde Dyb. It is not clear, however, when ice drainage of this magnitude could have taken place.

The ice-sheet margin after the Holocene thermal maximum

Collections of marine fossils from Little Ice Age moraines at, and close to, the present ice margin have been made

Table 4. Neoglacial radiocarbon age determinations (see also Table 2, Appendix 2)

Locality/Material	Laboratory number*	^{14}C age† ($\pm 2\sigma$) years B.P.	Rcorr ^{14}C age§ years B.P.	Calib. age¶ ka B.P.	$\delta^{13}\text{C}$ ‰	Reference
Paakitsoq, c. 69°25'N, 50°20'W						
<i>Macoma calcarea</i> shell	Ua-4577	3300 \pm 65	2900	3.1	-6.79	This study
Shell	Ua-699	3420 \pm 105	3020	3.2	0	Weidick <i>et al.</i> (1990)
<i>Mytilus edulis</i> shell	Ua-4576	3560 \pm 65	3160	3.4	-3.73	This study
Shell	Ua-1085	4520 \pm 135	4120	4.6	0	Weidick <i>et al.</i> (1990)
Tissarissoq, c. 69°06'N, 50°02'W						
<i>Mya truncata</i> shell	Ua-4581	3590 \pm 65	3190	3.4	1.88	This study
<i>Hiatella arctica</i> shell	Ua-4582	3940 \pm 65	3540	3.8	1.92	This study
<i>Mya truncata</i> shell	Ua-4580	3945 \pm 70	3545	3.8	2.50	This study
<i>Mya truncata</i> shell	Ua-4583	4075 \pm 70	3675	4.0	2.16	This study
<i>Odobenus rosmarus</i> tusk	Ua-2350	4290 \pm 100	3890	4.3	-13.05	Weidick (1992a)
<i>Mya truncata</i> shell	Ua-4579	5240 \pm 75	4840	5.6	1.85	This study
<i>Balanus</i> sp. plate	Ua-4578	5710 \pm 55	5310	6.1	0.98	This study
Alanngorliup Sermia, c. 68°54'N, 50°15'W						
Shell	Ua-1087	2620 \pm 110	2220	2.2	0	Weidick <i>et al.</i> (1990)
<i>Mya truncata</i> shell	Ua-4585	2935 \pm 60	2535	2.6	2.01	This study
Shell	Ua-1088	4000 \pm 115	3600	3.9	0	Weidick <i>et al.</i> (1990)
<i>Mya truncata</i> shell	Ua-4584	4930 \pm 60	4530	5.2	1.78	This study

* Ua: Ångström Laboratory, Uppsala.

† The radiocarbon age determinations have been corrected for measured or assumed isotopic composition by normalising to -25‰ on the PDB scale.

§ Rcorr: Reservoir-corrected. Corrected for a seawater reservoir effect of 400 years.

¶ Calibrated using the INTCAL04 dataset (Reimer *et al.* 2004) and the OxCal. v.3.10 software program (Bronk Ramsey 2001).

^{||} Assumed value.

around Jakobshavn Isbræ. The detailed subsurface maps of this area make it possible to follow the trends of the fjords beneath the ice, to establish the glacial transport route of the dated material, and hence to calculate that at maximum recession, the ice margin was some 15–20 km eastwards of the present position (Weidick *et al.* 1990; Weidick 1992a). The ages of 6.1–2.2 ka B.P. provide a minimum estimate for the period during which the ice margin at most localities was situated east of its present position (Table 4; Fig. 25). The ice margin presumably gradually advanced after the end of the Holocene thermal maximum at 5–4 ka B.P., although exceptions may have occurred locally. The Narssarsuaq moraine system near Narsarsuaq in South Greenland (Weidick *et al.* 2004b; Bennike & Sparrenbom 2007) and the Drygalski moraines (Fig. 22) crossing the root of the Nuussuaq peninsula north of Disko Bugt (Fig. 22; Kelly 1980), may be related to early Neoglacial events. If so, the margin of the ice sheet may locally have advanced beyond the present position in some places during early phases of the Neoglacial. The age of the Narssarsuaq moraine system is estimated to be *c.* 2 ka B.P. (Weidick *et al.* 2004b; Bennike & Sparrenbom 2007), whereas the age of the Drygalski moraines is unknown.

Investigations of sediment cores sampled west of Isfjeldsbanken indicate maximum Atlantic water influence during the period from *c.* 1650 to 500 calendar years B.P.,

which is related to a recession of the front of Jakobshavn Isbræ during the medieval warm period (Lloyd 2006).

Modelling of ice-margin changes for the last 1400 years is based on information on the surface and subsurface topography of the ice margin, palaeoclimate data from ice cores, measured weather data, and the rheology of the ice (Fig. 28). The calculations by Reeh (1983) demonstrate that quite short lengths of the ice margin exhibit local variations in behaviour. Two major advances took place, one at about A.D. 800 and another from A.D. 1500 to 1900 related to the Little Ice Age. The latter was often associated with advances, notably at about 1750 and in the late 1800s. Both of these advances can be observed in the response curves, with the older advance often apparently the more prominent. However, older moraines are rarely observable in the field, since they are often buried beneath younger ones. The difference between the *c.* 1750 and the *c.* 1900 maxima is only about 100–200 m. For the last 100–200 years, the modelled ice-margin changes agree with observational data, because fresh deglaciated terrain with a width of about 1 km is found.

This 1–2 km width of the ‘historical advance and recession’ of the ice after the middle of the 19th century is commonly quoted, but in reality large variations occur. An unusually large width is seen around Jakobshavn Isbræ (Fig. 29), where fresh moraines and ice-polished bedrock

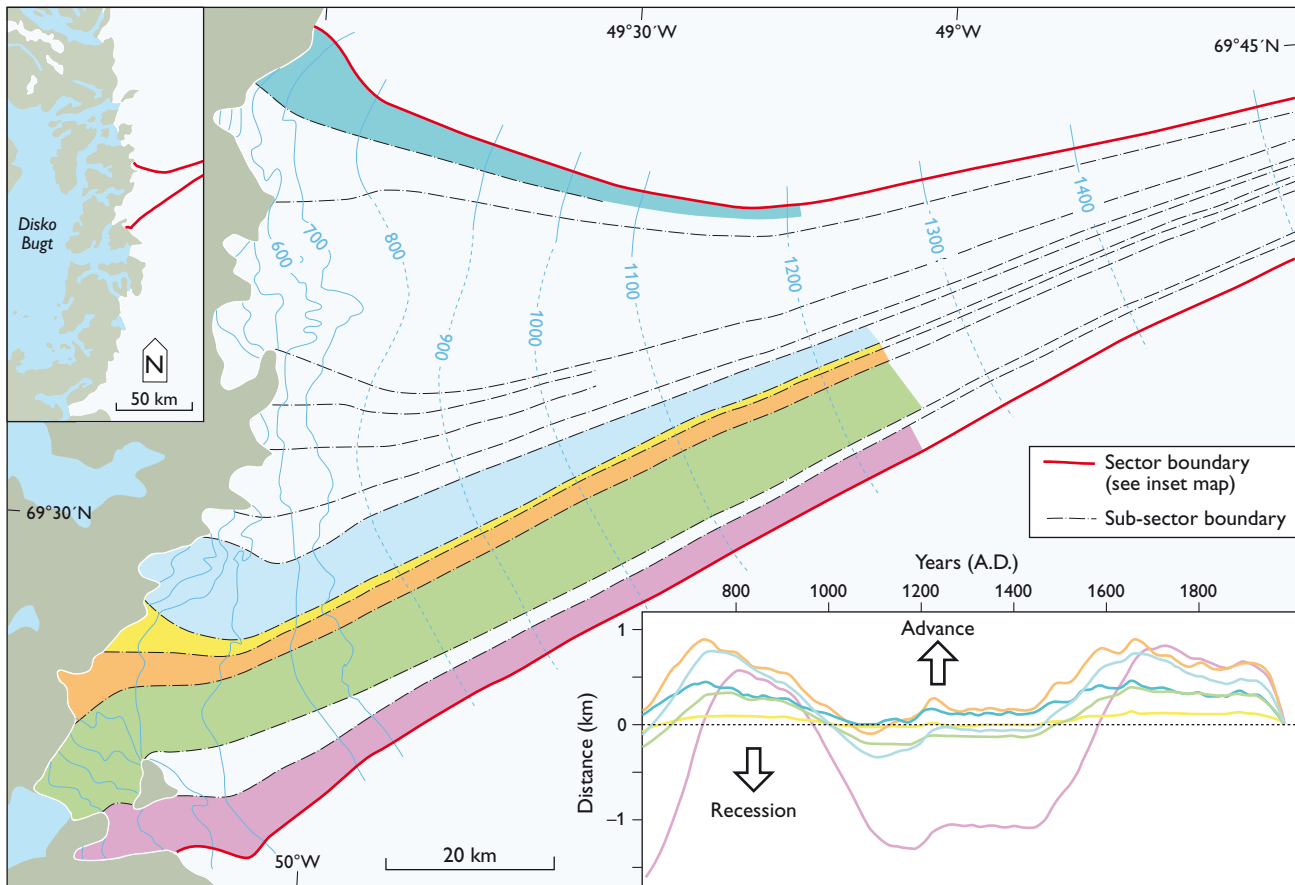


Fig. 28. Advances and recessions in the ice-sheet margin can be correlated with climatic variation deduced from the Dye 3 ice-core record. Lateral variation in the response was defined for a specific sector of the margin, north of Jakobshavn Isbræ (see inset map); sub-sectors (indicated by colours) show varying degrees of response, indicated on the inset, lower right. From Reeh (1983).

(the trimline zone) extend for over 30 km in front of the present glacier. The thinning of the ice, estimated from the height of the trimline zone, is 200–300 m around Jakobshavn Isbræ (Figs 27, 29). By contrast, just *c.* 25 km south of Jakobshavn Isbræ, the trimline zone nearly disappears around the outlets of Alanngorliup Sermia and Saqqarliup Sermia (Fig. 30). Historical records indicate nearly stationary conditions here since the middle of the 19th century (Weidick 1994a, b); the glaciers almost maintain their maximum extent from the Little Ice Age. Further south, towards Kangerlussuaq at 67°N, many lowland sectors were characterised by a minor readvance in the period between *c.* 1950 and 1985 (Weidick 1992a, 1994a, b); this could be a consequence of the temperature fall in the last decades of the 1900s, or that the ice sheet had still not adjusted to past climatic fluctuations, as modelled by Huybrechts (1994). The recent development of the trimline zone around Jakobshavn Isbræ has been studied from multispectral Landsat images (Csatho *et al.* 2005).

Observations of glacier change are unevenly distributed, with maximum change in South-West Greenland that appears to be indicative of a present warming trend. This is confirmed by observations and measurements in South Greenland (between 60° and 65°N; Mayer *et al.* 2002; Podlech 2004), while investigations in Melville Bugt (73° to 79°N, C.E. Bøggild, personal communication 2004) show the same warming trend; a similar situation may apply to Jakobshavn Isbræ (see also below).

In general, the Little Ice Age has left its mark in the form of fresh moraines, which are well defined in some regions, but not in others. Exact dating of parts of these moraines clearly points to a specific geological event; the link between moraine formation and climate, however, is not always straightforward. For a regional understanding of ice-margin history, only the whole zone or belt of ice-margin deposits can serve as a useful comparison with models of the ice-sheet response to climate change.



Fig. 29. A: Front of Jakobshavn Isbræ (northern side). The vegetation-poor zone (the trimline zone) is 200–300 m high. Photograph by J. Lautrup 1991. B, C: South side of Jakobshavn Isbræ; highest mountain is 368 m. The photograph in B (by M.C. Engell) is from 1902 whereas that in C (by A. Weidick) is from 1963. In 60 years, the glacier front has receded about 13 km to the east, and is seen faintly in the distance on the 1963 photograph.



Fig. 30. Western margin of Alanngorliup Sermia in 1988. The glacier has a steep margin close to vegetated terrain. Photograph by A. Weidick. (This glacier is recorded as Avannarleq Sermeq by Johansen & Nielsen 2001.)

It is not known to what extent the fluctuations of the ice margin in the Paakitsoq area can be correlated with the fluctuations of Jakobshavn Isbræ. However, during the period from 4.6 to 3.1 ka B.P. (Fig. 25), the ice margin at Paakitsoq was situated east of its present position. At *c.* 1200 years B.P., the ice margin advanced and reached a position near the subsequent Little Ice Age position.

It is supposed that a gradual net advance of the ice margin took place after the thermal maximum that ended around 4 ka B.P. although this was interrupted by minor stillstands or recessions, according to the modelling of Reeh (1983). Evidence for late Holocene fluctuations has also been recorded from eastern Disko Bugt (Lloyd 2006).

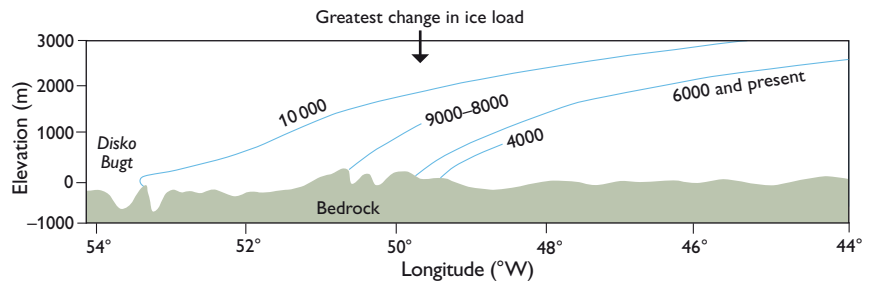
The beginning of the Little Ice Age advance may be related to the legend about Tissarissoq, the ice-filled bay south of Kangia. The name Tissarissoq is claimed to refer to a time when hunting in the bay was possible, that is to say when glacier ice did not cover the bay (Hammer 1883). This accords with a gradual glacier advance during the 1700s. However, of the 15 shell samples so far dated from the ice margin around Jakobshavn Isbræ (Fig. 25; Table 4), the youngest radiocarbon age (2.2 ka B.P.; Ua-1087) is a millennium before the Thule culture arrived in the region.

Relative sea-level changes around Disko Bugt

Thule winter houses and Norse ruins were reported to be partly below sea level by Thorhallesen (1776) and Arctander (1793), indicating recent submergence. These observations and further investigations in the first half of the 19th century by Pingel (1841, 1845) showed that this submergence had been preceded by emergence, as indicated by raised marine deposits. For the area around Disko Bugt, systematic descriptions of former raised shorelines and measurements of the following submergence were initiated in the last half of the 19th century (Steenstrup 1883a, b; Saxov 1958).

From the 1950s onwards, detailed investigations of changes in sea level, mapping of the marine limit and descriptions of marine faunas were carried out by GGU (Laursen 1950; Donner & Jungner 1975; Weidick 1975, 1976). This work has been followed up in the past few decades with more comprehensive studies of relative sea-level changes, often with support from the Arctic Station in Godhavn (Ingólfsson *et al.* 1990; Bennike *et al.* 1994; Rasch & Jensen 1997; Long *et al.* 1999, 2003, 2006; Rasch

Fig. 31. Provisional simplified profile of ice-margin stages between 10 000 years B.P. and the present day from south of Disko island to Jakobshavn Isbræ (Sermeq Kujalleq). The approximate position where the greatest change in glacier load has taken place is indicated. From Weidick (1993).



2000; Long & Roberts 2002, 2003). Much of the accumulated data was reviewed by Fleming (2000), and used for comparisons with results of geophysical modelling of the ice-sheet history in Greenland (Tarasov & Peltier 2002; Fleming & Lambeck 2004).

The relative sea-level changes observed in Greenland are mainly related to the combined effects of local glacio-isostatic responses of the Earth's crust to variations in glacier load, and global eustatic changes of sea level due to the storage and melting of ice on the continents. In the Disko Bugt region, the Holocene recession was complete by *c.* 6–5 ka B.P., and was followed by a Neoglacial expansion of the ice cover. The other major ice sheets in the northern hemisphere disappeared, the Fenno-Scandinavian ice sheet at 10–9 ka B.P., and the Laurentide ice sheet at about 8–7 ka B.P. In the Antarctic, Holocene recession of the shelf ice that began at the end of the last ice age has continued until the present day (Bindschadler & Bentley 2002); although, modelling predicts expansion during the next few centuries (Huybrechts *et al.* 2004).

Although *c.* 40% of the volume of the Greenland ice sheet has disappeared since the LGM (Huybrechts 2002), a substantial glacier load is still present in the central part of Greenland. The main losses of the glacier load after the LGM have occurred at the present ice margin (Fig. 31). By contrast, where other ice sheets have completely disappeared, the maximum glacio-isostatic uplift has occurred in what was formerly their central part.

The change of relative sea level has been dominated by Holocene emergence caused by recession and thinning of the ice margin. The altitude of the marine limit, which is defined as the maximum height of relative sea level after the last deglaciation, is usually determined by the upper limit of raised shorelines and/or the lower limit of perched boulders, or by studies of sediments in lakes situated above and below the marine limit. The trend of the marine limit indicates an elongated dome over the outer ice-free land, parallel to the present coast, with the highest values for the altitude of the marine limit in areas that show the largest Holocene recession of the ice-sheet margin (Fig. 32).

Higher values for the altitude of the marine limit than indicated on Fig. 32 were reported in some parts of the Disko Bugt region by Rasch (2000) and Long *et al.* (2006). These



Fig. 32. Elevation of the marine limit in Greenland in metres a.s.l. Compiled from Weidick (1992b), Bennike & Weidick (2001), Bennike (2002) and Weidick *et al.* (2004b).

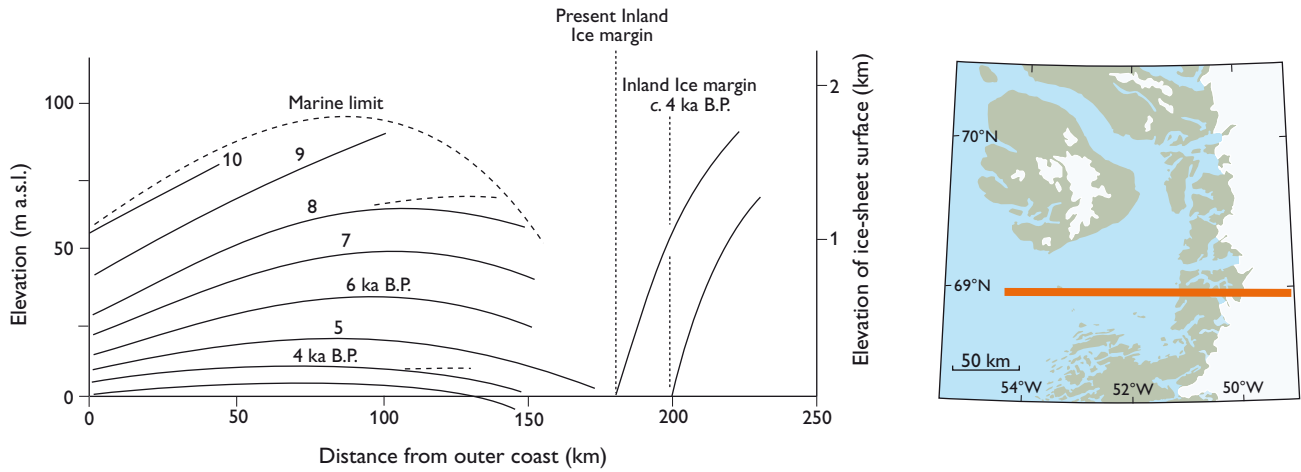


Fig. 33. Hypothetical shoreline relation diagram for Disko Bugt broadly following the 69°N line of latitude (see inset map); the position of the margin of the Greenland ice sheet at the present day and at *c.* 4 ka B.P. is indicated. Note that the dashed shorelines for 4 ka and 8 ka B.P. indicate the probable trends of these shorelines prior to the Neoglacial ice-sheet advance and accompanying proximal depression of the shoreline.

workers defined the marine limit largely on the basis of the lower limit of perched boulders. It should be acknowledged, however, that the shores in the region may locally have been affected by large waves so that the present lower limit of perched boulders may not be a true reflection of sea level. Large waves can be generated at glacier fronts by calving of icebergs, or by turnover of icebergs. The so-called kanelling, which occurred in the harbour of Ilulissat in the early parts of the 20th century (Reeh & Engelund 1971; Reeh 1985), was characterised by far-travelled large waves. Such waves have been described from several localities in Greenland; they are typically recorded from narrow fjords with calving outlets, and can reach heights over 10 m (Reeh 1985).

In addition, landslides have generated tsunamis in areas with steep slopes around the Vaigat strait and the eastern shores of Disko. In 2000, a tsunami that resulted from a landslide in the Vaigat strait had a run-up height of 50 m close to the landslide and a run-up height of 28 m at a distance of 20–25 km from the slide (Dahl-Jensen *et al.* 2004).

Minimum values for the marine limit in the central parts of Disko Bugt can be deduced from the uppermost marine terraces and beach ridges, which are found at 70–80 m a.s.l. This corresponds to an age of around 10–9 ka B.P. according to the relative sea-level curves of Long *et al.* (2006), which is close to the minimum ages for the last deglaciation (Fig. 22).

While the number of observations has increased considerably in recent years, the previously recorded maximum value of the marine limit south of Disko Bugt has been largely confirmed. However, the different versions of the marine limit that have appeared over the past decades

are still only generalised views, based on an uneven distribution of observations (Rasch 2000, Long *et al.* 2006). Data coverage is relatively good around Disko and Disko Bugt, but very scattered farther north. Only few observations are therefore available for the Ummannaq Fjord complex. Future detailed mapping of the marine limit will probably give a more varied picture (e.g. Ingólfsson *et al.* 1990). A fall in the marine limit from 85 m to 54 m a.s.l. over a distance of 8 km was reported for the south-eastern corner of Disko Bugt by Long & Roberts (2002), the change being attributed to a slowdown of ice-margin recession (cf. the Fjorde stage).

The dating of the limit at any one locality is usually based on extrapolation of local relative sea-level curves. The age of the marine limit in West Greenland decreases from the outer coastland towards the east (Fig. 33). Westwards, an apparent convergence of strandlines is seen, although their trend is often uncertain. A reverse trend has been suggested for the west coast of Disko (Funder & Hansen 1996), such that the marine limit becomes younger westwards. This was based on the occurrence of transgressive sequences in the area (Ingólfsson *et al.* 1990), and is comparable to transgressions that have been reported from western Norway (Andersen 1965; Kaland *et al.* 1984).

The form and trend of the marine limit are determined locally by the former glacier load and by the subsequent rate of thinning and recession of the ice cover. Detailed determinations of local sea-level changes, and the spatial trends of the individual isobases, are important for understanding the development of the landscape.

Compared to other regions in Greenland, the amount of detailed data on relative sea-level changes in the Disko

Bugt area is large (Rasch 1997, 2000). The older relative sea-level curves were mainly based on dates of marine shells (Donner & Jungner 1975; Donner 1978; Weidick 1996; Rasch & Jensen 1997). The oldest dates at any locality provide minimum ages for the local deglaciation and of the marine limit. However, the relationship between localities with fossil marine shells and the contemporaneous sea level is somewhat uncertain, and numerous sample localities from a large area are needed to provide enough data points. A more recent series of detailed curves has been constructed from isolation basins (Long *et al.* 1999, 2003, 2006; Long & Roberts 2002, 2003). The constructed uplift curves indicate a steady emergence throughout the early and mid-Holocene, followed by a late Holocene submergence, presumably caused by the advancing ice margin and increasing glacier load.

The relative sea-level curves indicate a larger initial emergence to the east, near the present Inland Ice margin, than farther west, and the hypothetical shoreline diagram has been drawn on this basis (Fig. 33). The north–south trend of the marine limit suggests the trend of the isobases should be broadly parallel to the present ice-sheet margin, but locally a more complicated pattern than shown in Fig. 33 can be expected (Rasch 2000).

A more exact and site-specific method of dating relative uplift is by dating of the timing of isolation of lakes at different altitudes (Fig. 34). This procedure has been carried out at six localities around Disko Bugt (Fig. 2): the Vaskebugt (Kangerluarsuk) area on Arveprinsen Ejland (Long *et al.* 1999), Akullit/Nuuk in the south-east corner of Disko Bugt (Long & Roberts 2002), Qeqertarsuatsiaq in south-western Disko Bugt (Long & Roberts 2003), Innaarsuit on southern Disko (Long *et al.* 2003), and near Qajaa and at Paakitsoq in eastern Disko Bugt (Long *et al.* 2006). The main drawbacks of this method are that it is time consuming, and obviously it can only be applied to areas where lakes exist at different elevations below the marine limit. Regional correlation of locally determined uplift may well be substantiated through geomorphological correlation of strandlines in the area.

From the investigations referred to above, it has been established that the present sea level was reached by 5–4 ka B.P. Emergence continued for some time, and the lowest relative sea level was reached at around 3–1 ka B.P. when it was about 5 m below the present; this was followed by the beginning of the present submergence. The period after *c.* 4.5 ka B.P. coincides with the period of human settlement in Greenland, and a number of the earliest known ruins are at or below water level at present high tide (Larsen & Meldgaard 1958; Rasch & Jensen 1997). The exact form of the late Holocene part of the relative sea-

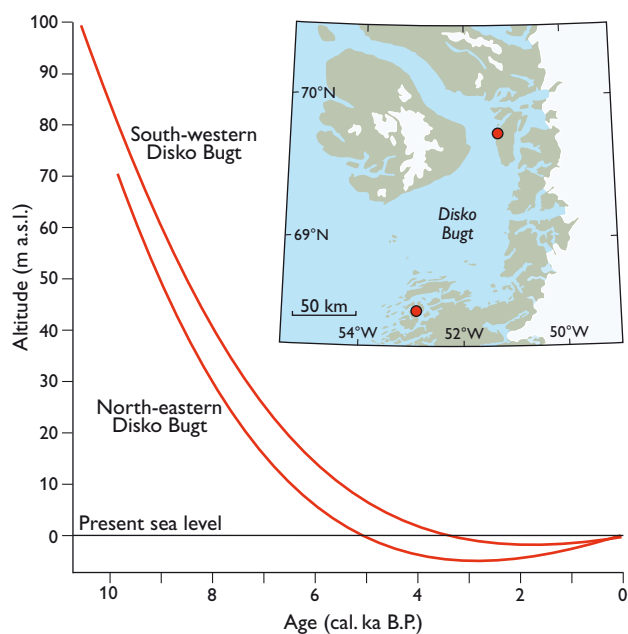


Fig. 34. Examples of two relative sea-level curves from the Disko Bugt area. The inset map shows the location of the sites in the Disko Bugt region. Modified from Long *et al.* (1999) and Long & Roberts (2003).

level curves is difficult to establish because the curves are flat, and the transition from emergence to submergence took place at shallow depth. At Tuapaat on southern Disko island, morpho-stratigraphic investigations of the coastal landscape by Rasch & Nielsen (1995) suggested that three or four transgressions have taken place during the past 2.5 ka B.P. Sea-level measurements were initiated at Godhavn in 1897, and demonstrate a subsidence of 0.475 m up to 1946, whereas an emergence of 0.3 m was recorded for the time period between 1946 and 1957 (Saxov 1958; Kelly 1980).

On the basis of repeated GPS observations in West Greenland between 1995 and 2002, the present-day vertical crustal movements have been determined (Dietrich *et al.* 2005). At Ilulissat an uplift of 1.6 mm/year was observed, which is presumably due to the recent thinning of the Inland Ice in this region. Uplift was also recorded at the outer coast south of Disko Bugt. In contrast, marked subsidence characterised the inland region south of Disko Bugt, with rates up to 4 mm/year.

Glaciology

From the geological section, it is clear that a presentation of the history of the Jakobshavn Isbræ ice stream must also include the history of the surrounding parts of the ice sheet. In the same way, the present major ice streams are a part of the local glaciation history of Disko Bugt.

Subsurface of the ice-sheet margin

The first mapping of large parts of the ice sheet was made in the period 1949–1951 during the ‘Expéditions Polaires Françaises’ using seismic methods (Holtzscherer & Bauer 1954). Their surveys covered the southern parts of the Inland Ice (south of *c.* 70°N in West Greenland and south of *c.* 72°40′N in East Greenland) and indicated a depression below present sea level in the central part of the ice sheet as well as a drainage channel towards Disko Bugt (Figs 4, 17). Later airborne radar surveys between 1968 and 1976 provided a measure of the thickness of the entire ice sheet (Gudmandsen & Jakobsen 1976; Overgaard 1981). These data provided a realistic impression of the gross features of the subglacial terrain, including highlands, uplands, lowlands, and drainage areas (Figs 4, 17). More recent data on the subsurface of the ice sheet are steadily improving our knowledge of the subglacial landscape (Bamber *et al.* 2001). In general, the elevation of the subglacial terrain falls from the marginal areas towards the central depression of the ice sheet, and as noted above, a depression connects Kangerlussuaq with the interior region, lying at or below sea level, (Fig. 4).

Ice streams have been defined as “part of an ice sheet, in which the ice flows more rapidly and not necessarily in the same direction as the surrounding ice” (Armstrong *et al.* 1973, p. 26). The strong flow of ice streams, relative to the slower moving ice on either side, leads to strong and chaotic break-up of the ice in ice streams. This limits penetration by radar waves, and hence the determination of ice thickness below ice streams. However, the thickness of Jakobshavn Isbræ has been determined by seismic methods (Clarke & Echelmeyer 1996). In the central parts of this ice stream, the ice thickness varies from *c.* 1.9 km near the grounding zone (cf. Fig. 35) with the ice surface at an altitude of *c.* 500 m a.s.l., to about 2.5 km at a distance of 40 km behind the grounding zone where the ice surface is *c.* 1000 m a.s.l. Thus the bottom of the subglacial trough is found at a depth of *c.* 1.5 km below sea level; the trough can be described as a canyon-like feature about 7 km wide,

surrounded by a hilly subglacial landscape with elevations close to sea level. Much of the basal interface is probably underlain by compacted, non-deformable sediments (Clarke & Echelmeyer 1996). Farther inland, the trough beneath the ice stream gradually levels out. About 120 km east of the grounding zone, the subsurface depression can be interpreted as a shallow subglacial valley. The subsurface map of Fig. 4 also shows that other subglacial areas near the present ice margin in West Greenland are dominated by uplands, with no clear indication of other ice streams of the size of Jakobshavn Isbræ.

Supplementary depth soundings have provided more detailed information around the deep drilling sites in the central parts of the ice sheet and in marginal areas studied in connection with possible exploitation of hydropower resources. For example, a *c.* 800 km² area of the ice margin in the Paakitsoq area, north of Ilulissat, was mapped at a scale of 1:250 000 with 100 m contour intervals (Thomsen *et al.* 1988). Other less detailed maps cover smaller areas of the ice margin at Tassarisoq (south of Ilulissat), Alanngorliup Sermia and Saqqarliup Sermia (Thorning *et al.* 1986; Thorning & Hansen 1987; Weidick *et al.* 1990). One revelation from this detailed mapping was that even for ice thicknesses of 600–800 m, the depressions and elevations of the subsurface are reflected in the topography of the ice surface, although in a smoothed and somewhat distorted form. It thus appears that topographical features of the ice-free marginal areas bordering the ice sheet continue beneath the ice, and are readily discernable on Landsat images with a low sun angle (Fig. 36). The deep trough of Jakobshavn Isbræ is clearly visible from its abundant crevasses (dark colour, relative high ablation), in contrast to the other less important ice streams draining into Disko Bugt (Sermeq Kujalleq and Sermeq Avannarleq in Torsukattak icefjord). The continuation of Torsukattak icefjord beneath the ice sheet rapidly levels off into the subglacial uplands of the area, as seen from the depth of the present fjord (*c.* 600 m) and radar soundings 25–30 km from the front (Overgaard 1981); this picture is confirmed by the image of Fig. 4. Restricted ablation and high relief may explain the local high production of calf ice from the glaciers draining into Uummannaq Fjord farther to the north (Fig. 5). Snowfall here is also heavier than in the interior parts of Disko Bugt (Ohmura & Reeh 1991).

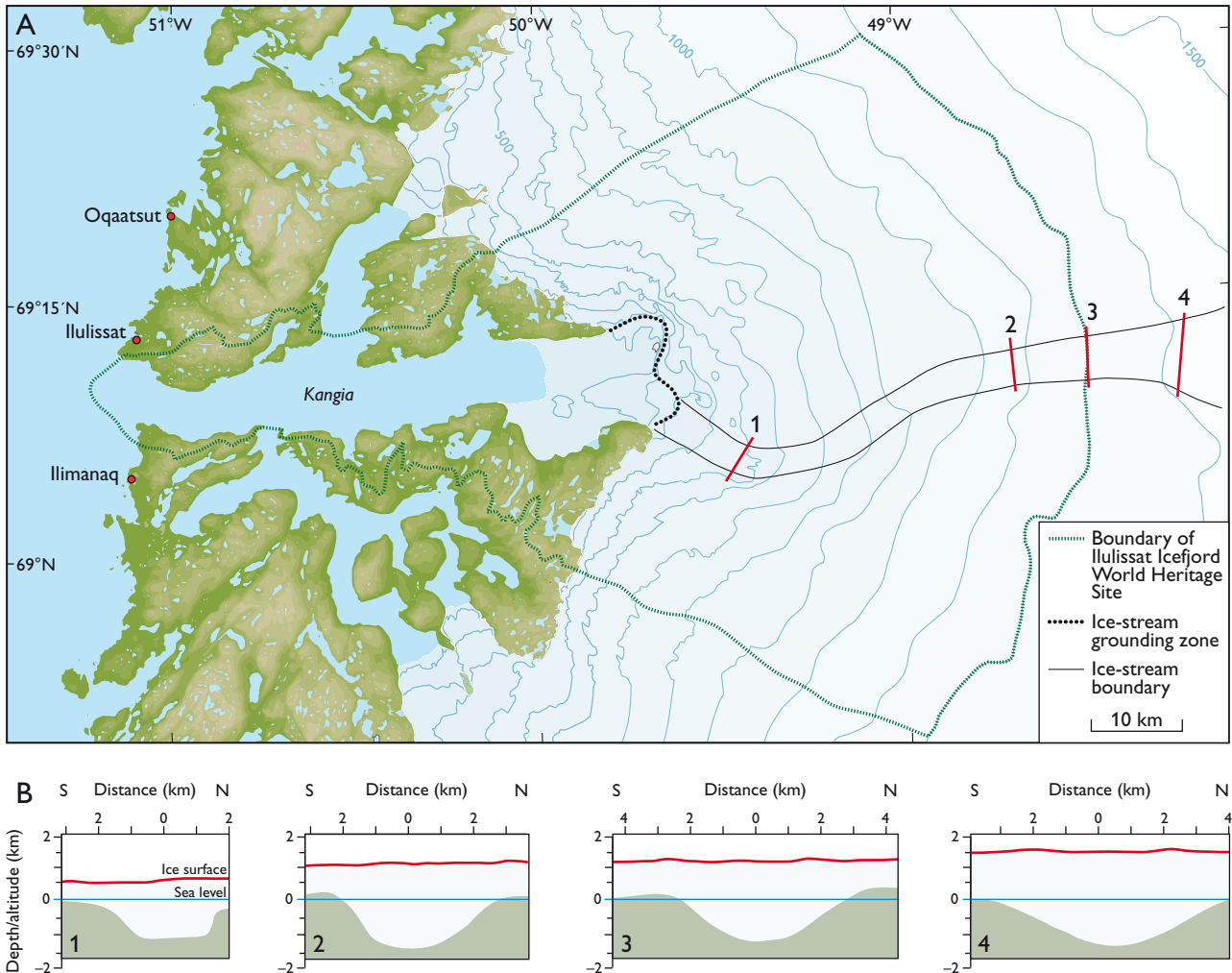


Fig. 35. A: Map of Kangia and the ablation zone of Jakobshavn Isbræ showing the positions of cross-sections 1–4 illustrated below. The ice-free areas have been mapped by GEUS, and the Inland Ice by H. Brecher (in: Fastook & Hughes 1994 and Fastook *et al.* 1995); supplementary data from the Inland Ice have been provided by S. Ekholm (National Survey and Cadastre, Copenhagen). Map compilation by Willy Weng and Annette T. Hindø (GEUS). The borders of the Jakobshavn ice stream (black lines) and the grounding zone *c.* 1985 (dotted black line) are from Echelmeyer *et al.* (1991, 1992). Ice-margin data from 1994; note that the present margin of Jakobshavn Isbræ, following the dramatic recent recession (Fig. 13), broadly coincides with the grounding line shown here. B: The four profiles of the ice stream (1 to 4) are based on seismic studies (Clarke & Echelmeyer 1996). N and S on the profiles indicate the north and south ends.

Present ice-margin surface

A topographic map of Greenland at a scale of 1:2 500 000 with 250 m contour intervals, including the entire ice sheet, was published by KMS [National Survey and Cadastre] in 1994, and the same topographic base was used in the Geological Map of Greenland at the same scale (Escher & Pulvertaft 1995). These maps show a low surface gradient from the highest central part of the ice sheet outwards

towards the ice margin, reflecting in some respects the low-land areas underlying the ice (Fig. 4). Detailed surface maps were produced in connection with the investigations along the EPF-EGIG line at Eqip Sermia (Holtzschcher & Bauer 1954), over the ice sheet margin at Paakitsoq (Thomsen *et al.* 1988), and around Jakobshavn Isbræ (Fastook & Hughes 1994; Fastook *et al.* 1995). The map around Jakobshavn Isbræ was used to delineate the ‘Ilulissat Ice-

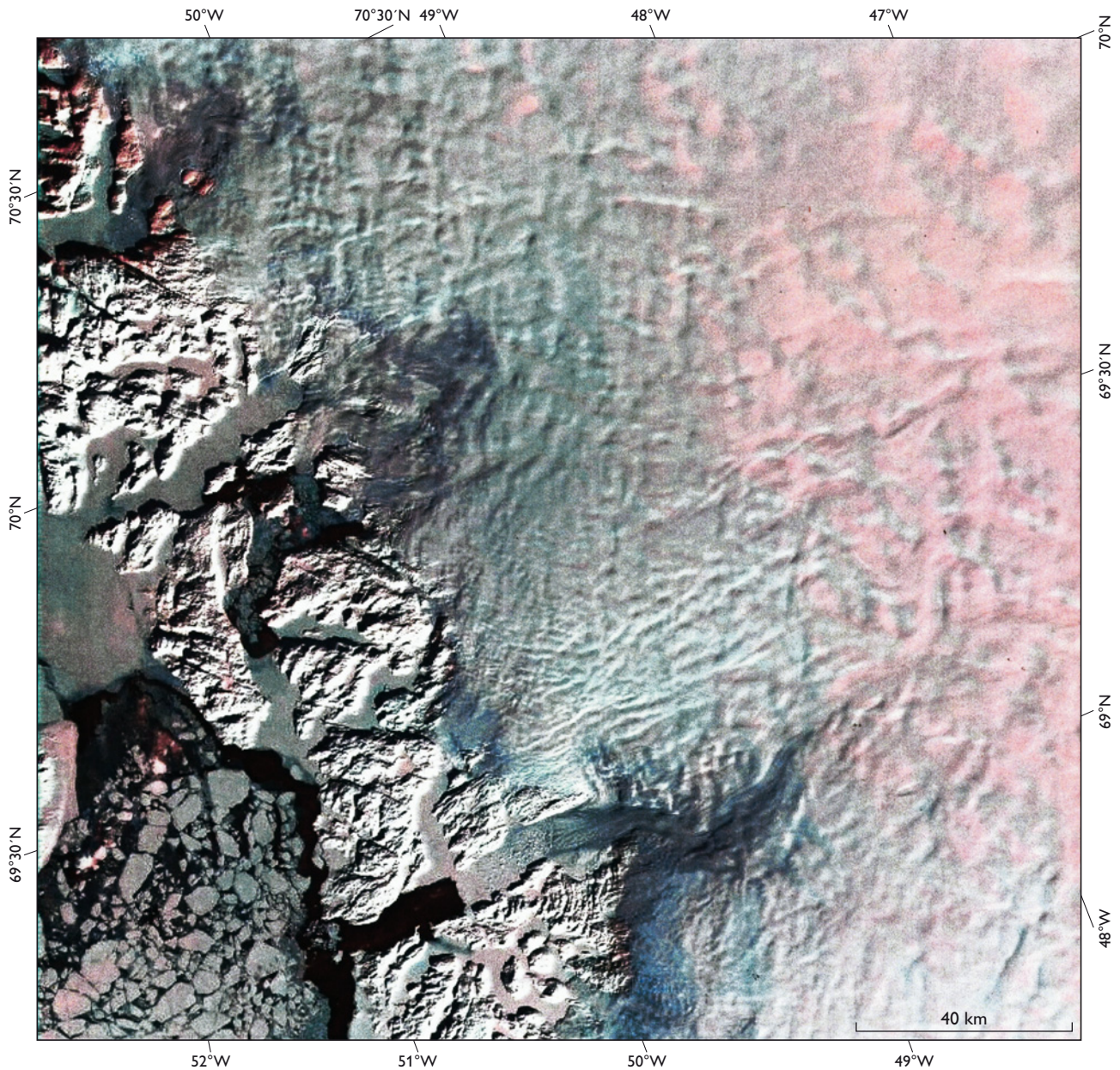


Fig. 36. False-colour satellite image (Landsat-7) of the interior of Disko Bugt and adjacent parts of the ice sheet, from 4 March 2002. The outer part of Kangia, and coastal areas north and south of the mouth of Kangia, are ice-free (black areas in lower part of the image). The morphology of the snow-covered, ice-free land continues under the surface of the ice margin, as can be seen from the surface features of the ice.

fjord' World Heritage proposal that was included on the World Heritage List in 2004 (Mikkelsen & Ingerslev 2002). This map is reproduced here (Fig. 35), and clearly shows the trend of Jakobshavn Isbræ.

In the Disko Bugt region, annual mass-balance field measurements have been carried out along the EPF-EGIG line (Holtzscherer & Bauer 1954; Bauer *et al.* 1968b, Ambach 1977), the GGU stake line (Fig. 15; Thomsen *et al.* 1988; Braithwaite *et al.* 1992) and a line along Jakobshavn

Isbræ (Figs 37, 38; Echelmeyer *et al.* 1992). The altitude of the equilibrium line, where the annual mass balance is 0, shows great annual variations, from *c.* 1000 m a.s.l. (extremely cold budget years) to more than 1200 m a.s.l. An in-depth discussion of climatic factors determining the variations in annual mass balance is given by Echelmeyer *et al.* (1992), who also record the albedo changes (increased ablation) due to inblown dust from the extensive trimline zone around Jakobshavn Isbræ. Calculated variations in

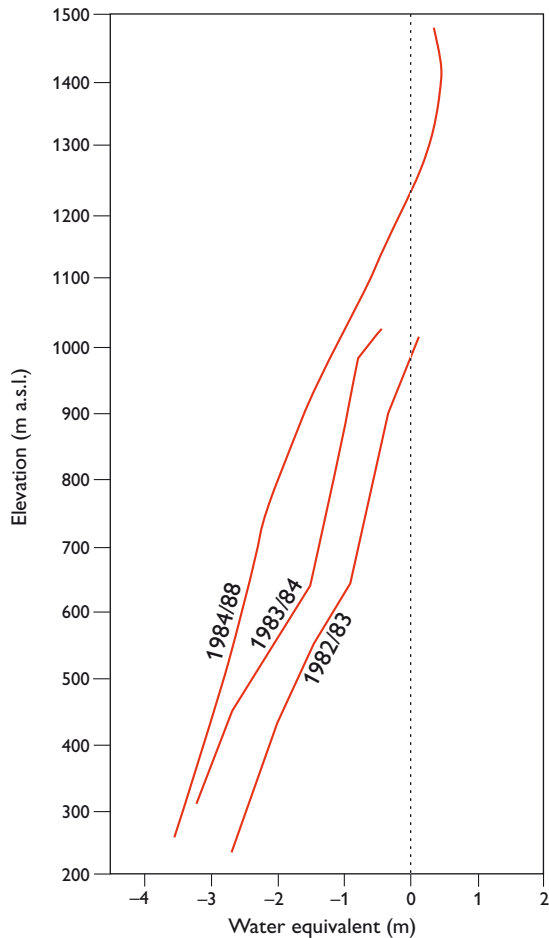


Fig. 37. Annual mass balance in relation to elevation along profiles in the Ilulissat area. 1982/83 and 1983/84 designate profiles along the GGU stake line (see Figs 15 and 38 for location; from Thomsen *et al.* 1988). 1984/88 shows measurements for this time period along Jakobshavn Isbræ (from Echelmeyer *et al.* 1992).

the annual ablation of the region for the period from 1961/1962 to 1989/1990 are reported by Braithwaite *et al.* (1992).

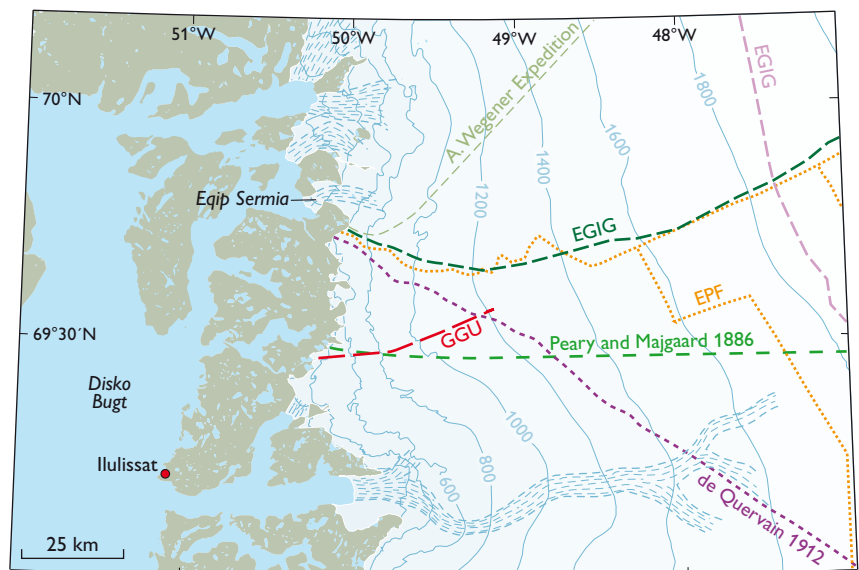
The trend of the annual mass balance along the GGU line, measured in 1982/83 and 1983/84 (Thomsen *et al.* 1988), is compared in Fig. 37 with the mass balance along Jakobshavn Isbræ, measured from 1984 to 1988 (Echelmeyer *et al.* 1992). The trends of the curves (the ablation gradient) are similar, but the difference between the equilibrium line altitudes (ELA) mainly illustrates the variations of the annual climatic conditions in the area, rather than variations due to the distance between the locations of the measurements (Fig. 38).

The ELA defines the upper limit of the ablation area, which is about 50 km wide. Above this, the accumulation zone extends to the top of the ice sheet around Summit. The decrease of intermittent summer-snow melt with increasing elevation is expressed by the division of the firn area into a region of superimposed ice over wet snow, a percolation facies and finally a dry snow facies. These facies were originally defined by Benson (1962), and have been modified by Williams *et al.* (1991) and Benson (1994, 1996).

The facies concept is applied in the current monitoring programme of the mass balance of the Inland Ice, which is based on data from satellites. The spectral variability of ice and snow surfaces is used to determine the different facies (Fausto *et al.* 2007).

The meltwater produced above the ELA during the summer is retained in the firn in increasing amounts with decreasing elevation. Refreezing of this meltwater leads to

Fig. 38. Map of the Jakobshavn Isbræ area, showing routes, stations and stake lines of the expeditions that have worked on the ice. EGIG, Expédition Glaciologique Internationale au Groenlande, mainly 1957–1960; EPF, Expéditions Polaires Francaises (Missions P.E. Victor 1948–1951); GGU, Geological Survey of Greenland, stake line 1982–1989. Modified from Wegener *et al.* (1930) and Weidick & Thomsen (1983).



increased temperatures in the firn. In the ablation zone, the superficial meltwater drains through crevasses and subglacial tunnel systems, where refreezing and closure of the meltwater conduits can take place. The complex system of drainage that can occur in glaciers has been described by Roethlisberger & Lang (1987). Descriptions of the drainage in the Paakitsoq area (site of the GGU line) are provided by Thomsen *et al.* (1988) and Zwally *et al.* (2002), and for Jakobshavn Isbræ by Echelmeyer *et al.* (1992).

Drainage and thermal conditions at the ice margin

The large ablation area of a continental ice sheet such as the Inland Ice is characterised by extensive englacial and subglacial drainage. Superficial meltwater drains into subglacial channels and conduits that are fed from crevasses and moulins on the ice surface. At the base of the ice, the water can drain in thin water films or in subglacial channels to the ice margin (Thomsen *et al.* 1988), and the water can emerge as upwelling plumes at the front of calving tidewater glaciers (outlets) such as Jakobshavn Isbræ (Echelmeyer *et al.* 1992). The contribution of basal melting from the Jakobshavn Isbræ drainage basin has been calculated to be *c.* 20% of the total loss by ablation (Echelmeyer *et al.* 1992). The complexity of the drainage is influenced by refreezing or by internal as well as external heating due to the movement of the glacier.

Glacier-dynamic modelling has been applied to the so-called ‘quiet’ sector, located between Jakobshavn Isbræ in the south and the ice streams draining into Torsukattak icefjord in the north. The ice margin in this sector is bordered by land areas or lakes. Modelling of the bottom conditions by Radok *et al.* (1982) suggested that the basal ice reaches its pressure melting point 250–300 km from the ice margin. Based on a model for calculating the response of the marginal sector of the Inland Ice to mass-balance changes, the ice margin at Paakitsoq (Fig. 28) could be divided into three zones: (1) An inner zone more than 292 km from the ice margin where ice movement is dominated by internal deformation. (2) A zone between 292 and 18 km from the margin where the ice is at the pressure-melting point, which leads to a significant degree of bottom sliding. (3) An outer zone, at 18–0 km from the ice margin, that is characterised by extensive sliding, high hydraulic pressure at the bottom of the ice, and conditions that allow for formation of cavities (Reeh 1983; Thomsen *et al.* 1988).

Ice-temperature measurements have been made at a few localities along both the EPF-EGIG and GGU lines. At Camp VI on the EPF-EGIG line, drilling from the surface

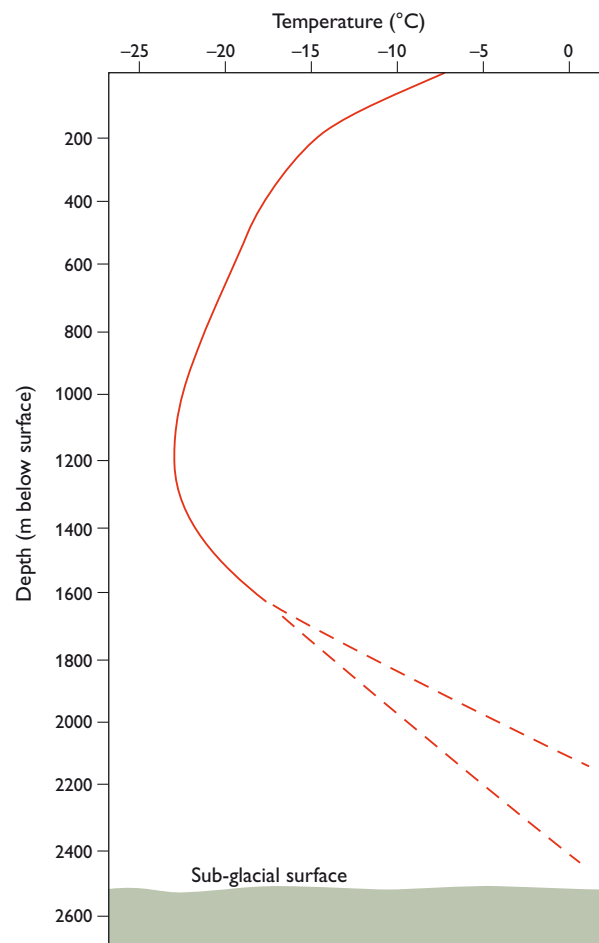


Fig. 39. Temperature profile through Jakobshavn Isbræ near its centre line, showing temperature versus depth below the ice surface (from Iken *et al.* 1993). The profile is situated about 50 km from the glacier front of 1988/89, at an elevation of 1000 m a.s.l., with the bottom of the ice at about 1500 m b.s.l. The dashed lines show possible limits for temperatures below the measured values. More recent modelling indicates that the thickness of the basal temperate ice (ice at the pressure-melting point) is around 230 m in the ice stream, and *c.* 30 m in the adjacent ice sheet (Funk *et al.* 1994).

at 1598 m a.s.l. extended only to a depth of 125 m; a temperature profile from this hole shows a decrease in temperature from *c.* -12.5°C to -16°C (Heuberger 1954; Robin 1983). The ice thickness here was about 1000 m. On the GGU line, temperatures were measured in a hole drilled at *c.* 500 m a.s.l. Temperatures were slightly negative throughout the ice hole, decreasing to -2.1°C in the upper 50 m (Thomsen *et al.* 1991).

For Jakobshavn Isbræ, it has been calculated that 2–3 km^3/year of meltwater is generated by deformational heating, although the meltwater does not seem to influence the movement of the ice stream (Echelmeyer & Harrison 1990; Echelmeyer *et al.* 1992).

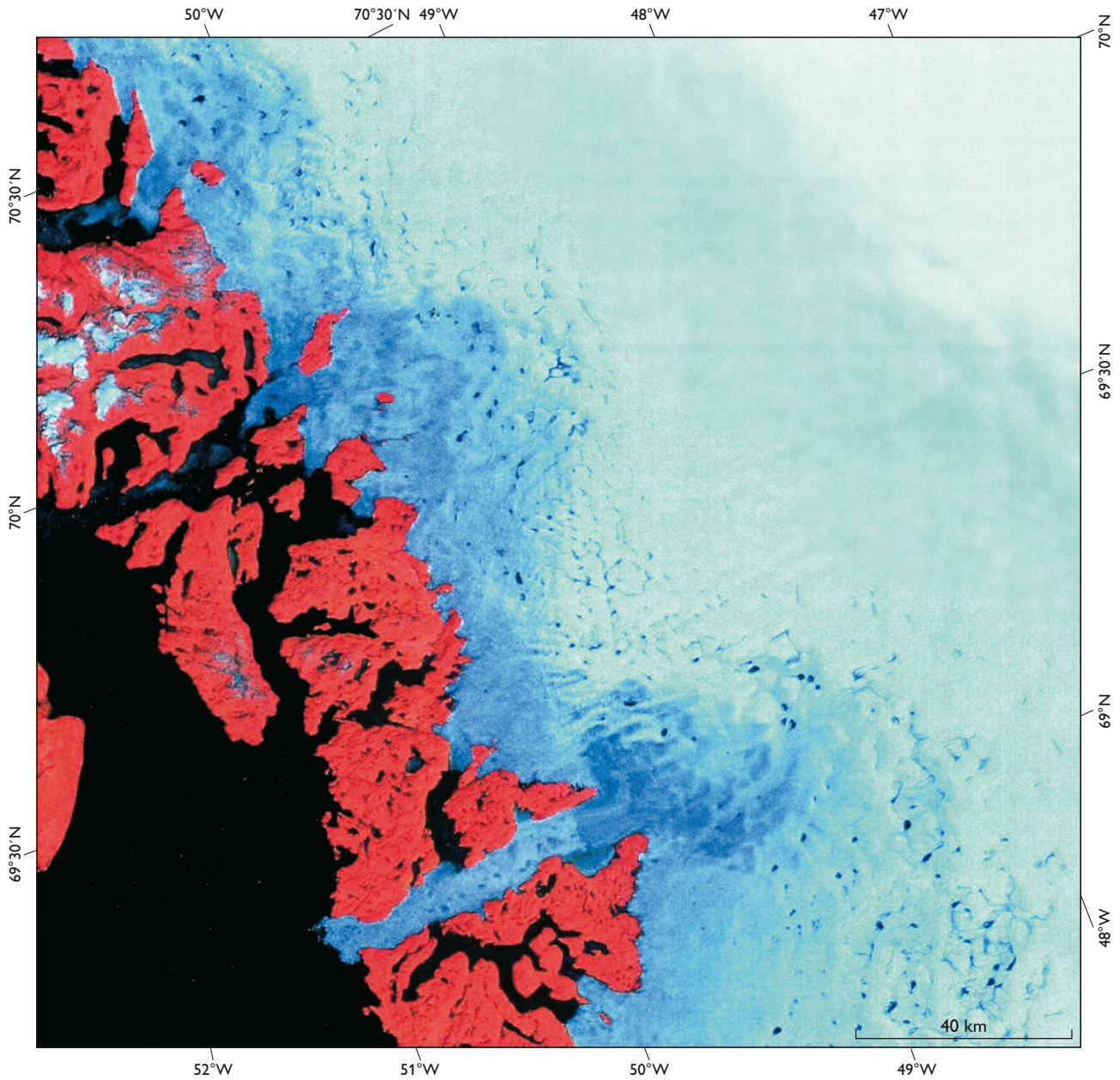


Fig. 40. False-colour satellite image (Landsat-7) of the interior of Disko Bugt and adjacent parts of the ice sheet, from 7 July 2001. The scene shows the zone with numerous lakes on the ice near the equilibrium line.

A temperature profile of Jakobshavn Isbræ was measured in 1988/1989 at a locality situated about 50 km from the glacier front at 1020 m a.s.l. The site is located at the centre line of the ice stream, where the surface ice moves at about 1 km/year, and the bed is situated at about 1500 m below sea level (Iken *et al.* 1993, Clarke & Echelmeyer 1996). The temperature profile (Fig. 39) shows a temperature minimum of -22°C c. 1200 m below the surface; this implies a thick, relatively warm and low viscosity bottom layer which is thought to facilitate the fast movement of

the ice stream. It has been suggested that this basal layer may contain Wisconsinan ice (Iken *et al.* 1993). This work was followed up by investigations of flow and temperature conditions at the transition between Jakobshavn Isbræ and the 'quiet' ice margin (Lüthi *et al.* 2002).

The surface features of the 'quiet' sectors of ice comprise lakes, rivers, crevasse formations and moulins, which have been mapped in detail and described along the lower parts of the EGIG line by Bauer *et al.* (1968a). Detailed maps and descriptions around the GGU line near Paakitsoq are

given by Thomsen *et al.* (1988). Thorough descriptions of Jakobshavn Isbræ and its surroundings are provided by Echelmeyer *et al.* (1992) and Fastook *et al.* (1995). The surface of the Inland Ice margin in the Disko Bugt region is characterised by numerous lakes up to an altitude of about 1400 m (Fig. 40). Above this altitude, corresponding approximately to the beginning of the wet snow facies of the accumulation area, lakes are present but are ice and snow covered and only faintly visible. The occurrence of numerous lakes extends down to about 1000 m a.s.l., corresponding to the zone of superimposed ice and the upper part of the ablation area. Lower down in the ablation area there are fewer lakes, due to the decreasing thickness of the ice cover that leads to increased crevassing during ice movement.

Where the ice margin is bordered by land, marginal lakes are common. Due to the bordering hilly upland, the occurrence and size of the lakes vary with changes in the position of the ice margin, just as the proglacial drainage shows variations. Changes in drainage patterns at Paakitsoq, between 1953 and 1959, can be documented by aerial photographs from the 1950s, 1960s and 1985 (Thomsen 1983).

Tinnilik, situated *c.* 40 km south of Jakobshavn Isbræ, is an ice-dammed lake at the ice margin that shows periodic drainage. The lake covers an area of 43 km² at its maximum extent and 20 km² at its minimum (Thomsen 1984). When the periodic drainage was first described in 1913 (Koch & Wegener 1930), it was stated that the filling/drainage cycle of the ice-dammed lake was about 10 years based on information from local people. This cycle seems to be more-or-less permanent, probably due to the nature of the damming glacier (Saqqarliup Sermia) which, in contrast to Jakobshavn Isbræ, has been almost stable over the last 150 years (Weidick 1994a, b). Braithwaite & Thomsen (1984) determined that during each drainage event nearly 2 km³ of water is released along Saqqarliup Sermia into the southern branch of the Tasiusaq fjord complex, and from there onwards to the south side of Kangia. Braithwaite & Thomsen (1984) also recorded the years of drainage for the period 1942–1983; the most recent drainage took place in 2003 (F. Nielsen, personal communication 2004).

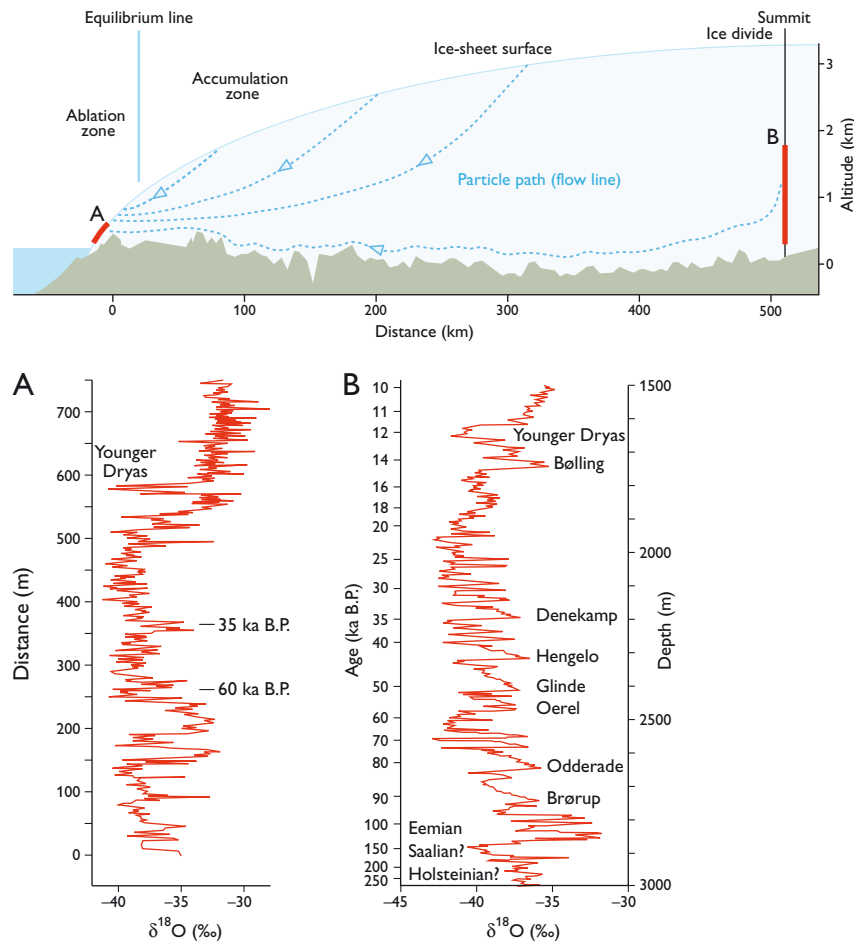


Fig. 41. Cross-section of the Inland Ice, showing schematic flow lines within the ice and the locations of (A) an oxygen isotope record near the margin and (B) the deep ice core at the summit of the ice sheet. A: Profile of the outer ice margin at Paakitsoq. The record covers major parts of the last ice age (Wisconsinan). From Reeh *et al.* (1991). B: A section of the ice core at Summit (Fig. 1) revealing successively older layers of snow and ice. The record covers about 250 000 years, and includes the Eemian and perhaps Holsteinian interglacials (Dansgaard *et al.* 1993).

The age of the ice margin has been investigated in detail along three profiles near Paakitsoq (Fig. 2). At certain localities along the ice margin, the ice stratification demonstrated from deep drill holes in the interior of the ice sheet can be preserved (Fig. 41; Reeh *et al.* 2002, Petrenko *et al.* 2006). Such localities, as at Paakitsoq, are characterised by a fairly smooth subsurface within so-called quiet marginal areas distant from larger ice streams. The profiles so far investigated at Paakitsoq cover a span of time from possibly 150 000 years B.P. to the present. These and similar investigations at the ice margin provide easy access to ice-age ice, a cheaply acquired supplement to the very expensive deep ice-core records, which can provide important information about past climates and the dynamics of the ice sheet.

Movement of the ice margin

The surface movement of the ice margin in Disko Bugt has been determined in the ‘quiet’ land-based area along the EGIG line north of Jakobshavn Isbræ, and in Fig. 42 it is compared with the horizontal movement of Jakobshavn Isbræ. The main differences are found in the areas around and below the equilibrium line. The thickness of the land-

based ice is up to 700–800 m below the ablation zone along the EGIG profile (Holtzschcherer & Bauer 1954; Bauer *et al.* 1968b). Along this profile, the horizontal surface movement increases westwards to a maximum of 100–200 m/year just above the equilibrium line, and then decreases towards the ice margin.

At the Swiss Station (for location, see Fig. 15), measurements of movements were made from 1996 to 1999 (Zwally *et al.* 2002). The station is situated near the equilibrium line at 1175 m a.s.l., at a site where the ice is 1220 m thick. The measurements show increasing ice velocities with increasing surface melt, indicating that bottom sliding is enhanced by rapid migration of meltwater to the bottom of the ice. This provides a mechanism for rapid, large-scale dynamic response of ice sheets to climate warming, for example during the transition from glacial to initial interglacial conditions.

With respect to calving tidewater outlets in fjords as well as calving outlets in proglacial lakes, it is generally known that the rate of movement increases towards the front reaching up to a few km/year. However, little information is available for calving outlets in the Disko Bugt region apart from at Jakobshavn Isbræ.

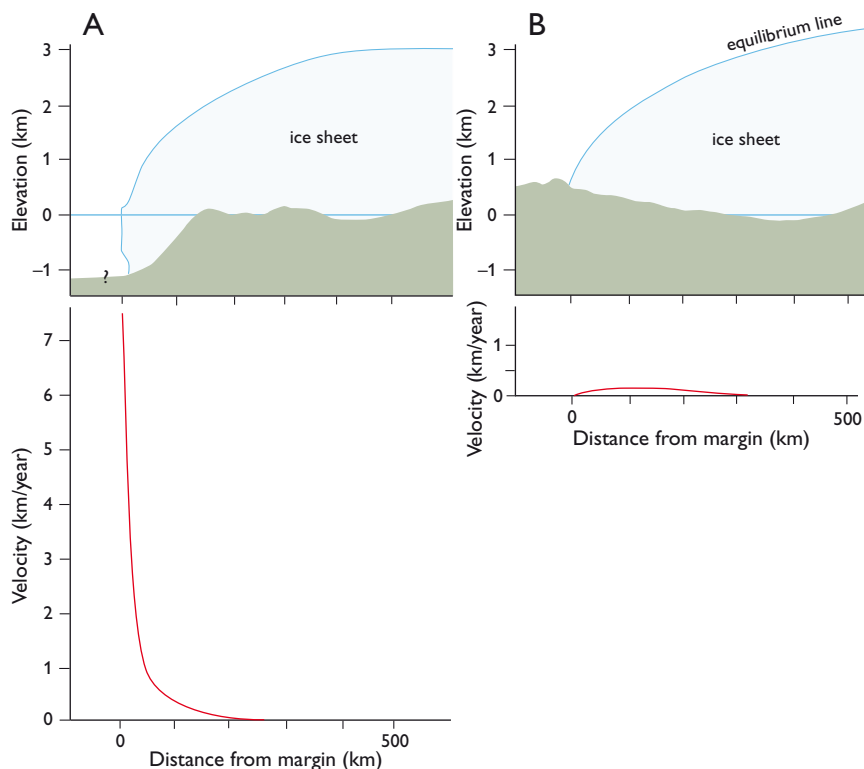


Fig. 42. **A:** Diagram showing a cross-section through Jakobshavn Isbræ together with a plot of the horizontal surface movement of the ice stream. Compiled from Fastook *et al.* (1995) and Joughin *et al.* (2004). **B:** Cross-section of the ice sheet along the EGIG line (Fig. 38) in a ‘quiet’ marginal area, about 60 km north of Jakobshavn Isbræ, in association with a plot of the horizontal surface movement of this part of the ice sheet, which has a maximum velocity of 200 m/year *c.* 100 km from the ice margin (from Bauer *et al.* 1968b).

Table 5. Velocity measurements of glaciers in the Disko Bugt region before 1950

Distance from flank of glacier (km)	Velocity		Comments	
	m/24 h	km/y		
Alanggorliup Sermia, July 1875 (Helland 1876)				
<0.5		<0.2?	Velocity measurement uncertain	
Jakobshavn Isbræ, 7–9 July 1875 (Helland 1876)				
<i>From south side</i>	0.40	14.7	5.4	Measurements close to front
	0.42	15.4	5.6	Frontal height scarcely over 40 m a.s.l.
	0.45	15.2	5.5	Width of the fjord estimated to 4.5 km (on present maps c. 7 km)
	0.45	15.2	5.5	
	1.05	19.8	7.2	
	1.06	19.5	7.1	
Jakobshavn Isbræ, 22 March – 24 April 1880 (Hammer 1883)				
<i>From south side</i>	0.28	5.1	1.9	Measured c. 5 km behind glacier front
	0.55	7.5	2.7	Maximum velocity in the central part of the glacier estimated to at least 16 m/24 h. Height of glacier front c. 63 m a.s.l.
	0.62	9.2	3.4	Width of fjord c. 7 km
	0.88	12.5	4.6	
	0.87	12.3	4.5	
Jakobshavn Isbræ, July 1902 (Engell 1904)				
<i>From south side</i>	1.29	15.0	5.5	Velocity measured near the front
	1.30	14.2	5.2	Width of fjord c. 7 km
	1.87	19.8	7.2	
	1.84	19.8	7.2	
	4.26	22.8	8.3	
Jakobshavn Isbræ, 27–28 September 1929 (Sorge in Wegener <i>et al.</i> 1930)				
<i>From south side</i>	2.1–3.7	18–21	6.6–7.6	Preliminary calculations based on four points. Glacier c. 6 km wide, presumably measured near front
6.6–7.7				
Sermeq Avannarleq in Torsukattaq fjord, 24–25 July 1875 (Helland 1876)				
<i>From north side</i>	0.21	3.8	1.4	Presumably measured near front
	0.37	5.7	2.1	Recorded frontal height 15 m a.s.l.
	1.93	8.8	3.2	Glacier c. 9 km wide
	4.07	10.1	3.7	
	4.94	10.2	3.7	
	4.97	9.4	3.4	
Sermeq Avannarleq in Torsukattaq fjord (Steenstrup 1883a)				
<i>From north side,</i>	2.70	7.8	2.8	Measured near glacier front
<i>5–7 May 1879</i>	2.70	6.3	2.3	Width of glacier estimated to c. 8 km
<i>From north side,</i>	3.01	5.0	1.8	
<i>21–22 May 1880</i>	3.01	7.8	2.8	
	2.17	5.0	1.8	
	1.51	5.0	1.8	

Jakobshavn Isbræ is an extreme example, where the depth of the glacier is controlled by a subglacial trough reaching 80–100 km inland under the ice and with depths that reach down to 1.5 km below sea level in the outer parts (Clarke & Echelmeyer 1996). The trough can be envisaged as a continuation of the proglacial icefjord (Kangia), which is believed to be around 1000 m deep. Such a depth is extreme for the fjords in the area, where depths of 400–800 m are more usual (Weidick *et al.* 1974b). Fjord depths at the present fronts are not known, and can only be estimated from soundings made at some distance from the active fronts.

The empirical relationship between calving rate and water depth at the glacier front (Pelto & Warren 1991) is therefore difficult to establish, although an attempt was made by Long & Roberts (2003) for the deglaciation of Disko Bugt. Measurements of movement and thickness of all the tidewater glaciers in Disko Bugt and the Umannaaq fjord complex were undertaken in 1957 and 1964 by the EGIG expeditions (Bauer *et al.* 1968a; Carbone & Bauer 1968), based on photogrammetric analyses of aerial photographs, and show a positive correlation between mean rate of movement and mean thickness of the glacier front.

With respect to the high movement rate of many of the calving glaciers in the Disko Bugt region, and variations of the velocity over time, a brief historical review is given below. The field conditions for the earliest measurements introduce an element of uncertainty, but do provide an order of magnitude of the possible variations in velocity over long time-spans. All the measurements given are related to the frontal areas, and for nearly all the glacier fronts the variations in position are within 2 km for the period since the end of the 19th century (Fig. 43; Weidick 1994a, b). The measurements of velocity at the front of most outlets were thus conducted at nearly the same location. The only exception is Jakobshavn Isbræ, which receded 26 km between *c.* 1850 and *c.* 1950 and has experienced continued recession since 2002/2003; in contrast to other outlets, the measurements were here undertaken from widely different positions of the glacier front during the recession.

Most early velocity measurements were carried out over short time intervals, and consequently where a low rate of movement was recorded, possible errors may be large, a fact that is stressed in some of the old descriptions (Helland 1876). In Table 5, older data are only given for the faster moving glaciers. Some original sources gave measurements in Danish feet (1 Danish foot = 0.31385 m), here converted to metres or kilometres. Rates of movement are given as m/24 h or km/year (Tables 1, 5). The latter is used for comparison with modern data and neglects possible variations in the course of the year.

Notes on individual outlets

The glacial histories of individual glacier outlets in the Disko Bugt region (Figs 2, 43) are summarised below.

Nordenskiöld Gletscher (Akuliarutsip Sermersua); 68°20'N, 50°51'W. This glacier does not drain into Disko Bugt, but its glacial history is closely related to the bay. The visits of the Nordenskiöld expeditions to the glacier in 1870 and 1883 gave rise to detailed descriptions of this glacier (Nordenskiöld 1885, 1886), but the velocity of the outlet was apparently first measured in July 1957 (Bauer *et al.* 1968a). The average velocity was found to be 3 m/24 h with a small production of calf ice (Fig. 5). From the descriptions, it appears that the build-up of a frontal moraine and proglacial delta hinders the production of calf ice. In the period 2000–2005, a slight increase in velocity was reported by Rignot & Kanagaratnam (2006).

Saqqarliup Sermia (68°54'N, 50°18'W) and Alanngorliup Sermia (68°55'N, 50°12'W). Alanngorliup Sermia was

visited in 1875 by Helland (1876), who found that the front had a height of *c.* 10 m. The velocity of the glacier was given as below 0.5 m/24 h, probably recorded in the lower reaches of the glacier although the location was not stated. Higher up in the glacier, where there is a tributary to Saqqarliup Sermia, the rate of movement is given as a maximum of 0.4 m/24 h. For both outlets, the mean frontal velocity in 1957 was measured at 0.9 m/24 h (Carbonnell & Bauer 1968). A slight velocity increase has also been reported for this outlet by Rignot & Kanagaratnam (2006).

Jakobshavn Isbræ (Sermeq Kujalleq; 69°11'N, 49°48'W). The first velocity measurements in 1875 were made along the southern side of the glacier. In the fast flowing, central part of the glacier, *c.* 1 km from the margin, a velocity of 19.8 m/24 h was measured (Table 5; Helland 1876). Helland also measured the movement of the glacier adjacent to the margin, where he found a velocity of not more than 0.02 m/24 h.

Subsequent measurements in 1880 recorded slightly lower velocities (Table 5; Hammer 1883; estimated maximum velocity >16 m/24 hours). The measurements appear to have been made *c.* 5 km behind the front, judging from a sketch map of measured points in Hammer's report. If the up-stream velocity decrease was similar to later values (Carbonnell & Bauer 1968), the frontal velocity may well have been of the same magnitude as given by Helland. However, the velocity does not always decrease immediately behind the glacier front; Joughin *et al.* (2004) reported that the marked decrease in speed of Jakobshavn Isbræ in the 1990s first occurred *c.* 14 km behind the front.

In 1902, the glacier was visited by Engell, who recorded a similar high velocity to that given by Helland for the central part of the glacier (*c.* 23 m/24 h; Engell 1904). Comparable values of between 18 m/24 h and 21 m/24 h were measured by Sorge in 1929 (Wegener 1930; Wegener *et al.* 1930).

Measurements in July 1958 gave a mean velocity of 13.1 m/24 h (Bauer *et al.* 1968a), whereas investigations in June 1964 gave a mean velocity of 19.1 m/24 h (Carbonnell & Bauer 1968). A very similar figure to the 'frontal velocities' of the central zone of the glacier given by Pelto *et al.* (1989): 21.1 (1964), 20.4 (1976), 21.0 (1978) 20.6 (1985) and 20.3 (1986) m/24 h. A decrease of velocity immediately behind the front was documented.

In a comment to the surprisingly large difference between the velocities in 1958 and 1964, it was pointed out that similar large changes were found at Rink Isbræ (Fig 1.), and that more measurements are needed to explain the difference (Carbonnell & Bauer 1968, p. 77). Subsequent measurements of Jakobshavn Isbræ show velocity variations

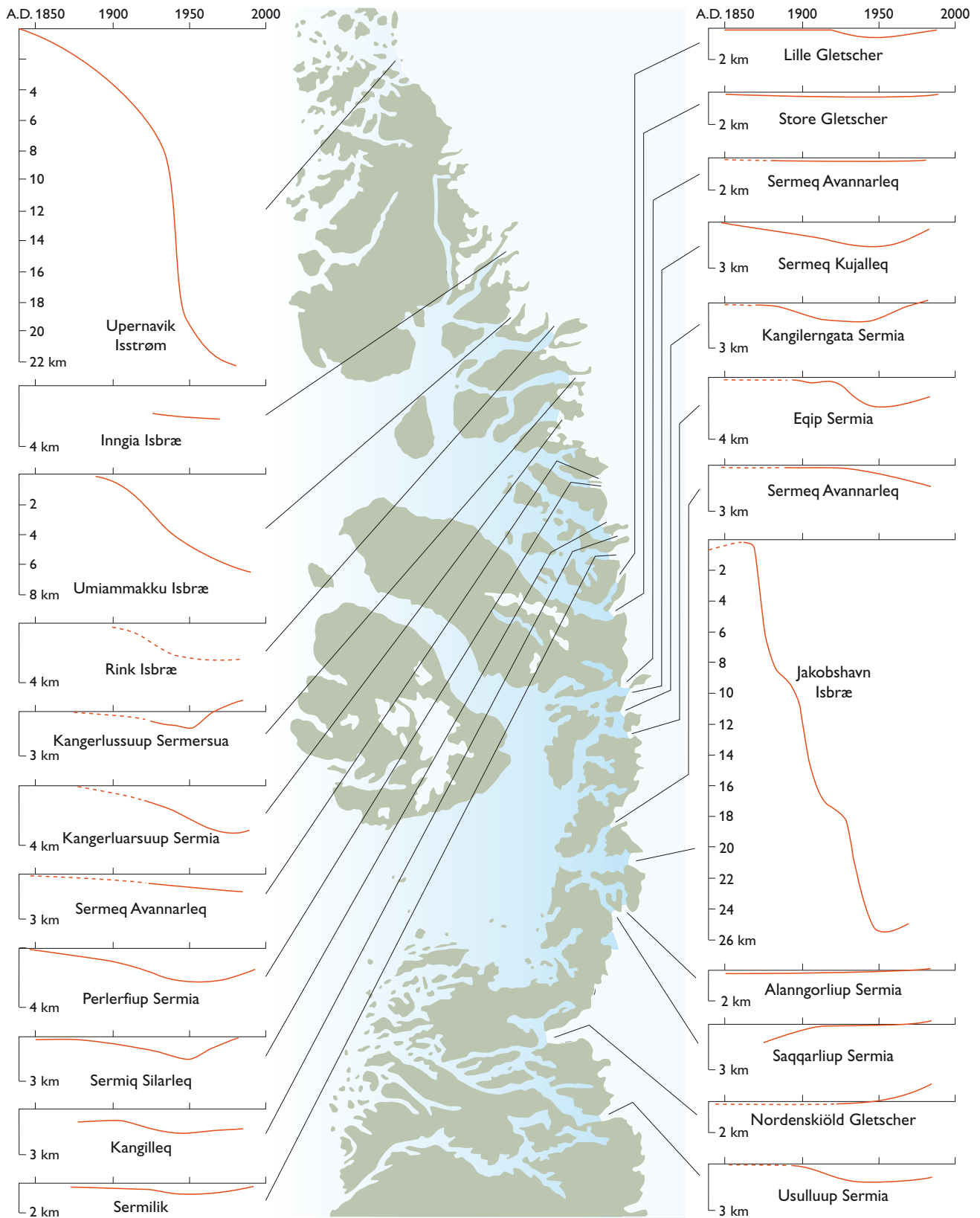


Fig. 43. Tentative reconstructions of fluctuations in the frontal positions of calving glaciers in central West Greenland from 1850 to 1985. Upernavik Isstrøm shows a recession of *c.* 23 km and Jakobshavn Isbræ shows a recession of *c.* 26 km (red lines), whereas the other glaciers show smaller fluctuations. From Weidick (1994b).

that are apparently related to the thickness of the ice margin. However, information is needed on the velocity in the time period between the measurements in 1929 and in 1948. It is also possible that the low velocity in 1958 could have been connected to a thickening of the glacier from *c.* 1950–2000, which was a period with a stable front. Studies of aerial photographs from the 1940s may provide data on thickness changes in the marginal parts of the ice sheet.

It is noteworthy that the maximum velocity, in spite of all the possible errors that may have affected the older measurements, has maintained a value of *c.* 5–9 km/year for more than a century. This is particularly relevant when considering the reasons for the marked velocity increase of the glacier after about 2000. Thus a 95% velocity increase took place between 1996 and 2005 according to Rignot & Kanagaratnam (2006). More details are provided below and in Fig. 44. In 2003, the velocity was 12.6 km/year and the discharge was *c.* 50 km³ ice/year (Joughin *et al.* 2004).

Sermeq Avannarleq in Kangia (69°2′N, 50°18′W). The velocity of this glacier was first measured during the EGIG expeditions, who recorded an average velocity of *c.* 1 m/24 h (Carbannel & Bauer 1968).

Eqip Sermia (69°48′N, 50°13′W). Velocity measurements of this glacier were first made from 31 July to 7 August 1912 by de Quervain, who recorded a maximum velocity at the front of 1.5–2.4 m/24 h (de Quervain 1925). Subsequent more detailed investigations on velocity and frontal fluctuations were made by Bauer during the EPF expeditions in 1948–1949 (Bauer 1955). A mean velocity of 3 m/24 h was recorded, and no difference was noted between September 1948 and June 1949.

The EGIG measurements from 12 to 17 July 1957 gave a mean velocity value of 3.1 m/24 h (Bauer *et al.* 1968a), whereas later measurements, from 5 to 9 July 1959 (Bauer 1968) and 29 June to 12 July 1964 (Carbannel & Bauer 1968) both gave values of *c.* 2 m/24 h. The differences are related to insufficient measuring points for the old data (Bauer 1968, p. 10). Detailed records of the frontal fluctuations of the glacier in the 20th century are provided by Bauer (1955) and Nielsen *et al.* (2000), who indicate advances around 1920, and during the 1990s. Between 2000 and 2005, the velocity of Eqip Sermia accelerated by 30% (Rignot & Kanagaratnam 2006).

Kangilerngata Sermia (69°55′N, 50°17′W). The movement of this glacier was first measured from 7 to 17 July 1957, when the average velocity was given as 2.3 m/24 h (Bauer *et al.* 1968a). Subsequent measurements from 9 to 22 June 1964 gave 3.3 m/24 h (Carbannel & Bauer 1968).

Between 2000 and 2005, the velocity of Kangilerngata Sermia also increased by 30% (Rignot & Kanagaratnam 2006).

Sermeq Kujalleq in the Torsukattak icefjord (70°00′N, 50°19′W). Velocity measurements of this large glacier started with the measurements of Bauer *et al.* (1968a) from 12 to 17 July 1957, which indicated a mean velocity of 7.2 m/24 h. From 9 to 22 June 1964, a mean velocity of 9.7 m/24 h was recorded (Carbannel & Bauer 1968). Sermeq Kujalleq slowed down by 11% between 2000 and 2005 (Rignot & Kanagaratnam 2006).

Sermeq Avannarleq in the Torsukattak icefjord (70°04′N, 50°19′W). The velocity was first measured by Helland (1876) from 24 to 25 July 1875. The measurements were made from the northern side of the glacier, and extended 4 km into the central parts where a velocity of over 10 m/24 h was measured (Table 5). Subsequent measurements by Steenstrup (1883a) in May 1879 and May 1880, from nearly the same position, did not reach as far into the glacier, but a velocity of 8 m/24 h was recorded *c.* 3 km from the glacier margin. This can be compared with later measurements of the mean glacier velocity from 12 to 17 July 1957 of 6.4 m/24 h (Bauer *et al.* 1968a) and on 9 to 22 June 1964 of 5.2 m/24 h (Carbannel & Bauer 1968). The velocity profile of the glacier is irregular. Maximum velocities of 9.5 m/24 h was found *c.* 1.5 km from the glacier margin (Carbannel & Bauer 1968, fig. 38). Sermeq Avannarleq, as Sermeq Kujalleq, slowed down between 2000 and 2005 by 11% (Rignot & Kanagaratnam 2006).

Sermeq Kujalleq (Store Gletscher: 70°24′N, 50°32′W). Although situated somewhat farther to the north, north of Disko Bugt, this glacier ('Store Qarajaq Bræ' of Steenstrup 1883a and 'Grossen Qarajaq Eisstrom' of von Drygalski 1897) deserves mention because of its history of exploration. Its velocity was measured in August 1878. The maximum velocity of 12 m/24 h, measured *c.* 3 km from its north side by Steenstrup (1883a), is close to the mean velocity recorded by von Drygalski (1897) in 1893 of 12.9 m/24 h, of 11.6 m/24 h by Bauer *et al.* (1968a) for 1957, and by Carbannel & Bauer (1968) of 13.4 m/24 h for 1964. These authors noted a significant upstream decrease in velocity for this glacier.

Historical information on the measurements of the calving tidewater glaciers in and around Disko Bugt leaves the general impression of high-speed behaviour of the major outlets to the fjords in the region; the scattered investigations lack sufficient details for further conclusions. The

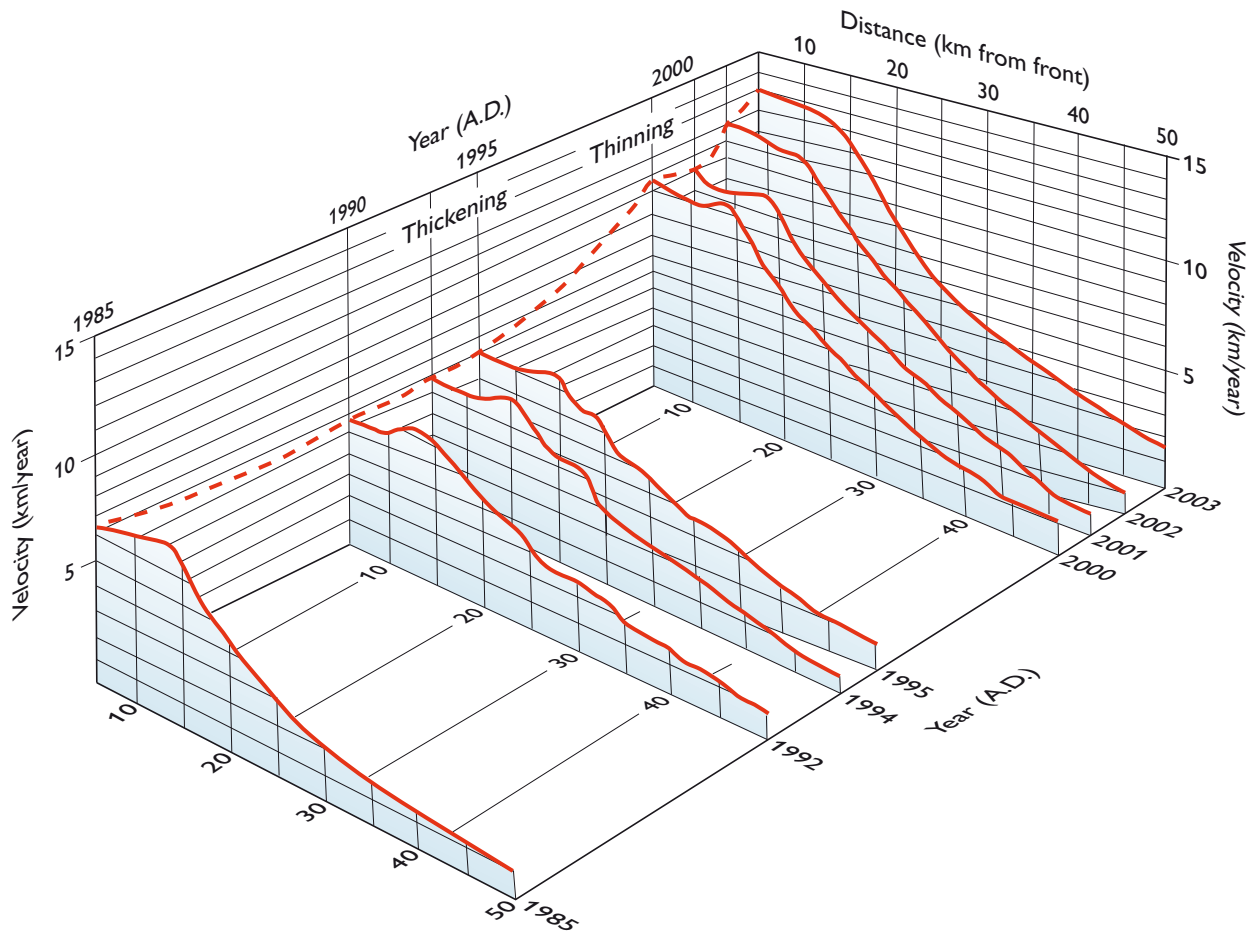


Fig. 44. Variations in the rate of movement of Jakobshavn Isbræ at distances *c.* 5–50 km behind the front. The marked increase in velocity that began in the late 1990s coincided with thinning of the front. Simplified from Joughin *et al.* (2004).

early measurements, especially those of von Drygalski's 'Grossen Qarajaq Eisstrom' and other outlets to Disko Bugt, reported the significant upstream decrease of the glacier velocities, although the investigations only covered the outermost 5–10 km of the glacier lobe.

Jakobshavn Isbræ is characterised by a high and rather constant flow rate at the glacier front, even though the position of the front has changed with time. Detailed coverage of the areal distribution of velocity is only known for Jakobshavn Isbræ, in the form of velocity contours (isotachytes of Ahlmann 1948). These cover an area of *c.* 80 × 80 km of the ice sheet upstream of the glacier front, with contours of 50 m/year, and record the situation at around 1985 (Fastook *et al.* 1995). The profile of Fig. 42 is derived from this source. A revision of the map for February 1992 and October 2000 (i.e. before the collapse of the front of Jakobshavn Isbræ in 2002/2003), is provided by Joughin *et al.* (2004), who describe the subsequent development illustrated by movement profiles from the front and reaching

50 km inland. Remote sensing data provide details of the frontal changes from 1985 to the present.

The velocity of Jakobshavn Isbræ has been more variable during the past few years than at any time since records began. A net thinning of the glacier of 200–300 m over the past 150 years, corresponding to an average thinning of 1.3–2.0 m/year, is apparent from the elevation of the trimline zone around the glacier. This thinning has taken place over a time period when the glacier has had a rather constant velocity of 5 to 9 km/year. Detailed records covering the past few decades indicate that the 1984 velocity of 6.7 km/year had decreased to 5.7 km/year by 1992, and that this lower velocity continued until 1997. The velocity then increased sharply, and for the years 2000, 2002 and 2003 the glacier attained velocities of respectively 9.4, 11.9 and 12.6 km/year (Joughin *et al.* 2004). A thickening of *c.* 1 m/year was related to the period of low speed (1991–1997), whereas the subsequent higher velocities were related to a thinning of around 6 m/year, Fig. 44. The

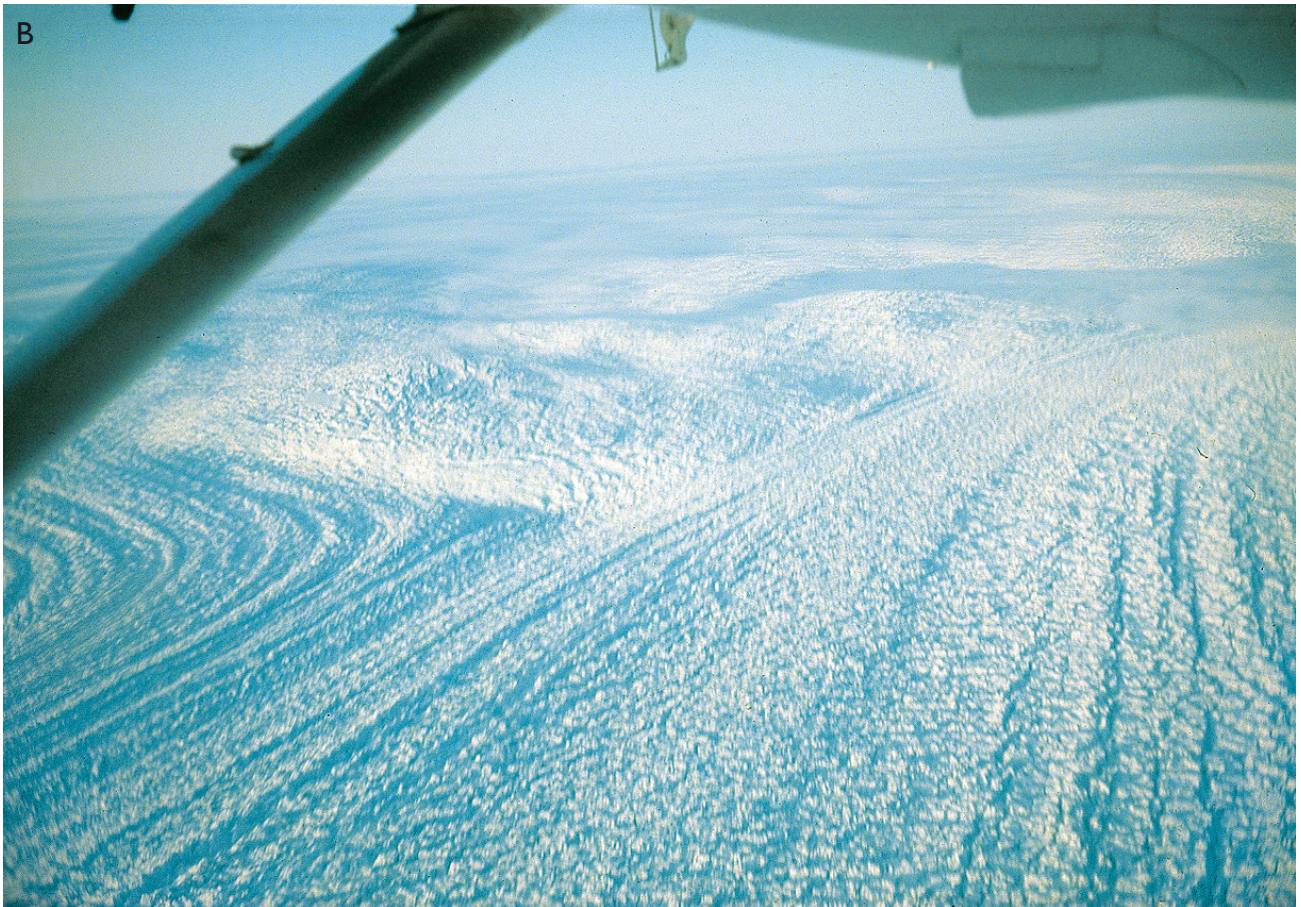
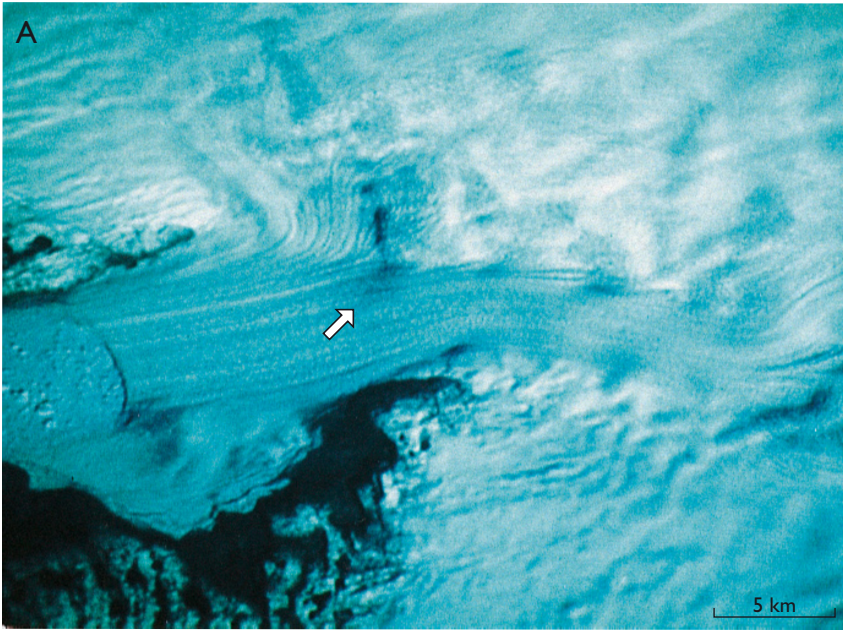


Fig. 45. Frontal area of Jakobshavn Isbræ. **A:** Landsat image of 27 September 1979. The glacier front and released icebergs can be seen near the left margin of the image. The ice stream is characterised by linear structures. A tributary from the north is separated from the main stream by a subglacial rumple. South of the front of Jakobshavn Isbræ is an area of stagnant ice (Tissarisoq, see also Figs 2, 25) that is separated from the ice stream by another rumple. The arrow indicates the view of the photograph below. **B:** Photograph of Jakobshavn Isbræ viewed from the WSW (see arrow on Fig. 45A) where the northern tributary joins the main ice stream. The bedrock topography under the ice is clearly reflected in the topography of the ice surface. Photograph by H.H. Thomsen, 1984.

sudden transition to rapid thinning that followed was at first confined to areas below *c.* 500 m a.s.l., but then spread inland and by about the year 2000 had reached up to 2000 m a.s.l. (Thomas *et al.* 2003).

The apparent quasi-stability of the front of Jakobshavn Isbræ from *c.* 1950 to 2000 has been related by Echelmeyer *et al.* (1991) to the presence of pinning points in the frontal areas of the floating front. However, little is known about the depths of the fjord below the floating outer parts of the glacier front that lie *c.* 22 km west of the lower seismic station of Clarke & Echelmeyer (1996), corresponding to profile 1 in Fig. 35. A possible threshold near the grounding zone (Figs 36, 45) may be viewed as a northern continuation of the curved bedrock lineaments on the south side of the glacier.

In contrast to the dramatic changes of the thickness and position of Jakobshavn Isbræ, the other calf-ice producing outlets to the north and south have shown only small changes in frontal positions during the past 150 years (Weidick 1994a, b). These small changes are perhaps linked to the shallower depths of these outlets demonstrated by the radar surveys of the Technical University of Denmark (Overgaard 1981) and apparent from low sun angle Landsat scenes.

For the areas around Jakobshavn Isbræ, the ice margin seems to have been nearly continuously receding over the

period from *c.* 1850 to 1950. Marginal zones farther to the north and south of Jakobshavn Isbræ, however, show a slight advance during the last decades of the 20th century (Fig. 43). The regional monitoring of glaciers in West Greenland on the basis of aerial photographs stopped with the last full coverage flown in 1985; subsequent regional coverage has been based on satellite imagery. Modelling of present response patterns of the ice sheet to climatic changes seems to match the present patterns of recession and readvance, and an important thickening of the south-western parts of the Inland Ice seems to have taken place (Huybrechts 1994). In contrast to the modelling, monitoring of outlet glaciers and marginal elevation changes based on repeated surveys by laser altimetry in 1993/1994 and 1998/1999 has revealed a significant thinning of the surroundings of Jakobshavn Isbræ, whereas the thickness of Jakobshavn Isbræ itself was constant or increasing (Abdalati *et al.* 2001). However, it may be misleading to compare the very generalised trends of the change of the ice margin positions based on scattered historical information with the detailed trends revealed during the past few decades (see e.g. Thomas *et al.* 2003 and Joughin *et al.* 2004 for Jakobshavn Isbræ).

With respect to the response of ice streams from the Inland Ice to climatic forcing, a holistic modelling approach has been presented by Hughes (2004), which lengthens and lowers the profiles of the Greenland ice streams.

Summary and outlook

The account of the onset and subsequent, repeated glaciations of Greenland, and the Disko Bugt region in particular, leaves more questions than answers. However, the detailed investigations of recent decades support the old idea that the preglacial, major fluvial drainage pattern of central Greenland was westwards towards the Disko Bugt region. The subsequent drainage of the Greenland ice sheet since its formation and during repeated glaciations was also predominantly westwards, and at the maximum extent of the Inland Ice, major ice streams in the present offshore region occupied and modified former river valleys and fjords.

The onset and extent of the early glaciations are still not clear. They must, however, have had characteristics similar to the glaciation of the present Antarctic or to high-arctic ice shelves, in that they showed a high sensitivity to climatic and eustatic sea-level changes and perhaps included ice streams of different character from those that drain the Inland Ice today. During the Illinoian, the glacial limit of the ice sheet may have been located at Store Hellefiskebanke and at the central part of Disko Banke. During the Wisconsinan, the ice margin may have extended only to the eastern part of Store Hellefiskebanke. On both occasions, the outer Egedesminde Dyb was probably occupied by a high-arctic type 'ice stream', but data to support this scenario are so far lacking.

A number of interglacial and interstadial deposits have been discovered along the outer coast of Disko island, and farther north at other localities along the extreme western parts along the outer coast. These have been referred to a number of marine events, but the extent of the ice sheet during and between the events is unknown. Modelling indicates, however, that during the last interglacial, the Eemian or Sangamonian, the Inland Ice was reduced to a degree where it was almost separated into a southern and northern ice sheet (Fig. 14).

The extent of the ice sheet during the last glacial maximum at 21 ka B.P. is still not established, neither from offshore stratigraphy nor from dates from marine sediment cores. It is likely that the pronounced warming at 14.7 ka B.P., combined with an initial rise of global sea level, caused a recession of the ice margin. During the cold interval of the Younger Dryas, the ice margin may have been located at the marked basalt escarpment that forms a submarine barrier, which would mean that the marine ice margin was situated at a depth of only 300–400 m.

In spite of the depth of this barrier, it must have had a blocking effect on ice movement, so that the large piedmont glacier that filled Disko Bugt during and immediately after the Younger Dryas was thinning rather than receding. In contrast, the few radiocarbon dates from Vaigat, north of Disko, provide minimum dates for the last deglaciation suggesting that the outlet here underwent gradual recession from 12.4 to 10.3 ka B.P. (Fig. 22).

The role of the inner part of Egedesminde Dyb, which has depths in excess of 1000 m, is unclear. The available descriptions and sea charts do not reveal a regular, U-formed conduit, but the western, steep basalt wall shows characteristics indicative of glacial plucking. We suggest that it was only the site of a major ice stream that helped to drain Disko Bugt during the initial thinning of the ice cover for a short time interval around 10.5–10 ka B.P.

It is certain that by *c.* 10 ka B.P. a major break-up of the Disko Bugt ice cover had taken place, and this probably occurred very rapidly, over a century or less. It is presumed that the change from the large, marine-based piedmont ice lobe to a land-based ice margin, with consequent changes in geometry, resulted in the prolonged halt or slowdown of ice-margin recession in the central parts of Disko Bugt that lasted from *c.* 9.9 to *c.* 7.9 ka B.P. As long as the front was situated close to Isfjeldsbanken near Ilulissat, the calving production of Jakobshavn Isbræ was probably much reduced. The cold event around 8.2 ka B.P. may also have contributed to the low recession rate.

Developments were different in the northern part of Disko Bugt, where the Torsukattak icefjord, the eastern continuation of the Vaigat strait, experienced a more continuous recession that brought the ice margin close to its present position before 8 ka B.P. However, the presence of moraines indicates minor deceleration or halts in the recession, perhaps at pinning points.

An iceberg bank similar to that near Ilulissat occurs at the junction between the Torsukattak icefjord and the Vaigat strait. The ice margin was presumably situated here from around 9.9 ka B.P., contemporaneous with the initial ice-margin position at the Jakobshavn iceberg bank. However, the ice margin was only situated at the mouth of Torsukattak icefjord for a short time, and also had a reduced frontal area. The relatively deep fjord may have favoured the continuous recession.

Around Jakobshavn Isbræ, the ice margin receded to its present location somewhat later, with 6.1 ka B.P. as a min-

imum age; recession of the ice-sheet margin continued to the east during the Holocene thermal maximum. Large undocumented fluctuations of Jakobshavn Isbræ may have occurred during the Neoglacial. We suggest that the high calf-ice production of Jakobshavn Isbræ began after the initial recession from Isfjeldsbanken, after *c.* 8 ka B.P. After the subsequent recession, at 5–4 ka B.P., the frontal position was at least 15–20 km east of the present location.

South of Disko Bugt, a relatively fast recession over the lowlands ended with the ice margin receding to its present position probably as early as 8 ka B.P. In this region, recession of the ice margin also continued beyond (east of) the present location. The extent of marine deposits in this region points to a partially marine ice margin during most of the recession. The depths of the fjords in this region, as far as is known, do not indicate the presence of troughs that could support major ice streams during the recession.

Ice streams are normally located in the ablation area of the ice sheet over Greenland, at sites where sufficiently deep troughs in the subsurface can act as conduits for the ice streams. With the gradual recession of the ice margin during the Wisconsinan to Holocene transition, periodic formation of ice streams might be expected during recession. The temporary role of such ice streams with respect to the total mass balance of the ice-sheet sector draining into the Disko Bugt region has still to be evaluated. Detailed mapping of the entire subsurface beneath the ice margin is essential in constructing the history of development of the ice sectors of the Disko Bugt region. If a prerequisite for the development of an ice stream is that it is located in a deep conduit in the marginal areas of the ice sheet, the life of Jakobshavn Isbræ with its present activity may be restricted to the Holocene period since *c.* 8 ka B.P.

The subsequent Little Ice Age readvance culminated in the area around Jakobshavn Isbræ with a major readvance in the middle and late 19th century. The extent of this readvance is unknown for the surrounding areas, with the exception of the Paakitsoq area. Historical information for the 20th century shows a marked recession only around Jakobshavn Isbræ, whereas the other ice-sheet sectors draining to Disko Bugt show a quasi-stability or even a tendency to advance over this period.

During the Holocene thermal maximum and the subsequent cooling, Jakobshavn Isbræ controlled much of the

ice drainage of central West Greenland, and its marked sensitivity to temperature changes must be linked to ablation and sliding mechanisms at the base of the ice-sheet margin. In general, Disko Bugt has played an important role for ice drainage of the central western slope of the ice sheet since the Wisconsinan. However, the role of individual factors such as ice streams, general ablation and related subglacial meltwater transfer to the bottom of the ice and marine versus land-based ice margin, still needs to be quantified. This can be achieved through better mass-balance investigations and dynamic modelling of the ice-margin change through time. Furthermore, the role of the individual troughs in Disko Bugt and offshore as conduits for the ice should be considered. The scattered positions of these troughs, and their relationship to temporary halts of the ice margin, point to a shift in the nature and position of ice streams during the recession of the ice margin since the last glacial maximum. Whether these offshore troughs had the same central role as the present ice stream of Jakobshavn Isbræ during the recession is still to be documented.

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Appendix 1

Index to Greenland place names

Names in quotation marks are informal names. Unless stated otherwise, the localities are in West Greenland.

Aasiaat (Egedesminde)	town	Innaarsuit	land area
Akuliarutsip Sermersua (Nordenskiöld Gletscher)	glacier	Isfeldsbanken	submarine threshold
Akulliit	island	Jakobshavn (Ilulissat)	town
Alanngorliup Sermia	glacier	Jakobshavn Isfjord (Kangia)	icefjord
Alluttoq (Arveprinsen Ejland)	island	Jakobshavn Isbræ (Sermeq Kujalleq)	glacier
Ammassalik (Tasiilaq)	town, East Greenland	Jøkelbugten	bay
Arfersiorfik	fjord	Kane Basin	large bay (basin)
Arveprinsen Ejland (Alluttoq)	island	Kangerluarsuk (Vaskebugt)	bay
Buksefjorden (Kangerluarsunnguaq)	fjord	Kangerluarsunnguaq (Buksefjorden)	fjord
Camp Century	former ice-sheet station	Kangerlussuaq (Søndre Strømfjord)	fjord/airport
Christianshåb (Qasigiannugit)	town	Kangerlussuaq	fjord, East Greenland
Claushavn (Ilimanaq)	town	Kangerlussuaq	branch of Uummannaq Fjord
Disko (Qeqertarsuaq)	island	Kangerlussuaq Gletscher	glacier, East Greenland
Disko Banke	shallow offshore area	Kangersooq (Nordfjord)	fjord
Disko Bugt (Qeqertarsuup Tunua)	bay	Kangersuneq	fjord
Disko Fjord (Kangerluk)	fjord	Kangia (Jakobshavn Isfjord)	icefjord
Dye 3	former ice-sheet station	Kangilerngata Sermia	glacier
Egedesminde (Aasiaat)	town	Kangilliup Sermia (Rink Isbræ)	glacier
Egedesminde Dyb	submarine trough	Kap Farvel (Nunap Isua)	cape
Eqalorutsit Killiit Sermiat	glacier	Laksebugt (Eqaluit)	bay
Eqaluit (Laksebugt)	bay	'Lerbugten'	bay
Eqip Sermia	glacier	Lersletten (Naternaq)	clay plain
Gade Gletscher	glacier	'Lersletten' (Narsarsuaq)	clay plain
Godhavn (Qeqertarsuaq)	town	Marraq	clay plain
Godthåb, (Nuuk)	capital of Greenland	Melville Bugt (Qimusseriarsuaq)	bay
Gunnbjørn Fjeld	mountain, East Greenland	Nanortalik	town
Harald Moltke Bræ (Ullip Sermia)	glacier	Narsap Sermia	glacier
Hareøen (Qeqertarsuatsiaq)	island	Narsarsuaq ('Lersletten') at Ilimanaq	clay plain
Helheimgletscher	glacier, East Greenland	Narsarsuaq	airport, South Greenland
Hellefisk-1	offshore well	Naterna (Lersletten)	clay plain
Ikerasaap Sullua (Qarajaq Isfjord)	icefjord	Nordenskiöld Gletscher (Akuliarutsip Sermersua)	glacier
Ilimanaq (Claushavn)	town	NorthGRIP	ice-core site
Ilulissat (Jakobshavn)	town	Nordfjord (Kangersooq) at Disko island	fjord
Ilulissat Icefjord	icefjord (world heritage site)	Nuuk	peninsula
Inland Ice	Greenland icecap	Nuuk (Godthåb)	capital of Greenland
		Nuussuaq	peninsula

O qaatsut (Rodebay)	settlement
Orpissooq	fjord
P aakitsoq	bay
Paamiut (Frederikshåb)	town
Pattorfik	coastal stretch
Peary Land	region, North Greenland
Pinguarsuit	rock knoll
Pituffik (Thule Air Base)	airbase
Q ajaa	ruin site
Qapiarfiit	land area
Qaqortoq (Julianehåb)	town
Qarajaq Isfjord (Ikerasaap Sullua)	icefjord
Qasigiannuguit (Christianshåb)	town
‘Qarsortoq’	coastal cliff
Qeqertarsuaq (Disko)	island
Qeqertarsuaq (Godhavn)	town
Qeqertarsuatsiaq (Hareøen)	island
Qeqertarsuatsiaq	island south of Aasiaat
Qeqertarsuup Tunua (Disko Bugt)	bay
Qilertinnuguit	mountain
R ink Isbræ (Kangilliup Sermia)	glacier
Rodebay (Oqaatsut)	settlement
‘ S andbugten’	bay
Sanddalen	valley
Saqqarliup Sermia	glacier
Serfarsuit	headland, Kangersuneq Fjord
Sermeq (Upernavik Isstrøm)	glacier
Sermeq Avannarleq in Kangia	glacier

Sermeq Avannarleq in Torsukattak	glacier
Sermeq Kujalleq (Jakobshavn Isbræ)	glacier
Sermeq Kujalleq in Torsukattak	glacier
Sermeq Kujalleq (Store Gletscher) in Qarajaq icefjord	glacier
Sermermiut	bay/ruin site
Sigguup Nunaa (Svartenhuk Halvø)	peninsula
Sikuiuitsoq near Ilulissat	fjord
Sisimiut (Holsteinsborg)	town
Søndre Strømfjord (Kangerlussuaq)	fjord/airport
Store Gletscher (Sermeq Kujalleq in Qarajaq icefjord)	glacier
Store Hellefiskebanke	shallow offshore area
Storstrømmen	glacier
Sullorsuaq (Vaigat)	strait
Summit	ice-sheet station
Svartenhuk Halvø (Sigguup Nunaa)	peninsula
Swiss Station	ice-sheet station
T asiusaq	fjord
Thule Air Base (Pituffik)	airbase
Tininnilik	ice-dammed lake
Tissarissuq	former part of ice-sheet margin
Torsukattak	icefjord
Tuapaat	coastal stretch
U pernavik	town
Upernavik Isstrøm (Sermeq)	glacier
Ussuit	bay
Uummannaq Kangerlua (Uummannaq Fjord)	fjord
Uummannaq Fjord (Uummannaq Kangerlua)	fjord
V aigat (Sullorsuaq)	strait
Vaskebugt (Kangerluarsuk)	fjord

Appendix 2

Radiocarbon analyses

New radiocarbon ages reported in Tables 3 and 4 were determined by accelerator mass spectrometry (AMS). Dating was performed at the Ångström Laboratory at Uppsala, Sweden, under the supervision of Göran Possnert. The outer part of the shell material was removed by hydrochloric acid (HCl) to prevent contamination. The ages are reported in conventional radiocarbon years B.P. (before present = A.D. 1950). The dates have been corrected for isotopic fractionation by normalising to a $\delta^{13}\text{C}$ value of -25‰ on the PDB scale. The $\delta^{13}\text{C}$ measurements were performed on a conventional mass spectrometer. The dates have been corrected for a seawater reservoir effect by using an apparent age of 400 years (Bennike 1997). The reservoir-corrected ages have been calibrated into calendar years

before present (= A.D. 1950) using the INTCAL04 dataset and the OxCal version 3.10 software program (Bronk Ramsey 2001).

With respect to previous radiocarbon age determinations, compiled in Tables 2–4, those marked AAR and AA were also determined by accelerator mass spectrometry, whereas the other analyses were carried out by conventional methods. Older dates on marine material have been reservoir corrected by subtracting 400 years (laboratory codes AAR, AA) or no corrections were applied (laboratory codes I, K, Hel). The latter dates have been corrected for isotopic fractionation by normalising to a $\delta^{13}\text{C}$ value of 0‰ on the PDB scale, or no correction for isotopic fractionation was applied. These dates have a ‘built-in’ correction.

