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 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-

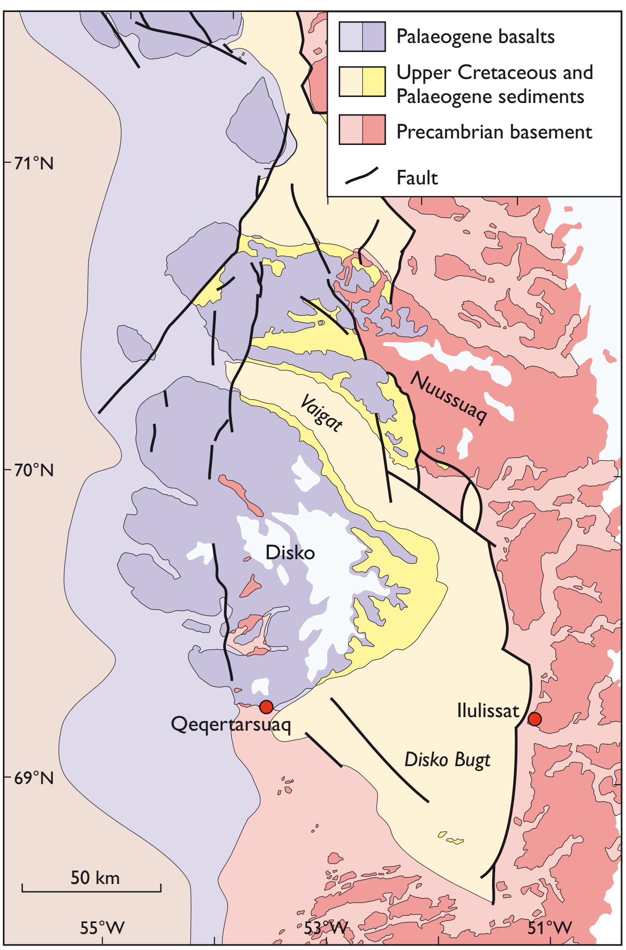


Fig. 16. Onshore (full colours) and offshore (pastel shades) bedrock geology of the Disko–Nuussuaq area. Modified from Chalmers *et al.* (1999) and Larsen & Pulvertaft (2000).

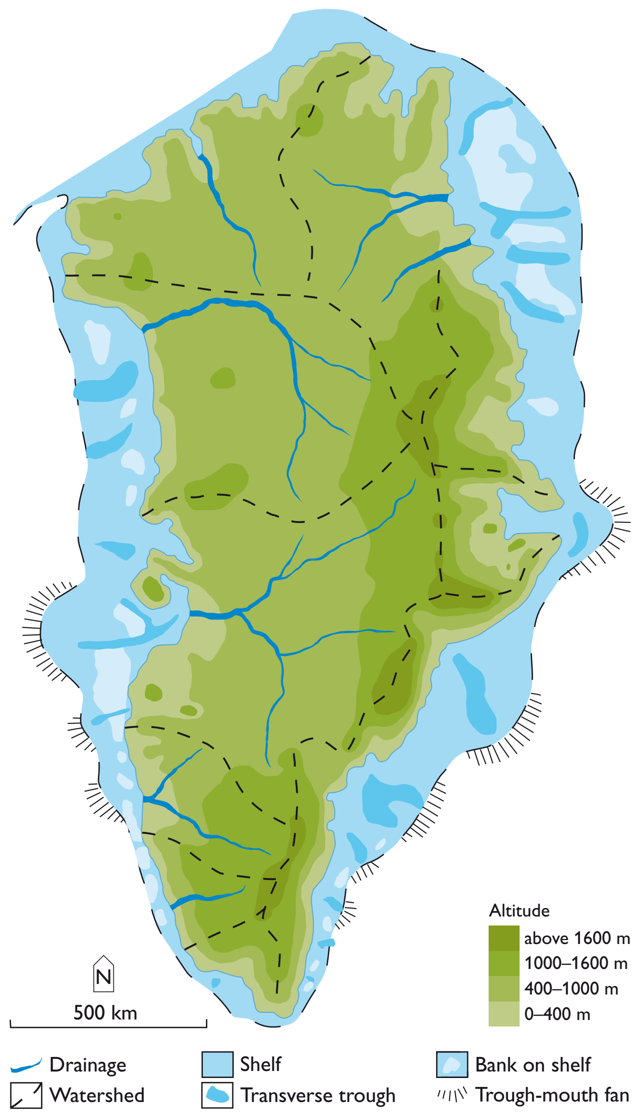


Fig. 17. Map of Greenland without the present ice cover, with altitudes corrected for the present load of the Inland Ice (cf. Fig. 4). In this reconstruction, large parts of central Greenland are shown to be drained by river systems, of which one major system flows into Disko Bugt. The trends of the 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

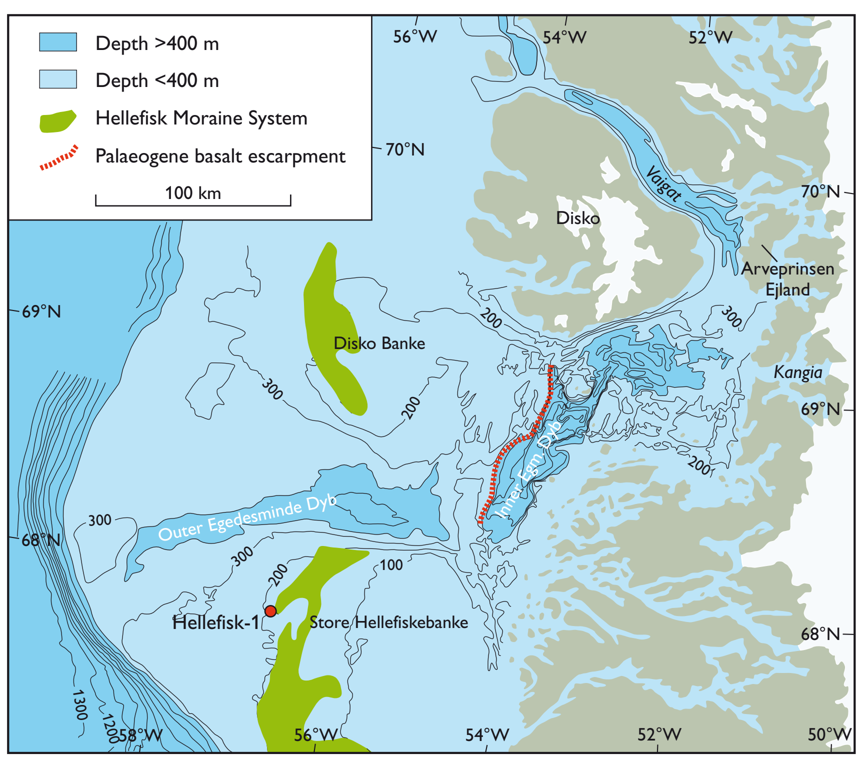


Fig. 18. The continental shelf offshore Disko and Disko Bugt. Bathymetry of the southern area (including Egedesminde Dyb) from Brett & Zarudzki (1979), and of the northern area (including Vaigat) from Chalmers *et al.* (1999). The Hellefisk Moraine System, comprising Disko Banke and Store Hellefiskebanke, 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 (Holtzscherer & 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, stepwise 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

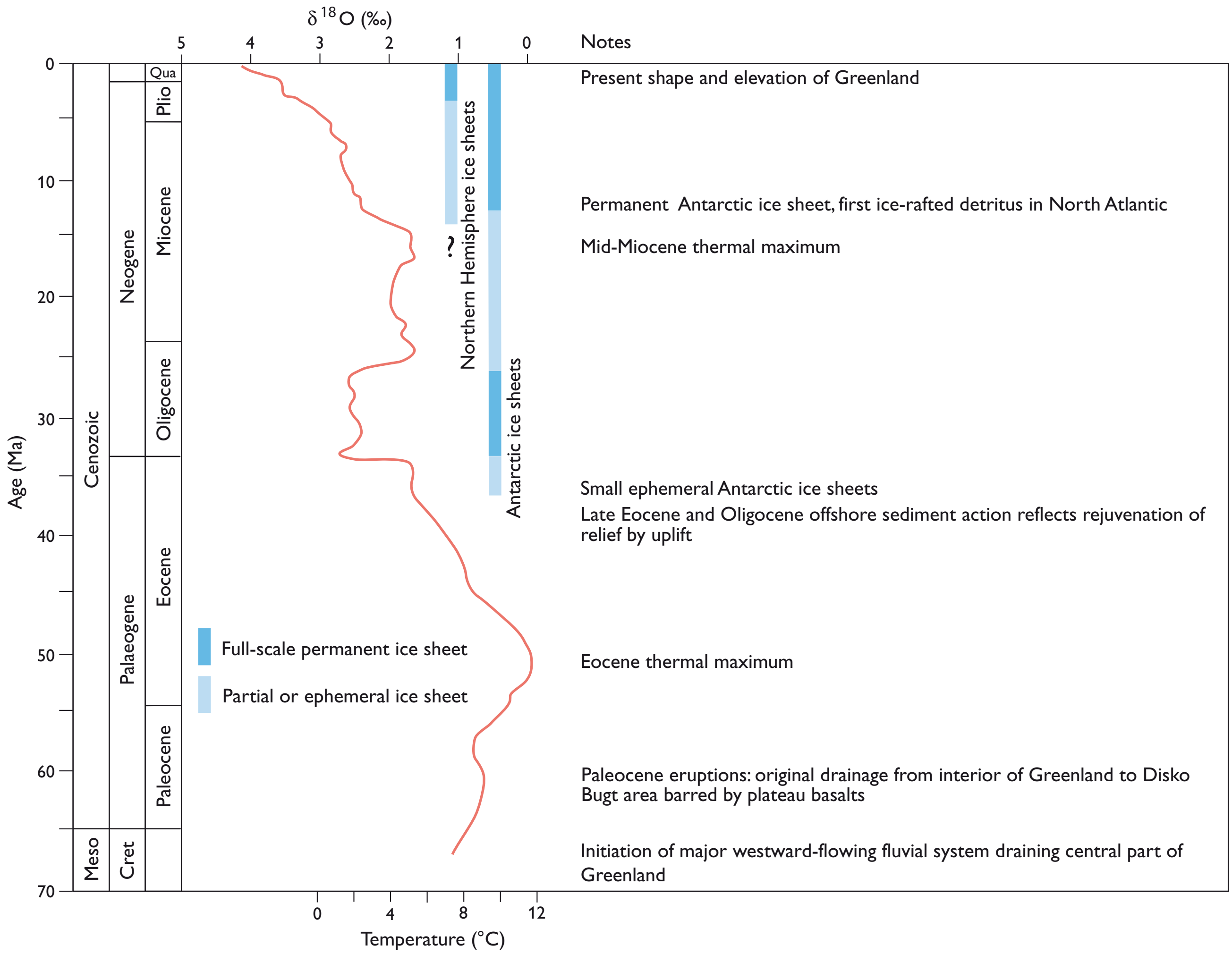


Fig. 19. Global cooling during the Cenozoic is illustrated by a temperature curve based on deep-sea oxygen isotope records (from Zachos *et al.* 2001); the temperature evolution is compared with global ice-sheet development and selected local events in Greenland. **Cret**, Cretaceous; **Meso**, Mesozoic;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

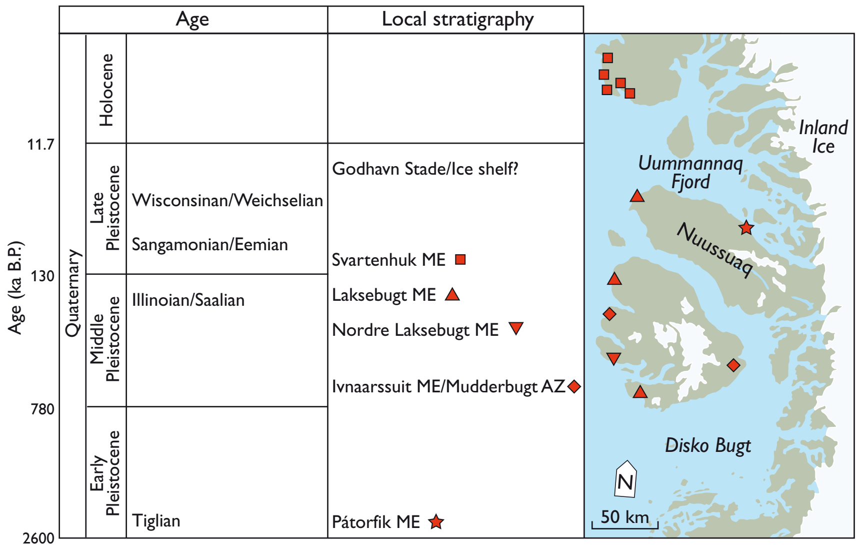


Fig. 20. Stratigraphic summary of interglacial and interstadial deposits between Disko Bugt and southern Svartenhuk Halvø, based on Kelly (1986) and Bennike *et al.* (1994). **AZ**, aminozone; **ME**, marine event. The accompanying map shows the location of these deposits; the symbols represent one or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

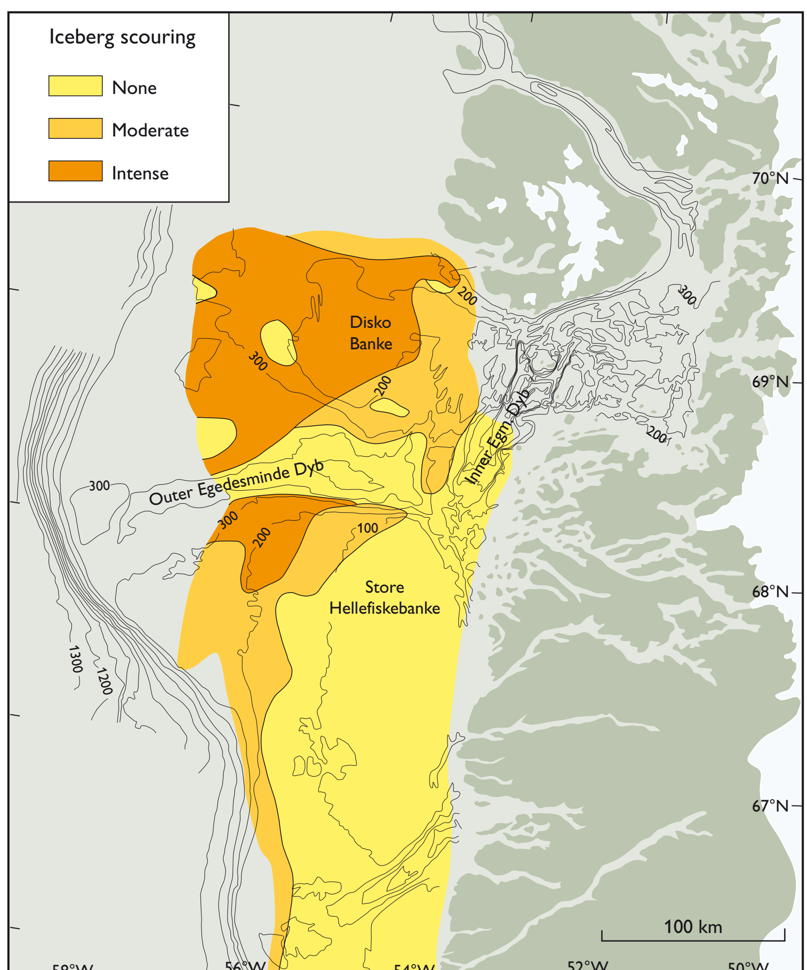
It was noted above that the Hellefisk glacial event may have been followed by the Svartenhuk marine event, which is referred to the last interglacial stage. The Sangamonian/Eemian interglacial has been described as more humid and warmer than the present interglacial, with a mean summer temperature up 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 (Letréguilly *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 morphological 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.

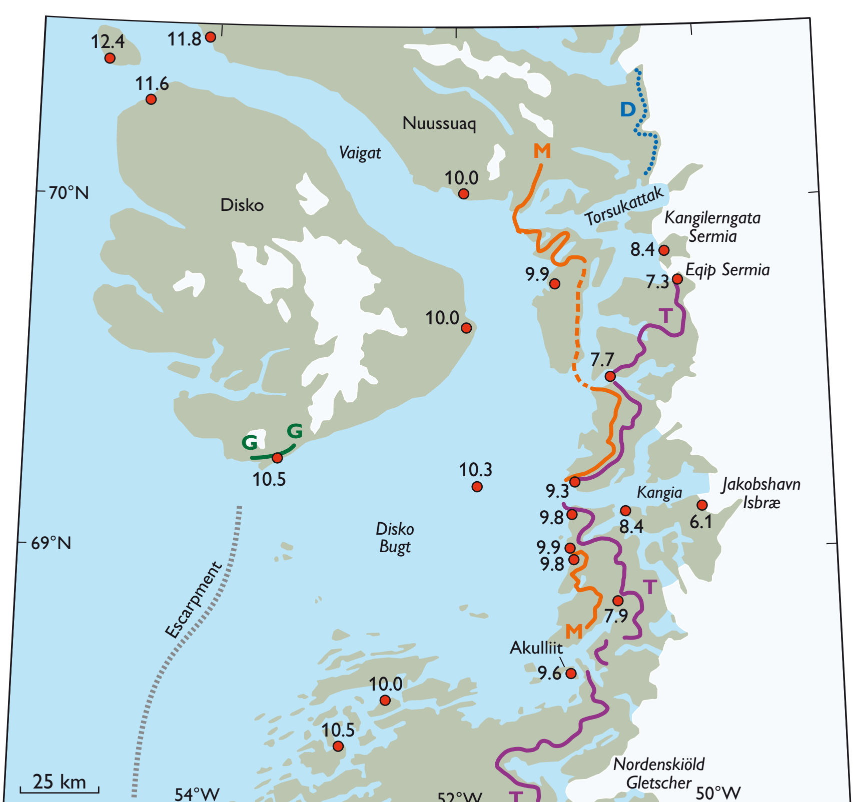


Fig. 22. Radiocarbon dates pertaining to the last deglaciation of the ice-free areas around Disko Bugt. Dates are given in calibrated thousand years before present (cal. ka B.P.), using the INTCAL04 data set for calibration (Reimer *et al.* 2004). Based on dates published by Ingólfsson *et al.* (1990), 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 Marrait 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 *et al.* (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 *et al.* (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 *et al.* (1990)
E Disko	69°40′	52°00′	K-3660	Shells	8700 ± 120	8700	10200 – 9500	9.9	Ingólfsson *et al.* (1990)
Arveprinsen Ejland	69°46′	51°15′	Beta-107879	Gyttja	8820 ± 100		10200 – 9550	9.9	Long *et al.* (1999)
Godhavn	69°17′	53°28′	AAR-5	Shells	9650 ± 250	9250	11250 – 9750	10.5	Ingólfsson *et al.* (1990)
Central Disko Bugt	69°11′	51°49′	AA-37711	Foraminifers	9483 ± 65	9083	10500 – 10150	10.3	Lloyd *et al.* (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 (Qeqertarsuatsiaq) 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 Qasigiannguit 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)
Qapiarfiit	69°52′	50°19′	2–3	K-3663	Shells	7600 ± 110	7600	8.4	–0.1	Ingólfsson *et al.* (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 Isfjeldsbanken; 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 Qasigiannguit 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 Isfjeldsbanken 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

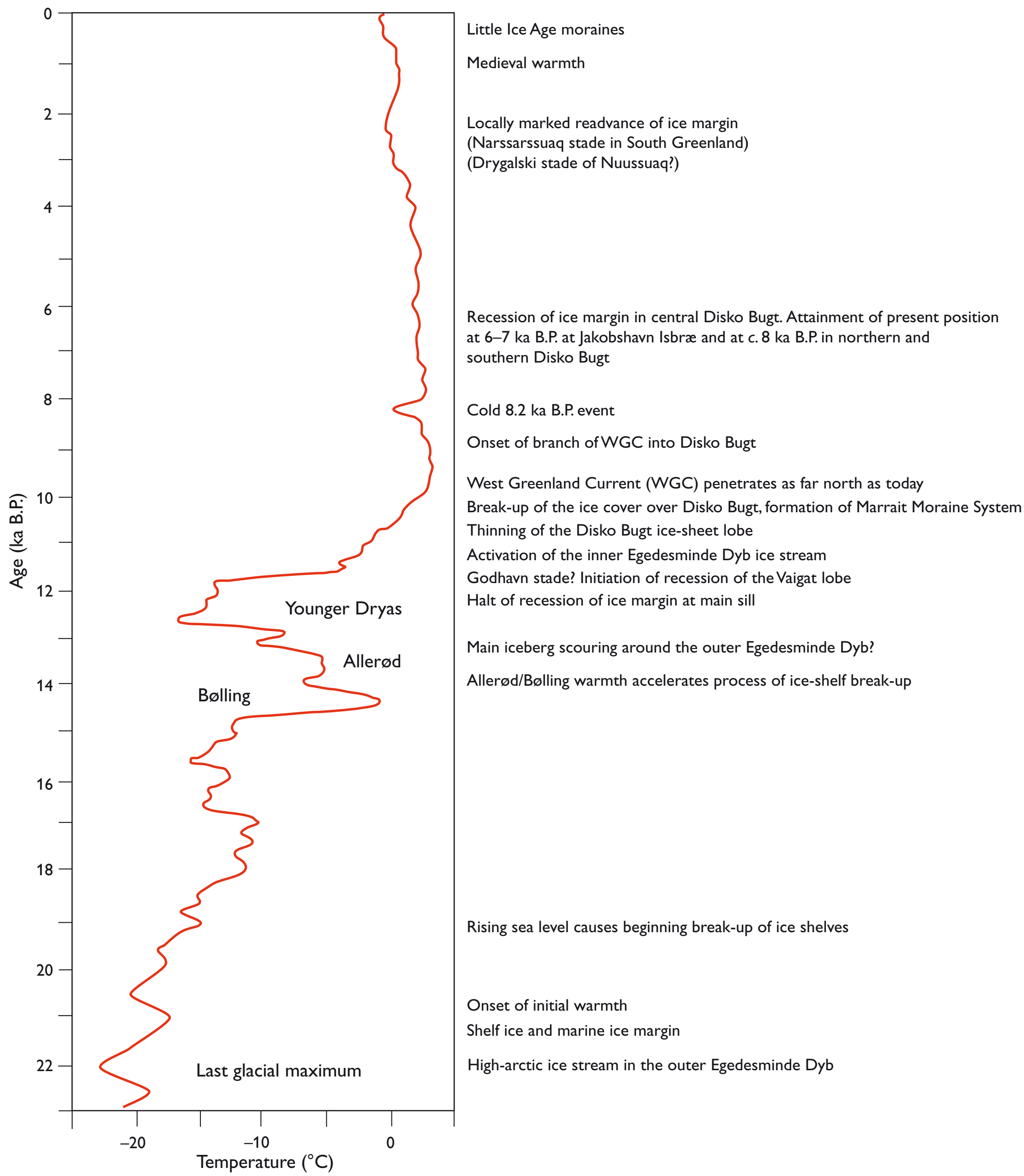


Fig. 23. Trend of the temperature development since the last glacial maximum, compiled from ice-core records (Dansgaard *et al.* 1984; Dahl-Jensen *et al.* 1998; Dansgaard 2004) and events related to the recession of the ice-sheet margin around Disko Bugt.

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 (Holtedahl 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 Isfjeldsbanken 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 (Donner & Jungner 1975; Ingólfsson *et a*l. 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 'Tasiussaq 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 Eqaluit (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 laiddown 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 Marrait 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).

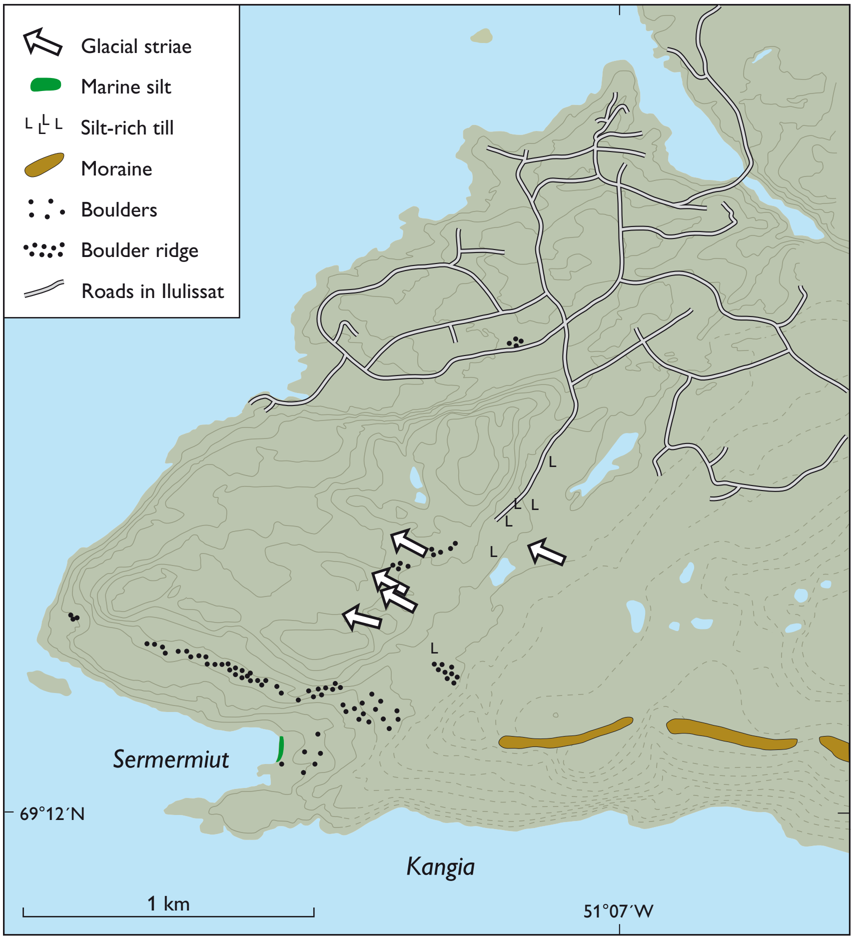


Fig. 26. Sketch-map of the area around Ilulissat (Jakobshavn) and Sermermiut showing Quaternary features. The map is based on the Geodetic Institute (now part of National Survey and Cadastre, Copenhagen) map sheets 1:2000, Jakobshavn and 1:8000, Sermermiut. Contour interval is 10 m. Redrafted from Weidick (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 Qapiarfiit 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 correlatives 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 Marrait 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 Marrait moraine system (around 9.9–9.4 ka B.P.), which probably developed in response to the change of environ-

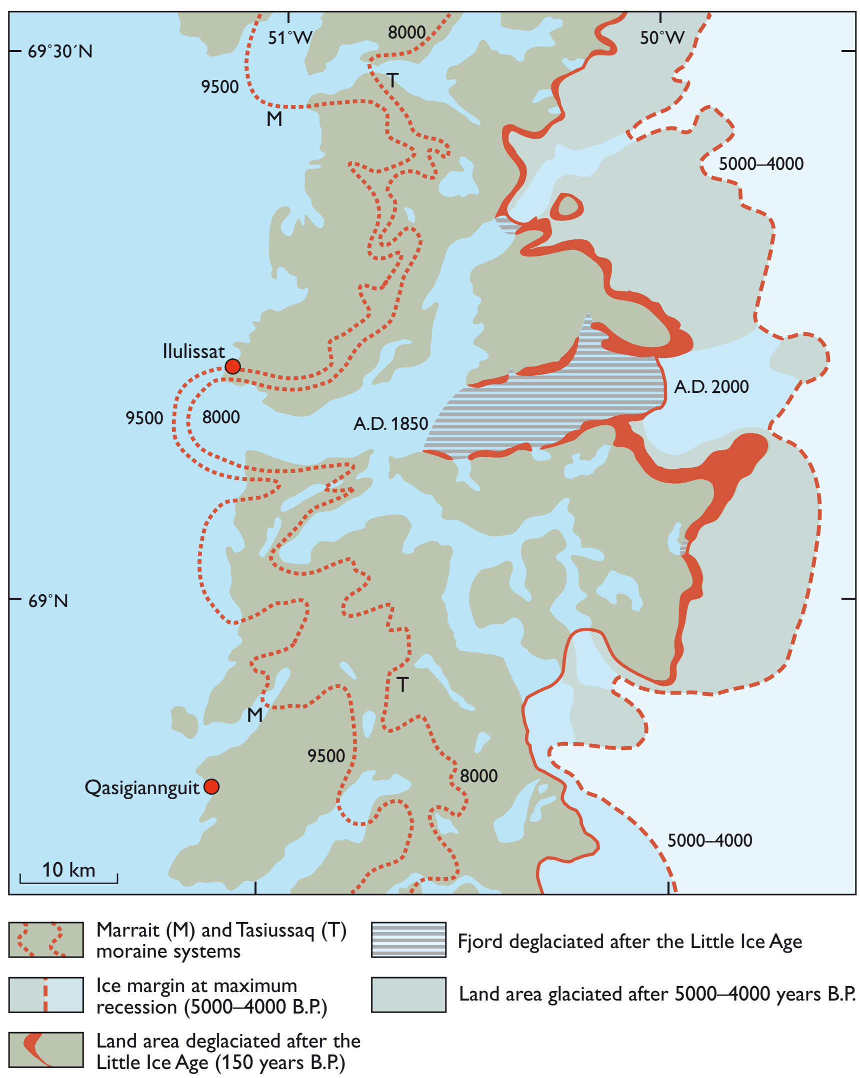


Fig. 27. Provisional reconstructions of the position of the ice margin in the Ilulissat area at *c.* 9500, 8000, 5000–4000 and 150 years B.P. The change of the ice margin between the Little Ice Age maximum (150 years B.P.) and the present day (the trimline zone) is shown in red for land areas and horizontal 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 calf 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 morphological 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 (Letréguilly *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† (± 2σ) years B.P.	Rcorr ^{14}C age§ years B.P.	Calib. age¢ ka B.P.	$\delta^{13}C$ ‰	Reference
Paakitsoq, *c.* 69°25´N, 50°20´W						
Macoma calcarea shell	Ua-4577	3300 ± 65	2900	3.1	–6.79	This study
Shell	Ua-699	3420 ± 105	3020	3.2	0¶	Weidick *et al.* (1990)
Mytilus edulis shell	Ua-4576	3560 ± 65	3160	3.4	–3.73	This study
Shell	Ua-1085	4520 ± 135	4120	4.6	0¶	Weidick *et al.* (1990)
Tissarissoq, *c.* 69°06´N, 50°02´W						
Mya truncata shell	Ua-4581	3590 ± 65	3190	3.4	1.88	This study
Hiatella arctica shell	Ua-4582	3940 ± 65	3540	3.8	1.92	This study
Mya truncata shell	Ua-4580	3945 ± 70	3545	3.8	2.50	This study
Mya truncata shell	Ua-4583	4075 ± 70	3675	4.0	2.16	This study
Odobenus rosmarus tusk	Ua-2350	4290 ± 100	3890	4.3	–13.05¶	Weidick (1992a)
Mya truncata shell	Ua-4579	5240 ± 75	4840	5.6	1.85	This study
Balanus sp. plate	Ua-4578	5710 ± 55	5310	6.1	0.98	This study
Alanngorliup Sermia, *c.* 68°54´N, 50°15´W						
Shell	Ua-1087	2620 ± 110	2220	2.2	0¶	Weidick *et al.* (1990)
Mya truncata shell	Ua-4585	2935 ± 60	2535	2.6	2.01	This study
Shell	Ua-1088	4000 ± 115	3600	3.9	0¶	Weidick *et al.* (1990)
Mya truncata shell	Ua-4584	4930 ± 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 Narssarssuaq 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 Narssarssuaq 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

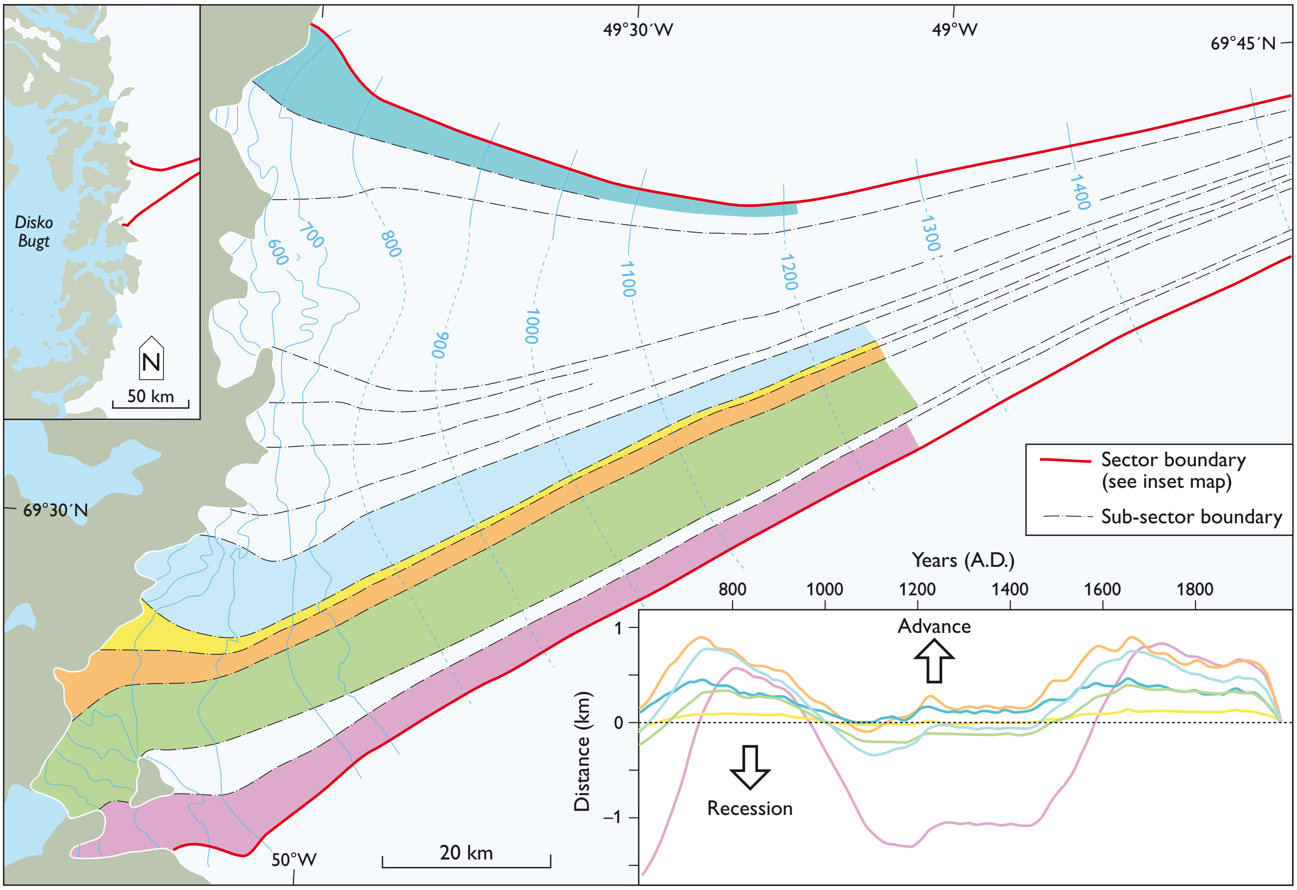


Fig. 28. Advances and recessions in the ice-sheet margin can be correlated with climatic variation deduced from the Dye 3 ice-core record. Lateral variation in the response was defined for a specific sector of the margin, north of Jakobshavn Isbræ (see inset map); sub-sectors (indicated by colours) show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

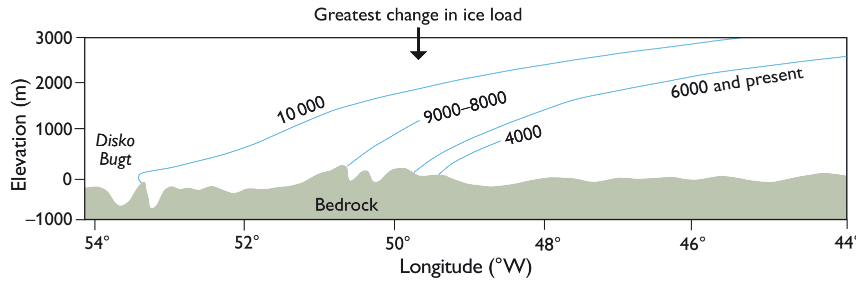
The beginning of the Little Ice Age advance may be related to the legend about Tissarissoq, the ice-filled bay south of Kangia. The name Tissarissoq is claimed to refer to a time when hunting in the bay was possible, that is to say when glacier ice did not cover the bay (Hammer 1883). This accords with 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).

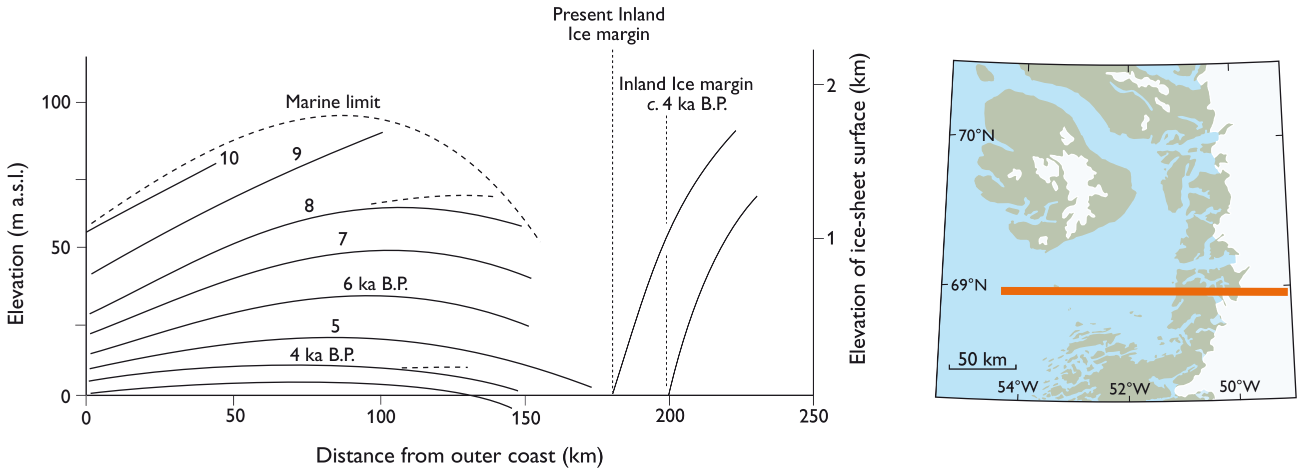


Fig. 33. Hypothetical shoreline relation diagram for Disko Bugt broadly following the 69°N line of latitude (see inset map); the position of the margin of the Greenland ice sheet at the present day and at *c.* 4 ka B.P. is indicated. Note that the dashed shorelines for 4 ka and 8 ka B.P. indicate the 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 Vaigait 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 Uummannaq 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), Akulliit/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-

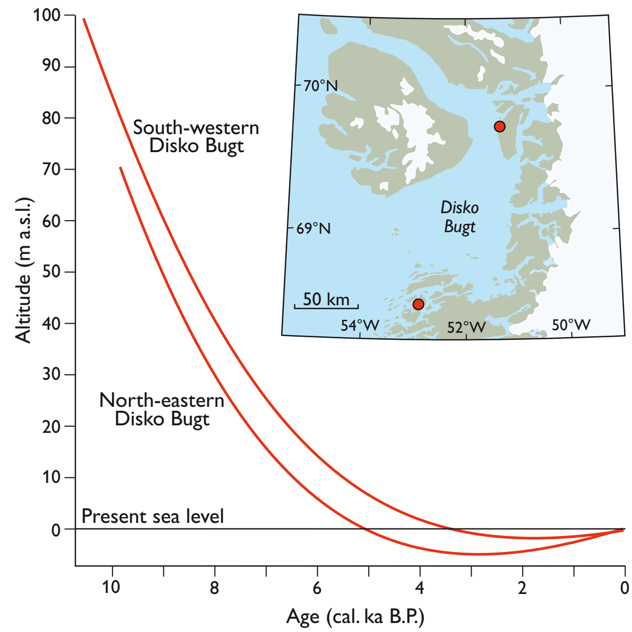


Fig. 34. Examples of two relative sea-level curves from the Disko Bugt area. The inset map shows the location of the sites in the Disko Bugt region. Modified from Long *et al.* (1999) and Long & Roberts (2003).

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.