Glaciation and the Quaternary of Greenland

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Fig. 360. Glacier lobes emanating from the Sukkertoppen ice cap. Head of Evighedsfjord, West Greenland. Route 207 H No. 04940. Copyright Geodetic Institute.

Tertiary prelude to the Quaternary

Tertiary deposits

The Tertiary deposits in Greenland are mainly of Paleogene age and are characterised by basalts, intercalated with limnic and marine sandstones and shales. The Neogene period is characterised by faulting and denudation processes which formed the major present landscape features of Greenland.

Paleogene deposits

The deposits occur scattered throughout central West and central East Greenland but correlation between them is difficult because the exact stratigraphic position and the age of the beds are unknown.

Fossil plants in the Lower Paleocene beds in West Greenland indicate a climate like the present day climate of Japan or China (warm temperate: 10° C mean annual temperature, 25° C summer temperature) and a depositional environment of great river systems deriving their material from areas now under the present Inland Ice (Koch, 1964). The Paleocene marine faunas show more affinities to the European faunas than to those of North America (Floris, 1972).

East Greenland marine faunas at Kap Brewster show a great stratigraphic range, from Middle-Upper Eocene (*Cyrena* beds) through Lower Oligocene (*Coeloma* beds) to Miocene (*Chlamys* beds) (Hassan, 1953). The marine and limnic deposits are interbedded with basalt layers which in West Greenland show clear indications of submarine eruption in the lowermost layers, whereas the upper plateau basalts were subaerially erupted. The basalt successions in both East and West Greenland are over 6 km thick. Since much of the basalt accumulation took place at sea level, the extrusions must have in-

Table 18. Chronology of the Tjörnes deposits

PLEISTOCENE			
1 m.y.	Redará	(2 glaciations)	
	Búrfell	(1 glaciation)	
2–3 m.y.	Breidavík	(4 glaciations)	
	Furuvík	(2 glaciations)	
PLIOCENE			
3–4 m.y.	*Cardium* Zone		Pacific Boreal species dominate. Correlated with English Red Crag Formation.
	Mactra and *Tapes* Zones		Pacific species. Climatic conditions as at present in the Mediterranean. Pollen flora of the intercalated lignite indicates mixed coniferous and deciduous forest and a warm temperate climate. The zones are correlated with the English Coralline Crag Formation.

Compiled according to Einarsson *et al.* (1967)

volved a widespread contemporaneous down-warping of the pre-basaltic substratum in these areas.

Neogene deposits

The basalt eruptions continued into Late Eocene or Oligocene times since the highest flows pre-date the *Chlamys* beds (Birkenmajer, 1972). No deposits from the later parts of the Tertiary period are known which has been explained by there being a period of extensive denudation.

The Skeldal conglomerate near Mesters Vig in East Greenland is, with some reservation, referred to the Pliocene-Pleistocene (Fränkl, 1953) but carries no fossil record. However, very good stratigraphical and palaeoclimatological evidence of the Pliocene-Pleistocene transition occurs at Tjörnes on the north coast of Iceland, only 500 km from Greenland. The Tjörnes deposits (Einarsson *et al.*, 1967) give the sequence in Table 18 interbedded with basalt flows. The marine faunas represent an influx of Pacific molluscs to Iceland during the late Pliocene or early Pleistocene indicating a relatively warm Arctic Ocean and an open Bering Strait.

Climatic conditions during the Tertiary

The poleward gradient of temperatures already existed in the Mesozoic, as Birkelund (1965) has shown from faunal evidence and Lowenstam (1964) on the basis of oxygen temperatures, but this tendency intensified during the Tertiary, culminating with the strongly marked 'ice age gradient' in the Quaternary. A graph of available data on temperature conditions in the North Atlantic area is shown in fig. 361 where the gradient change can clearly be seen.

There is little information available on the development and extent of oceans and shelf areas and the circulation of the ocean currents through the ages. Berggren (1970) has stated that during Cretaceous and Tertiary eras, a branch of the Gulf Stream flowed along the eastern margin of Newfoundland and Labrador and that this was suddenly terminated with the development of icebergs in the Labrador Sea about 3 million years ago. Frakes & Kemp (1973) stated that interchange of water in the main branch of the Gulf Stream, between the North Atlantic and the Arctic Basin, was probably due to the widening of the gap between Greenland and Europe. However, the increased interchange of water throughout the Tertiary and the gradual cooling do not seem to be directly related on the basis of present information. In this context the Icelandic Tjörnes deposits indicate an open poleward connection between Pacific and Atlantic waters at the end of the Tertiary period.

Palaeogeographic conditions

Variation in situation and extent of Greenland

According to the ideas on sea floor spreading, the Davis Strait rift formed during the Jurassic, and the North Atlantic rift east of Greenland in the Cretaceous period (Pitman & Talwani, 1972). Since that time Greenland has been free of the continents and surrounded by deep ocean basins. Today only a narrow stretch at Thule has a continental shelf in common with North America. However vague the

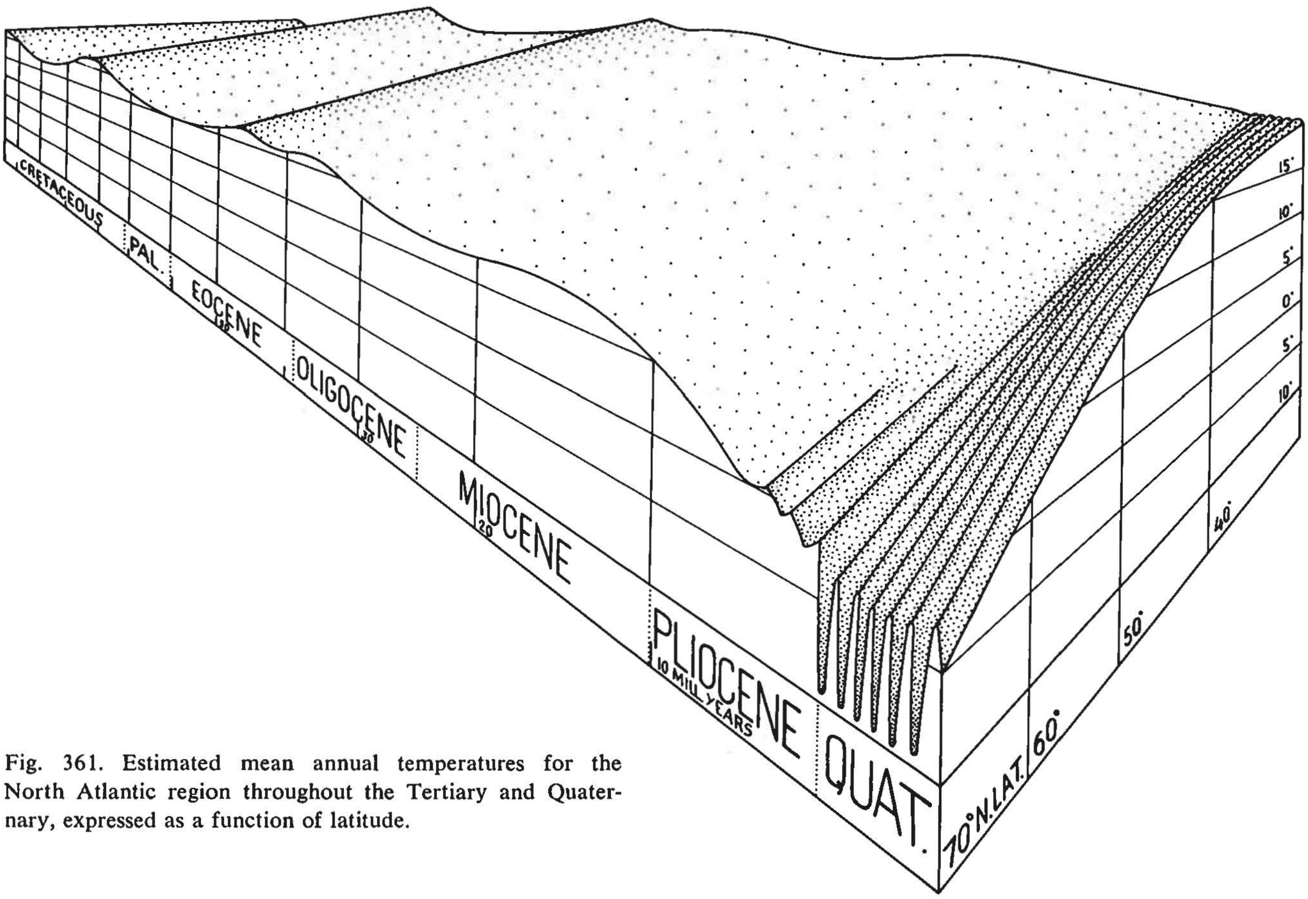


Fig. 361. Estimated mean annual temperatures for the North Atlantic region throughout the Tertiary and Quaternary, expressed as a function of latitude.

palaeogeographic evidence of Greenland is it seems justified to envisage Tertiary Greenland as an island similar to its present situation and possibly also to its present size.

Vertical movements

The present Greenland coastal strip is mainly dominated by alpine landscapes or elevated plateaux with one high summit level and some secondary erosional surfaces. Towards the north, the summit levels of the southern and eastern alpine parts generally merge into more widespread plateaux of the Norwegian type. The highest and oldest erosion levels appear to cut the Tertiary basalt, as well as Precambrian bedrock and seem therefore to have been formed since the deposition of the basalts, i. e. in Miocene time or younger.

The post-Miocene peneplanation must have lasted for a long time since several kilometres thickness of the basalt flows were removed, and uplift therefore took place in the late Miocene, Pliocene or Pleistocene. Whether the uplift resulted in elevations greater than at present and was followed by subsequent downgrading to present altitudes is not known. It is possible that grading to some degree kept pace with uplift resulting in a nearly continuous uplift to the present maximum heights, and with some secondary erosion surfaces developing during this time. The 'falling sea level' curve of Zeuner (1946) indicates that with a rate of decrease in relative sea level of 1 m every 13 000 years the average summit level of 2000 m would be 26×10^6 years old (Lower Miocene) which might be little more than a mere coincidence.

Final pre-glacial form and conditions necessary for ice sheet formation

Fig. 362a shows the approximate outline of the pre-glacial relief of Greenland assuming that the limitational and altitudinal conditions were approximately the same as the present deglaciated part of Greenland, and that the subglacial valleys mapped by the P. E. Victor expeditions (Holtzscherer & Bauer, 1954) in fact represent Tertiary river drainage systems. The altitudes of the present subglacial base are corrected for the present glacial load of the Inland Ice.

A crucial factor is the length of time during which this pre-glacial landscape existed. It is generally accepted that the present Antarctic Ice Sheet reached its present extent in the Eocene (Geitzenauer *et al.*, 1968), Oligocene (Frakes & Kemp, 1973) or Miocene

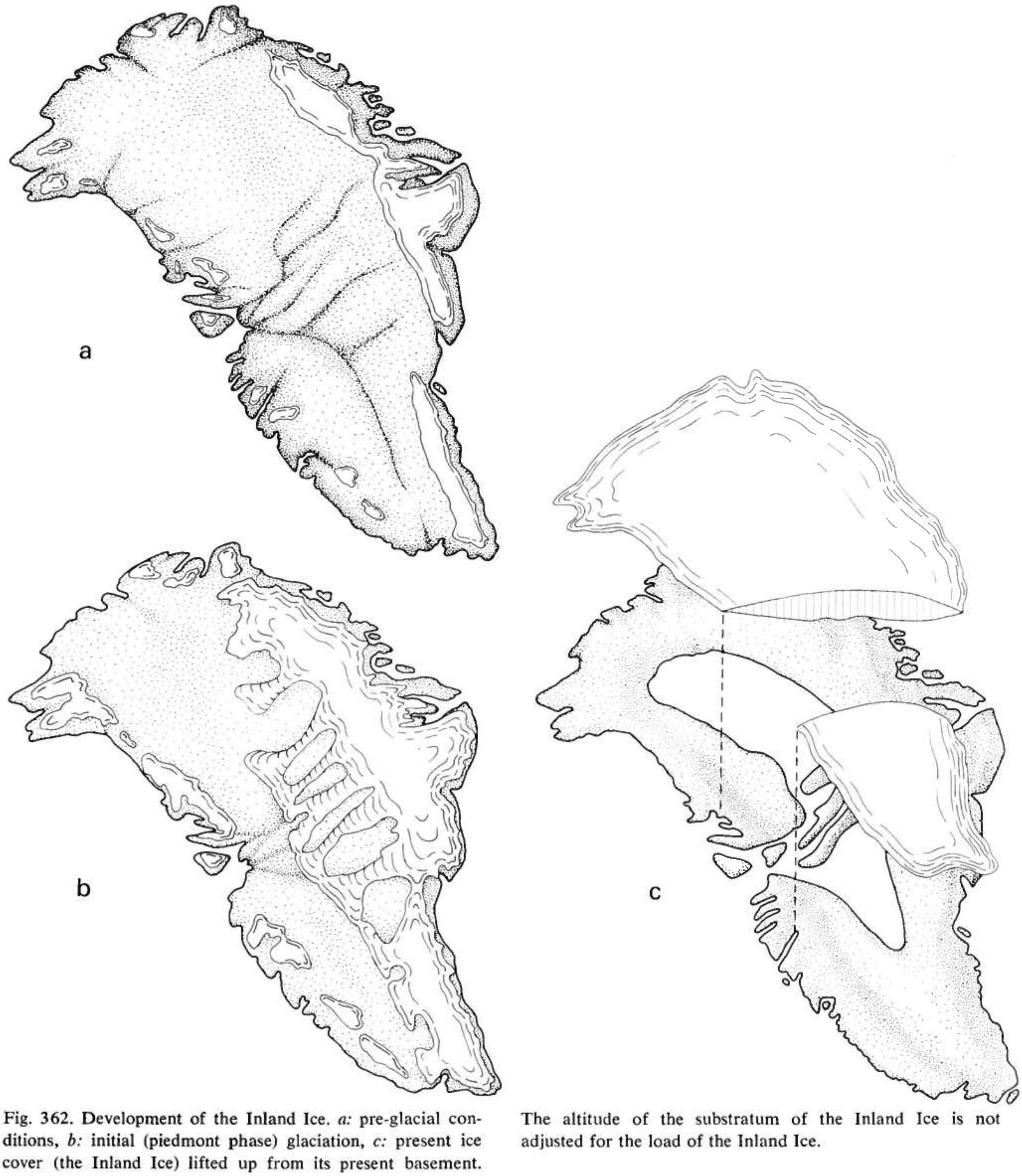


Fig. 362. Development of the Inland Ice. *a:* pre-glacial conditions, *b:* initial (piedmont phase) glaciation, *c:* present ice cover (the Inland Ice) lifted up from its present basement. The altitude of the substratum of the Inland Ice is not adjusted for the load of the Inland Ice.

(Bandy *et al.,* 1971; Flint, 1971) and the same Tertiary (Miocene) age was supposed by Wager (1933) for the Greenland Inland Ice. However, if the palaeoclimatic curve of fig. 361 is accepted, the Miocene climate of Greenland was similar to that of present day southern Europe with a glaciation limit (lowermost limit of glacier formation) at least at 2000 m above sea level near the coast and increasing to 3000 m in the interior of Greenland, according to the continental updoming of the glaciation limit. With the present deglacial altitudes of Greenland only small local glaciers would have existed in the highest areas. In comparison the glaciation limit today can be seen to vary from 500–700 m above sea level near the coast to 1500–1700 m further inland (fig. 363).

Another line of evidence is the statement of Berggren (1970) that glacial marine sediments from the North Atlantic marine cores were first recorded 3 million years ago. The lack of ice rafted debris in older sediments favours the viewpoint that Tertiary Greenland was not glaciated.

The Pliocene-Pleistocene boundary

The Pliocene-Pleistocene boundary is still a matter for discussion. Two different opinions can be seen in Einarsson's stratigraphic division of the Tjörnes deposits (Table 18) and Berggren's stratigraphic scheme in fig. 364. For both it is evident that a non-glacial climate prevailed in Greenland until around 3 million years ago, assuming present altitudinal conditions.

It is considered that the Inland Ice was formed during the first glaciation which may have been Donau or Eburon. Successive glaciations are assumed to have increased in intensity in the northern hemisphere reaching a maximum 0.4–0.2 million years ago (Illinoian-Saale glacial stages). Steenstrup (1883) expressed the view that the Inland Ice would not have been formed under prevailing climatic conditions, an idea favoured by more recent knowledge of the altitudes of the deglaciated part of Greenland and the trend of the present glaciation limit (cf. figs 362a, 363).

Wegmann (1939) considered that once mountain glaciers had spread sufficiently to form piedmont glaciers then subsequent development of the main ice sheet would be rapid, because of the autocatalysis of glaciation (fig. 362a–c). The ice sheet is distinct from local glaciers by both its wide extent and independence of the topography of the substratum. The Inland Ice is here defined as the lens-shaped body of firn and ice situated in the interior of Greenland; it may have been in existence continually since the early Pleistocene, though an interglacial disappearance cannot be completely excluded (see later).

Different estimates have been made for the length of time for the formation of the Inland Ice and other ice sheets. These range from 100 000 years (Holtzscherer & Bauer, 1954), greater than 50 000 years (Flint, 1971), 30 000–15 000 years (Weertman, 1964; Barry, 1966), to 10 000–5000 years (Lamb & Woodroffe, 1970). If the development and disappearance of the Wisconsin-Weichsel ice sheets are envisaged as having taken place within 70 000 years, then the shorter estimates for its formation are to be preferred. During development of the ice sheet there may have been an increase in precipitation such as was discussed by Loewe (1971) for the glaciation of North America.

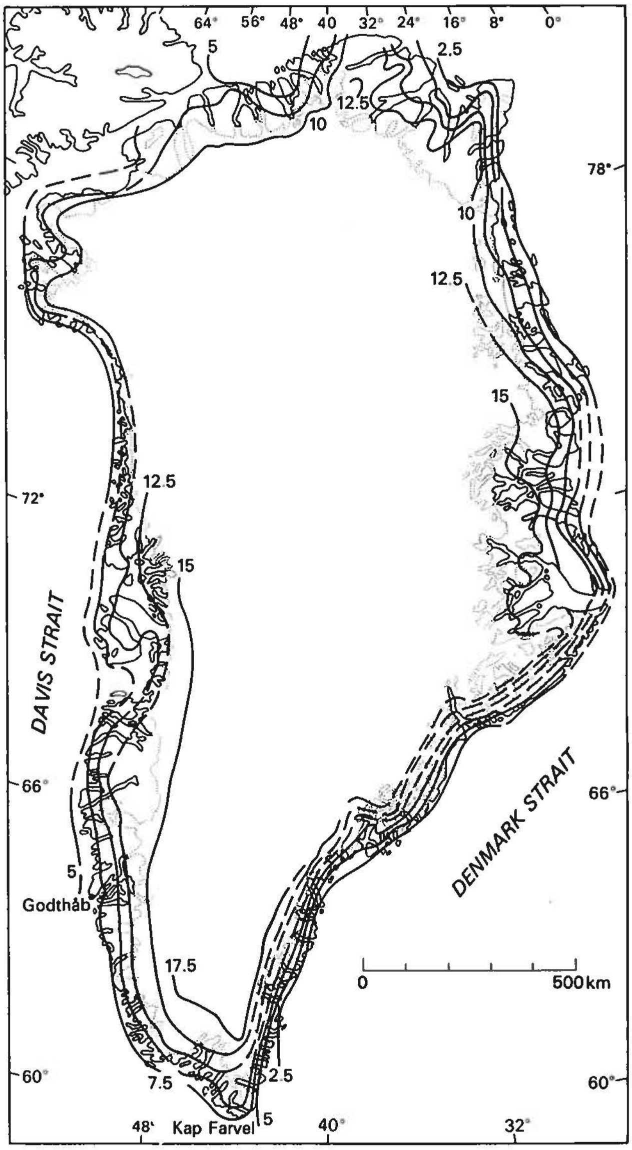


Fig. 363. Present altitudes of the glaciation limit. Figures are given in hundreds of metres.

Pleistocene

Glaciation of the outer coast

That Greenland was at least once nearly completely covered by ice is suggested by the glacial sculpturing of the outer coastal region and the N–S trending ridges on offshore banks which represent moraines deposited during the glacial epochs.

Some of the outermost mountains were never covered by the Inland Ice even during the maximum glaciation as can be shown by an upper limit of ice margin deposits, erratics and schliffgrenzen (upper limit of glacial abrasion of bedrock). The potential nunatak regions of fig. 365 are based on this evidence. The upper limit varies in West Greenland from 300–900 m at the outer coast to over 1000 m in the

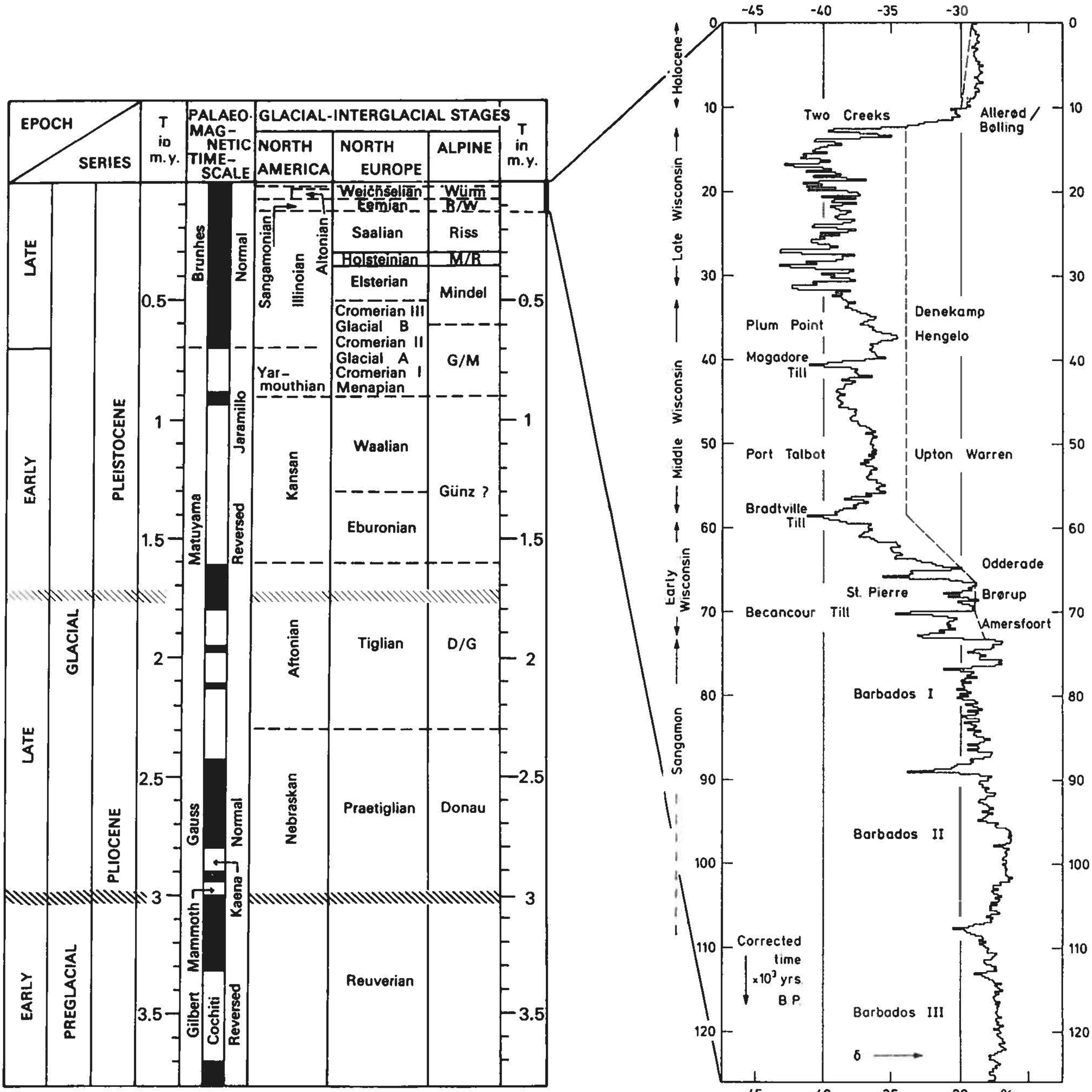


Fig. 364. *Left:* shortened stratigraphic scheme for the North Atlantic regions (modified from Berggren, 1970). *Right:* climatic conditions in Greenland throughout the last 125 000 years according to the Camp Century ice core (Dansgaard *et al.*, 1973).

central parts of the ice-free areas. Data on such limits in North and East Greenland are minimal but show the same trend of increasing altitude inland and to the same heights. These limits are indicated on the Quaternary Map of Greenland 1:2 500 000 (1971) published by the Geological Survey of Greenland.

The nunatak hypothesis

The geological evidence for the existence of nunataks during the maximum extent of the Inland Ice is supported by certain botanists (Gelting, 1934; Bøcher, 1956, 1959) who consider that the local occurrence of certain plant species reflects their survival on nunataks during the various glaciations. South-facing valleys on the nunataks, even during the ice ages, may have had a local climate sufficiently warm for the survival of relatively warm-loving plant species. However, Iversen (1953) has opposed this idea and considers the spreading of seed by birds, wind or ocean currents to be more important. The fact that the plant species are not always found

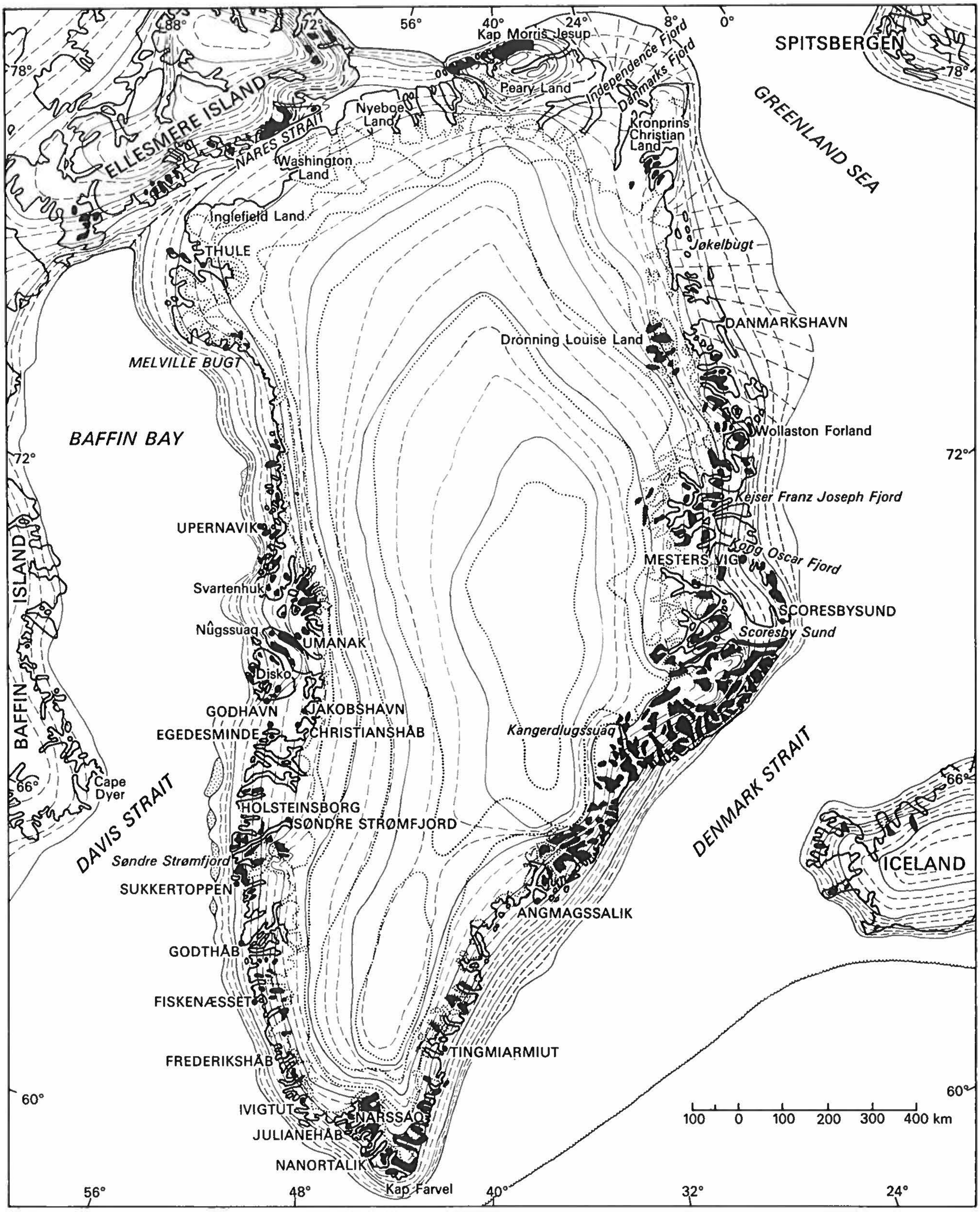


Fig. 365. Presumed extent of the Inland Ice during maximum glaciation. *Black areas:* potential nunatak regions. *Dotted areas:* parts of the West Greenland banks presumed to have been above sea level. *Dotted lines:* margin and contour lines of the present Inland Ice. *Saw-tooth line:* maximum southern extent of polar pack ice.

at the same locations as the potential nunatak areas could be explained by a secondary Holocene dispersal.

Glaciations older than Wisconsin-Weichsel

Repeated glaciations of the outer coast are evident from the occurrence of double cirques and U-shaped valleys. It can be seen from the direction of glacial striae at several localities that the initial glacial conditions led to development of the cirques before the full development of the continental ice sheet extended over the coastal region.

It is not yet possible to recognise distinct features of individual ice ages. The morphology of the submarine moraines on the banks off the west coast of Greenland shows that the proximal moraines slope gently seaward to submarine terraces at –70 to –110 m, which may indicate former sea levels related to stages marked by the moraines, some of which may be older than Wisconsin-Weichsel.

Assuming that successive glaciations reached their maximum extent 0.2–0.4 million years ago, it is possible that the outermost of the moraine systems forming the banks off West Greenland (cf. fig. 365) represents an Illinoian-Saale glaciation. In East Greenland, the Kap Mackenzie stadial may be referred to an Illinoian-Saale glaciation (Funder & Hjort, 1973).

Davies (1972) put forward the idea that in North Greenland, due to increasing aridity throughout the Quaternary, the Inland Ice reached its greatest extent possibly during the Kansan-Mindel glaciation, whereas during the Illinoian-Saale it was less extensive and by the Wisconsin-Weichsel it extended not much beyond that at present. In general, global climates during the Quaternary and late Tertiary periods (the last thirty million years) are presumed to have been marked by aridity and progressively colder interglacials (Fairbridge, 1973). Therefore the land around the Arctic Basin was probably dry throughout the Quaternary. So far, the evidence supports Davies' concept of increased dryness throughout Quaternary time, not *vice versa.*

The idea of the decreasing extent of the subsequent glaciations in North Greenland is supported by investigations on glacial drift in Inglefield Land. According to Tedrow (1970) three drift sheets can be distinguished in the southern part of the area: (1) an older, boulder-rich till from a glaciation extending over Inglefield Land in which the boulders are strongly weathered; (2) weathered yellow-coloured drift from glacial readvance reaching to around 10 km from the coast at Rensselaer Bugt; (3) grey-coloured drift, well preserved landforms, and situated closer to the present Inland Ice than (2).

A build-up of river terraces was presumably connected to this stage (3). Two feet below the surface of such a terrace organic matter was found which has given an age of 20 800 ± 2900 B.P.

According to this division at least Tedrow's stage (1) may be older than the Wisconsin-Weichsel glacial epoch.

However, in general the theory of a small expansion of the Inland Ice during the last glacial period (Wisconsin-Weichsel) conflicts with the fact that the late and rapid Holocene glacio-isostatic uplift of the whole of northern Greenland can only be explained as a reaction from the retreat of a young, presumably Wisconsin-Weichsel glaciation over the whole area extending to the outer coasts. Furthermore, Nichols (1969) arrived at the conclusion that the Inglefield Land area was completely glaciated during the Wisconsin-Weichsel which is substantiated by C^{14} dates on marine Holocene deposits. The chronology of Table 19 for Inglefield Land is based on Tedrow's division but the contradiction with Nichols' results must be borne in mind.

Interglacial stages in Greenland

In West Greenland the first indications of interglacial deposits were observed by Steenstrup (1883); in his description of the locality Pátorfik on the north coast of the Nûgssuaq peninsula, he records a shell-bearing moraine. Subsequent investigations and dating of deposits in this area revealed the occurrence of *Alvania wyvillethomsoni* Friele var. *patorfikensis* Laursen in the deposits indicating an interglacial age; the species is not encountered in postglacial marine deposits (Rosenkrantz, 1968; Símonarsson, 1970). In this respect the species may be parallel to *Venerupis senescens* from the Danish Eemian. The C^{14} age of the shells in the deposits is more than 35 000 years old.

Subsequently early dates on shell material from West Greenland have been reported from Nordre Isortoq (Kelly, pers. comm.) and Svartenhuk Halvø (Laursen, 1972) at approximately 67°20′ and 71°30′ respectively (fig. 366). Early dates may also be expected in the shell material in till between the fjords Kangerdluarssugssuaq and Søndre Strømfjord at altitudes up to 390 m above sea level (Sugden, 1972), though not enough material is yet available for C^{14} dating.

Near the Inland Ice margin at Frederikshåbs Isblink and at the head of Godthåbsfjord in West Greenland, marl concretions were collected by S. Hansen and K. Gripp (Gripp, 1932) and investigated

Table 19. Summary of the Quaternary of Greenland

			West Greenland	East Greenland	North Greenland
HOLOCENE		Historical time (A.D. 1500–A.D. 1974)	Smaller readvances until 1900 in West and East Greenland and until the 1920s in North Greenland. Overall general small recession of glacier lobes since 1920.		
		5000–1500 B.P.	Minor fluctuations of the margin of an essentially stable Inland Ice.		
		6000–5000 B.P.	Readvance of glacier lobes over marine deposits. Adjustment of the Inland Ice to present stability.		
		10 000–6000 B.P.	Main postglacial recession of the Inland Ice margin, interrupted by halts marked by the following substages:		
			Mt. Keglen 7000–7200 B.P.	Rødefjord 7000 B.P.	
			Fjord 8400–8100 B.P.		Outer moraines in Independence Fjord 8200 B.P.?
			Avatdleq 8700 B.P.		
			Taserqat 10 000–9000 B.P.?	Milne Land 10 000 B.P.	
UPPER PLEISTOCENE	Wisconsin-Weichsel glacial age	Late (20 000–10 000 B.P.) Stadials	Oldest moraines on the banks? Or identical with the Avatdleq or Fjord stages (at Sanerâta timâ 63°30′N)	Milne Land stadial maximum extent of Late Wisconsin-Weichsel Inland Ice with an ice margin at the entrance of the fjords and with the outer coast relatively ice-free.	Inglefield Land only partially covered by the Inland Ice during the time of this stadial?
		Middle (50 000–20 000 B.P.) Mainly an interstadial complex	Date of Frederikshåbs Isblink (21 710 B.P.)	Two groups of dates: 24 300–19 500 B.P. >40 000–33 600 B.P. Both groups include the Jameson Land interstadial	Date of lower terrace in Inglefield Land: 20 800 B.P.
		Early (> 50 000 B.P.)		Extensive glaciation of outer coasts of East Greenland: Flakkerhuk and Kap Mackenzie stadials. Possibly Kap Mackenzie stadial must be referred to an Illinoian-Saale glaciation.	
	Sangamon-Eem interglacial age				Occurrence of interglacial ice in the Camp Century core. Older soil in Inglefield Land?
			Interglacial disappearance of the Inland Ice?		
	Illinoian-Saale glacial age		Inland Ice to the offshore banks in West Greenland?	Extensive Kap Mackenzie glaciation in East Greenland?	
MIDDLE PLEISTOCENE			Unknown		
LOWER PLEISTOCENE			Formation of the Inland Ice at Pliocene-Pleistocene transition?		

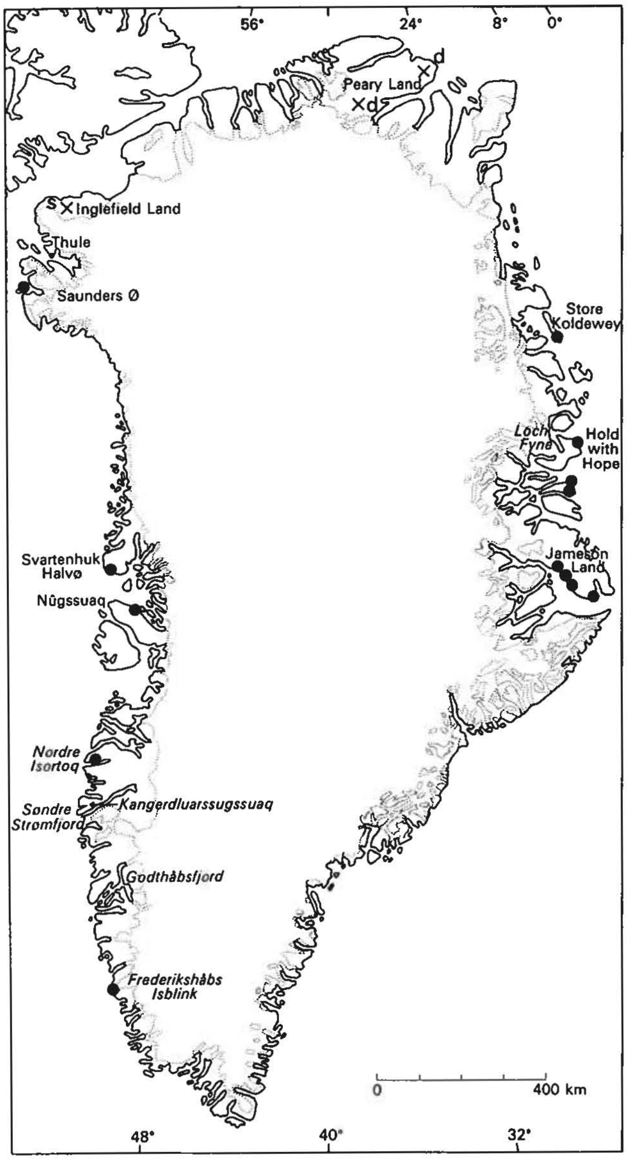


Fig. 366. Location of samples dated as more than 20 000 years B.P. Shell material marked by filled circles; crosses mark other material (*d* driftwood, *s* soil).

East Greenland in Jameson Land (Funder & Hjort, 1973). Funder & Hjort correlate a Kap Mackenzie stadial of total glaciation with the Illinoian-Saale glaciation, and the less extensive Flakkerhuk stadial with the early Wisconsin-Weichsel glaciation (cf. Table 19), which may indicate the occurrence of the Sangamon-Eem interglacial in East Greenland. This twofold division of glaciation can be continued farther north as is revealed by old radiocarbon dates (C. Hjort, personal communication) on shell material from the outer coast at Hold with Hope (73°40′ N) and from Store Koldewey (76°10′ N; shell material described by Jensen, 1917).

In North Greenland, C^{14} dating of shells in till-like deposits on Saunders Ø near Thule gave ages of more than 32 000 years B.P. (Davies *et al.*, 1963). According to Tedrow's division some of the high terrace levels between his stages (1) and (2) may be related to a Sangamon-Eem interglacial. From the outer coast of Peary Land, Davies (*in* Trautman & Willis, 1966) described wood (willow) 137 m above sea level at the top of a terminal moraine presumed to mark the maximum Wisconsin advance of the continental ice sheet (fig. 368). In the interior part of Peary Land, at the Midsommersøer, Fredskild (1969) also reported wood in Holocene marine terraces with an age greater than 35 000 B.P. The latter may represent redeposited interglacial driftwood.

So far, none of these deposits can definitively be placed in any specific interglacial or interstadial period but their frequent occurrence throughout Greenland at least implies the possibility for localising interglacial deposits. In this context, sparker profiles or corings of the banks off West Greenland may contribute to a better knowledge of the division of glacial–interglacial periods in Greenland, situated as they are at the very margin of the maximum glaciations.

for pollen by M. Bryan (1954). The presence in two of the specimens of *Picea mariana* pollen suggests some of the concretions are of interglacial age. However, long distance transport of *Picea* pollen has been reported in postglacial Greenland sediments (Fredskild, 1973) and Bryan's conclusion may therefore be invalidated.

In East Greenland at Loch Fyne, a peat layer was found underlying gravel and overlying ground moraine, such that the peat may be considered as interglacial or interstadial (Backlund, 1931); no C^{14} dating has so far been made on the material. However, C^{14} dates older than 35 000–40 000 B.P. have been obtained on shell material from the outer coast of East Greenland in Jameson Land (Funder & Hjort,

Extent of the Inland Ice during interglacial periods

Climatic conditions in the Sangamon-Eem were sufficiently warm, as shown by the Toronto flora of Sangamon age in Canada, that a partial or complete disappearance of the Inland Ice is at least possible (Flint, 1947). The Sangamon sea level at 6 m above the present also corresponds well with the sea level rise of 6–7 m which would accompany total disappearance of the Inland Ice. Mercer (1968) has suggested, however, that the Sangamon high shorelines reflect essentially the melting of the west Antarctic ice sheet, which being grounded below sea level would be sensitive to climatic changes. He considers the land-based Greenland Inland Ice to

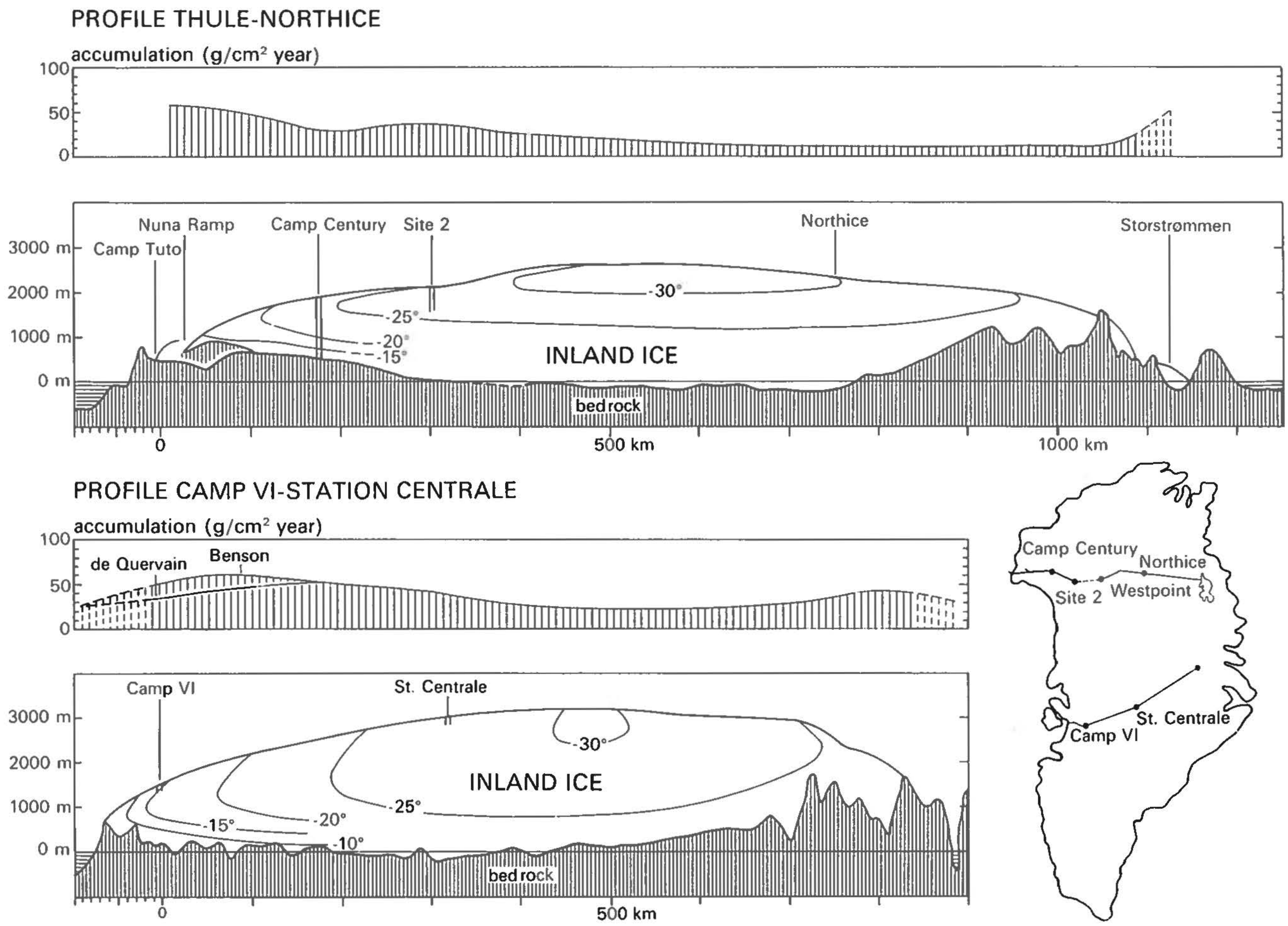


Fig. 367. Profiles of the Inland Ice across north and central Greenland. Positions of the sections are shown on the sketch map to the right. Shown on the profiles are the thickness of the Inland Ice, estimated temperature conditions in the Inland Ice and the depth of the core holes at Camp Century, Site 2, Camp VI and Station Centrale. Above each profile is shown accumulation over the same area.
Data of the northern section: accumulation, Benson (1962, fig. 29, p. 38); surface temperatures, ibid. (1962, fig. 37, p. 54); internal temperatures, Hansen & Landauer (1958, fig. 3) and Weertman (1968, fig. 7, p. 2698); altitude of substratum, Bull (1955, 1957) and Roethlisberger *et al.* (1965). Data of central section: accumulation, Benson (1961, fig. 3, p. 22) and de Quervain (1969, fig. 32, p. 134); internal temperatures, Robin (1955, fig. 5, p. 531); altitude of the substratum, Holtzscherer & Bauer (1954, fig. 13, p. 20).

have been less vulnerable to changes in temperature.

The ice core taken on the Inland Ice at Camp Century (fig. 367) also shows that at least the northern part of the Inland Ice survived the last interglacial, if the time scale used is correct. However, Emiliani (1969) says that the conditions for the Greenland Inland Ice to exist appear to be more critical than those of Antarctica "apparently because it is not centrally located with respect to the pole and because it extends to considerable lower latitudes." Emiliani believes that significant melting of the Greenland Inland Ice did take place after interglacial conditions were established, causing high interglacial sea levels.

Any reduction in the size of the Inland Ice would result in a relative decrease in the accumulation area compared to the ablation area. The smaller the ice sheet becomes, the greater the amount of accumulation required to maintain its existence. The annual accumulation required to maintain a pleniglacial Inland Ice is assumed to be 12–15 cm water equivalent and for the present Inland Ice with marginal retreat of *c.* 200 km in comparison about 31–37 cm water equivalent (Table 20). A retreat of a further 100 km would necessitate a mean accumulation of at least 70 cm water equivalent (Weidick, 1975) and bring it very close to its threshold of existence, beyond which it would disappear as did the other ice sheets in the northern hemisphere.

There is evidence in some places that the Inland Ice margin receded at places 20 km behind its present position during the Holocene postglacial climatic optimum. A recession of 100 km should therefore be inside the range of the effects of an interglacial longer and warmer than the time and climate of the Holocene (cf. fig. 364). The total disappearance of the

Table 20. Estimates of mass balance data, area and volume of the pleniglacial and present Inland Ice

	Pleniglacial	Present
Accumulation area	2.25 10^{6} km^2	1.44 10^{3} km^2
Ablation area	0.05 10^{6} km^2	0.29 10^{6} km^2
Total area	2.30 10^{6} km^2	1.73 10^{6} km^2
Mean accumulation over the accumulation area	12–15 10^{-5} km water	31–37 10^{-5} km water
Mean ablation over the ablation area	50–70 10^{-5} km water	100–110 10^{-5} km water
Mass balance		
Total gain by accumulation	305 km^3 water/year	500 km^3 water/year
Loss by ablation	30 km^3 water/year	295 km^3 water/year
Loss by calving	275 km^3 water/year	205 km^3 water/year
Volume	3.21 10^{6} km^3 water	2.35 10^{6} km^3 water

Inland Ice during a long and warm interglacial period cannot be excluded, but it is still unproven.

Wisconsin-Weichsel

Division

A threefold division of the Wisconsin-Weichsel glacial period into Early, Middle and Late seems to be generally accepted (Flint, 1971) and has been used here in Table 19. The Middle Wisconsin-Weichsel (*c.* 50 000–20 000 B.P.) is a period of relatively warm spells between the cold periods of Early and Late Wisconsin-Weichsel.

Extent of glaciations

In West Greenland there is no evidence of the extent of the Inland Ice in the Early Wisconsin-Weichsel, though a radiocarbon date on shells from fresh moraines in front of Frederikshåbs Isblink furnished an age of 21 710 years B.P. indicating that the present glacier lobe covers interstadial sediments. In the same area the subsequent readvance of Late Wisconsin-Weichsel reached the outer coast, but presumably not beyond, since dates of undisturbed marine beds at Marraq and Sanerâta timâ (both near Godthåb) indicate the position of an Inland Ice lobe between 14 000 and 8000 B.P. Further north in West Greenland the recession of the Inland Ice in the Holsteinsborg area was initiated before 10 000 B.P. (Weidick, 1972a, b), and the same is the case in the Kap Farvel area (Fredskild, 1973) and in the Frederikshåb area (Kelly, 1974). The ice margin in these areas must have been situated at or beyond the present coast.

In East Greenland the observations and datings of Funder & Hjort (1973) reveal sound evidence of a Middle Weichselian shrinkage of the Inland Ice (Jameson Land interstadial complex). At this time the western part of Scoresby Sund (Hall Bredning) was largely unglaciated. The subsequent readvances of the Inland Ice (Milne Land stadial) did not reach as far as the earlier Flakkerhuk stadial preceding the Jameson Land interstadial. Neither in Scoresby Sund nor in Kong Oscars Fjord did the ice of the Milne Land stadial reach the open coast, but the Inland Ice seems first to have receded from the position at this stage at around 10 000 years B.P.

The date of 20 800 years B.P. on soil in Inglefield Land, North Greenland (Tedrow, 1970), points to the possibility that parts of Inglefield Land have not been covered by ice since then. However, evidence furnished by data from both sides of the waters between Greenland (Davies *et al.*, 1963; Nichols, 1969) and Canada (Prest, 1969) seem to stress the possibilities of a confluence of the Innuitian ice sheet (Blake, 1970, 1972) with the Inland Ice. This fits well into the pattern of a subsequent deglaciation of the Thule area before 8500 B.P. (Davies *et al.*, 1963) or even 10 000 B.P. (Blake, 1972) and of Inglefield Land in a period between prior to 7800 B.P. in the south and prior to 5900 B.P. in the north. If both lines of evidence are justified, the soils of Inglefield Land have not suffered glacial abrasion during the readvance of the Inland Ice in the Late Wisconsin.

In other parts of North Greenland the Holocene uplift of Peary Land indicates an ice cover of the same extent as in West and East Greenland (Weidick, 1972b). Peary Land was mainly glaciated from a local centre around Frederick E. Hyde Fjord which

Fig. 368. Mudderbugten (to the left) and Kap Eiler Rasmussen (to the right) at eastern end of Peary Land, North Greenland. Moraine terrain modified by marine action. White line indicates the approximate marine limit of the area (approximately 70–90 m above sea level; Troelsen, 1952; Davies, 1963). 18 km north of Mudderbugten is the location of willow on top of terminal moraine (Wisconsin-Weichsel?) dated to more than 32 000 years B.P. (Davies *in* Trautman & Willis, 1966). Route 548 D-N No. 1208 (11.7.1949). Copyright Geodetic Institute.

merged with the Inland Ice north of Independence Fjord (fig. 368).

Main events

The main events are summarised in Table 19. During Early Wisconsin-Weichsel the Inland Ice may have extended to the outer coast, or beyond to the offshore banks in West Greenland, i. e. close to its maximum extent. Evidence of a subsequent substadial retreat in Middle Wisconsin-Weichsel is generally accepted in East Greenland as well. The following expansion of the Inland Ice during Late Wisconsin-Weichsel was not of the same extent as that preceding the interstadial. Although being a rough generalisation, the Inland Ice seems to have either reached or maintained its maximum position in the Late Wisconsin-Weichsel before 10 000 B.P. in West Greenland, 10 000 B.P. in East Greenland, and possibly 8000–9000 B.P. in north-east Greenland.

In Table 20 the pleniglacial conditions are estimated for an Inland Ice reaching the offshore banks. However, under pleniglacial conditions only small variations in accumulation would suffice to vary the extent of the Inland Ice to those of Late Wisconsin-Weichselian and even to interstadial conditions.

Holocene ice cover and its changes

Ice margin deposits

In West Greenland reconnaissance mapping of all the ice margin deposits has been made (Weidick, 1968), and was followed by detailed mapping between Nordre Strømfjord and Søndre Strømfjord, where ice margin deposits are especially abundant, by Kelly (1969), Ten Brink (1971, 1975) and Weidick (1972b).

In North Greenland, widespread ice margin deposits were first found in Hall Land by Koch (1928) and later mapped and described in detail by Davies (1963, 1972). Davies also described the deglaciation of the Thule area in north-west Greenland, although based on very scattered ice margin deposits.

The extensive moraine systems of East Greenland were described first around Ella Ø by Gelting (1934). More recently Funder (1972) has described those of the Scoresby Sund region in an account of its deglaciation.

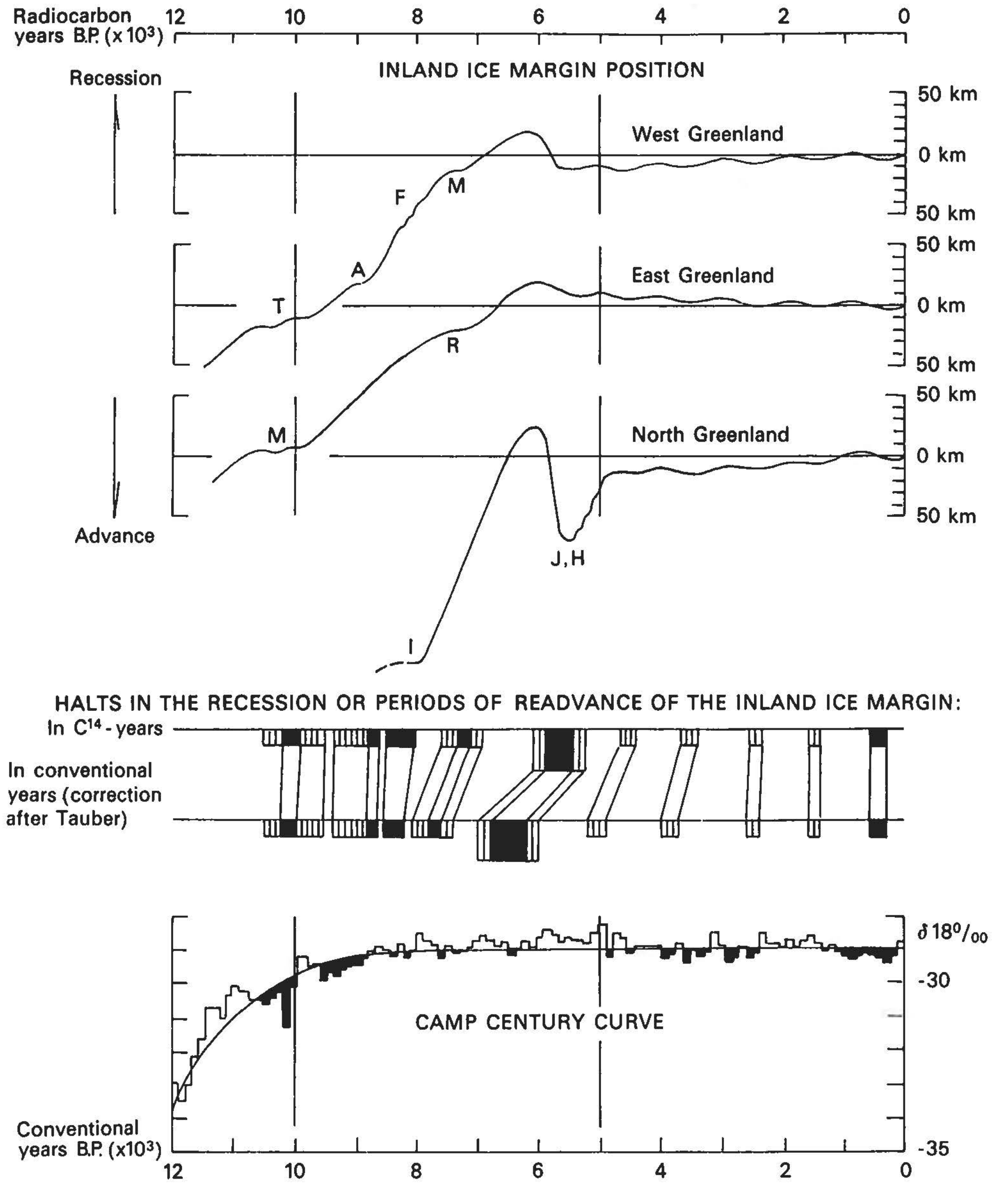


Fig. 369. *Top:* Generalised curves showing the fluctuations of the Inland Ice margins in West, East and North Greenland.
Letters indicate the following stages in the ice margin position: West Greenland: T = Taserqat, A = Avatdleq, F = Fjord and M = Mt. Keglen stage.
East Greenland: M = Milne Land and R = Rødefjord stages according to Funder (1972).
North Greenland: I = stages at the entrance of Independence Fjord, J = Jørgen Brønlund Fjord and H = Hall Land stages.
Centre: Approximate age of halt periods in the recession, or of readvance of the Inland Ice margin, according to the top curves. Black: dated periods at present. Hatched areas: alternative ages of dated periods or minor readvances of the last 6000 years. The uppermost series indicates age of the periods in radiocarbon years, while the lowermost series gives the same periods after the correction proposed by Tauber (1970). The ages thus should be more comparable to those of cold periods of the Camp Century record given below.
Bottom: The Camp Century record according to Dansgaard *et al.* (1973). Relatively cold spells are given in black.

The recession of the Inland Ice in East and West Greenland can be dated by sea levels of known age related to the successive moraine systems, and a correlation of this information has been made with the glacial events in North Greenland, where marine levels have also been dated. Fig. 369 is compiled from the best known sectors of West, East and North Greenland and expresses these glacial events as the linear change of the ice margin.

The Holocene fluctuations of the Inland Ice margin

The curves of figs 369 and 370 are not representative for the whole of Greenland as they are based on key areas. It is known, for example, that the Holsteinsborg area of West Greenland was still partially glaciated in 8000 B.P. while in the Frederikshåb area to the south (Kelly & Funder, 1974) and in Disko Bugt to the north (Weidick, 1968; Donner & Jungner, 1975) the ice had retreated almost to its present position. However, in general figs 369 and 370 show that:

(1) Nearly all recession occurred between 11 000–10 000 B.P. and 6000–5000 B.P.

(2) The recession rate culminated somewhat before 8000 B.P. in West and East Greenland, and approximately a millenium later in North Greenland. This is in accordance with the concept that initial deglaciation began in South Greenland.

(3) Recession went beyond the present limits of the Inland Ice, and after 6000–5000 B.P. was succeeded by a period of pulsations bringing the ice sheet to its present position.

(4) The pulsations since 6000–5000 B.P. are best described as variations in the marginal positions of ±20 km about an equilibrium position close to the present. The size of the fluctuations seem to be strongly dependent on topographic factors.

Changes in mass balance of the Inland Ice since Wisconsin-Weichsel

The postglacial recession of the Inland Ice must have been initially caused by a rise of mean temperature, together with related changes of other climatic parameters. Tentative estimates of the change in mass balance of the Inland Ice are shown in fig. 370. This model is based on the assumption that the pleniglacial glaciation limit and firn line were approximately 600 m lower than at present (cf. fig. 363).

According to the Camp Century ice core record a rather sudden shift from pleniglacial to postglacial temperatures occurred about 13 000–12 000 years B.P., and it is therefore likely that the shift in the position of the glaciation limit (and firn line) from glacial to present altitudes also occurred at this time. However, this change in glaciation limit involved only a small increase in the ablation area and is insufficient to explain the subsequent rapid shrinkage of the Inland Ice, unless accumulation increased only slowly and did not reach present magnitudes until 8000–6000 B.P.

In parts of East and North Greenland it is possible that an increased winter accumulation contributed to maintenance of the Inland Ice for a time. The increase in accumulation may have initially been restricted to the fringe of the Inland Ice, subsequently developing into a general increase in accumulation over the entire surface related to the warming up of the North Atlantic waters and the disappearance of the Laurentide ice sheet (Weidick, 1975).

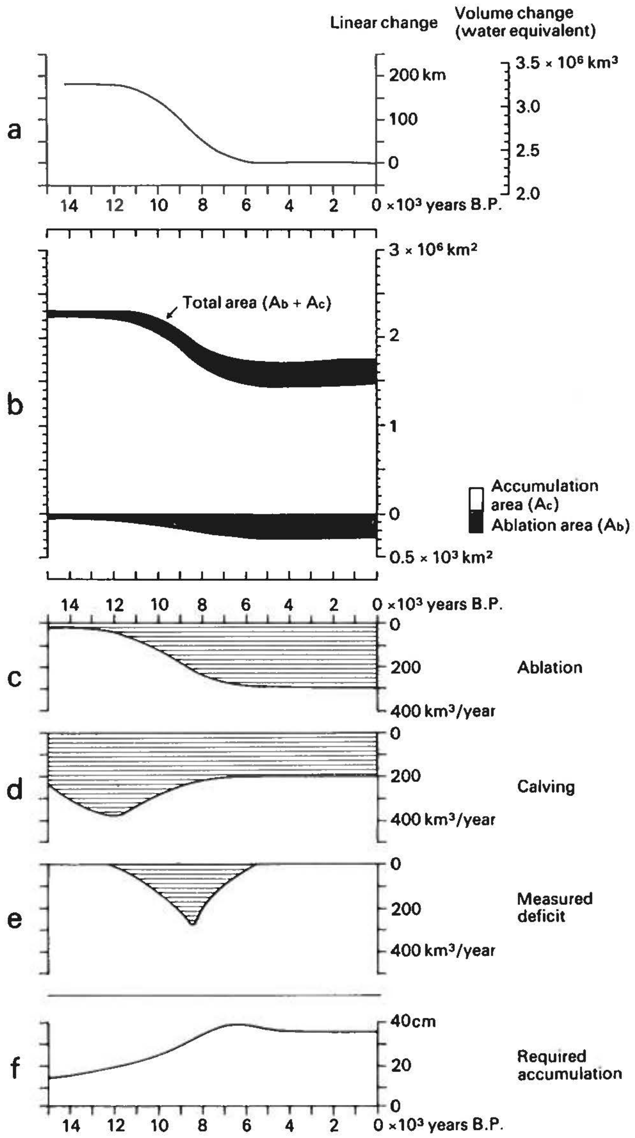


Fig. 370. Changes of the mass balance of the Inland Ice through Holocene time, estimated on the basis of the model of fig. 369.

a. The average recession of the Inland Ice margin. The recession is also converted to volume shrinkage.

b. Shrinkage of the total area (A) and the accumulation area (A_c) and the coherent growth of the ablation area (A_b). The ablation area is corrected for the altitude of the marginal basement.

c. Loss by ablation in km³/year of water equivalent according to the change in the ablation area given in *b*.

d. Loss by calving in km³/year water equivalent.

e. Loss in volume due to shrinkage according to the trend given by the volume change in *a*.

f. Accumulation (cm water/year) required to match the losses given by the curves of *c*, *d* and *e*.

Fig. 371. Marine beaches at Flade Isblink, northern East Greenland, partly truncated by the existing glacier. Route 665D-SV No. 15278 (11.8.1951). Copyright Geodetic Institute.

The stabilisation of the Inland Ice since 6000–5000 B.P. can only be explained by an increase in accumulation. This may have been initiated at the onset of Atlantic time (7500–5300 B.P.) but could only be effective after a build up period in the accumulation area. So far, this viewpoint is tentative and no conclusions with regard to changes in precipitation can be drawn from palynological evidence (Fredskild, 1973; Kelly & Funder, 1974); however, current investigations of the firn in the Inland Ice (Dansgaard *et al.*, 1973) may produce conclusive data on accumulation fluctuations throughout the Holocene.

Holocene minimum extent of the Inland Ice

The occurrence of shell, peat and wood remnants in many marginal deposits of the lobes of the present Inland Ice as well as of local glaciers suggests a smaller ice cover in the past. In places such deposits have even been found beneath the present ice margin or in the ice itself (Weidick, 1972a).

Marine terraces at levels up to approximately 40 m above sea level are reported covered by present lobes of the Inland Ice in Bredefjord, South Greenland (Kelly, 1974) and in Nordre Strømfjord, Arfersiorfik and Disko Bugt in West Greenland (Jahn, 1938; Weidick, 1972a). Lobes from local glaciers have also been seen to cover such deposits in the Sukkertoppen district and possibly also in the western parts of Disko (Sugden, 1972; Weidick, 1972a).

In North Greenland, Davies *et al.* (1963) presented evidence that the Pitugfik Gletscher and Harald Moltke Bræ had Holocene frontal positions at least 11 and 16 km respectively east of their present positions; further north, the Marie Gletscher was stated in 1923 to have had a glacier front partly covering a delta formed at a sea level 10 m above the present (Koch, 1928). In the eastern parts of North Greenland shells have been observed on the side of Academy Gletscher at the head of Independence Fjord 10 km behind its 1920's front (Koch, 1928). East of here the local glacier of Flade Isblink is seen to cover at least a lower system of 12–14 m marine terraces (Laursen, 1954; Weidick, 1972a; fig. 371).

Widespread occurrences of marine deposits inundated by the margin of the Inland Ice are described from Germania Land, East Greenland (Koch & Wegener, 1917; Jensen, 1917) and ice coverage of the 35 m marine terrace level in Scoresby Sund is reported by Funder (1972).

The observations referred to above indicate that before or at 6000 B.P. the Inland Ice must have had its margin 10–20 km behind its present situation. The observation of Jahn (1938) of a subsequent readvance to a position 12 km west of the present margin in Arfersiorfik fjord is contemporaneous with the formation of a 10–12 m marine terrace, and there is the possibility that the readvance occurred at 6000–5000 B.P. in West Greenland. The same seems to be the case in North Greenland according to the observations by Davies (1963), whereas in East Greenland the Inland Ice after its minimum has slowly advanced to its present position.

Late Holocene glacier fluctuations

All the information given points to a relatively constant size of the Inland Ice as well as of local glaciers throughout the last 5000 years with only smaller oscillations, the variations of which may as well be topographically as climatically controlled; drainage capture of one sector from another during changes in thickness could lead to advances which were not climatically controlled. The extreme variations of the Inland Ice margin position probably did not exceed 20 km.

Attempts to establish a chronology of the glacial events during the last 5000 years are presented by Beschel (1961), Weidick (1968, 1972a) and Ten

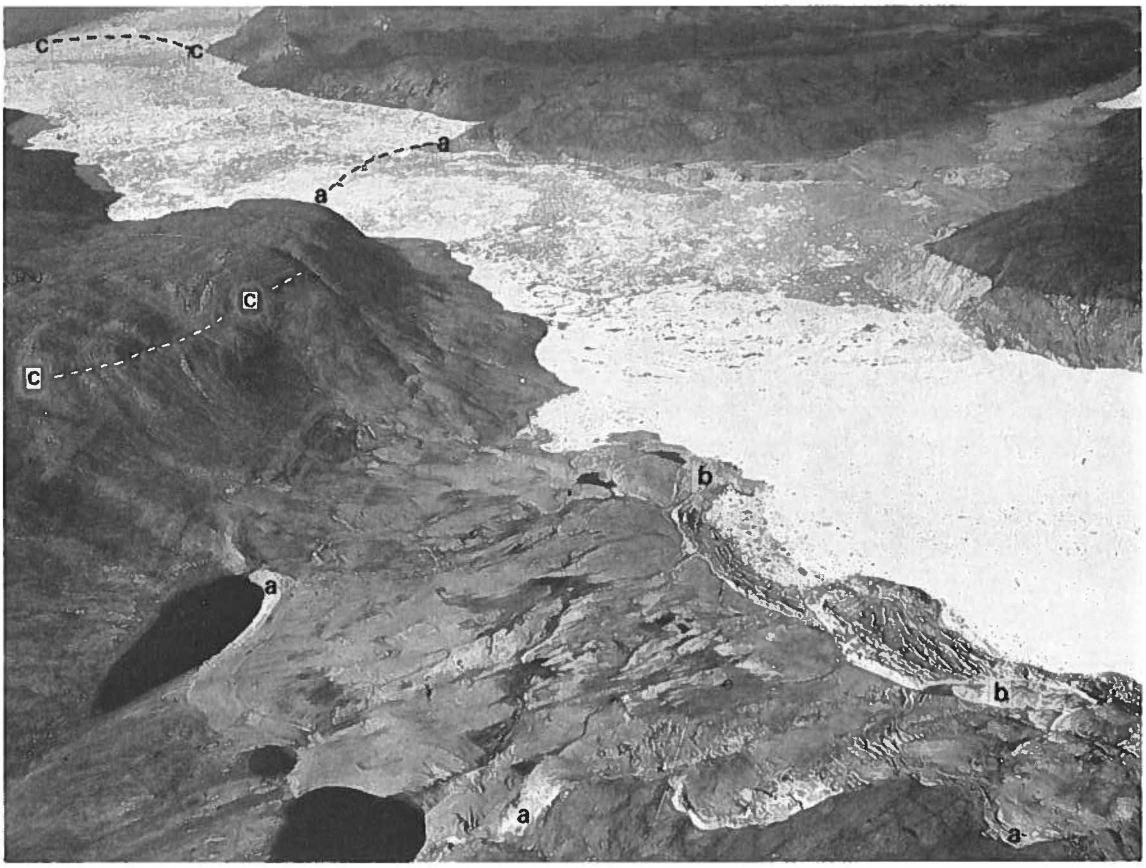


Fig. 372. Moraine terrain at Kangiata nunâta sermia at the head of Godthåbsfjord, West Greenland. In the foreground and nearest the fjord are the trim line zone and moraines from the 18th century (a-a) and from the readvances in 1850–1890? (b-b). Deposits from the fjord stage (8400–8100 B.P.) are seen on the left on the uppermost parts of the hill sides (c-c). Route 506 B-N No. 6270 (21.8.1948). Copyright Geodetic Institute.

Brink (1971), but it must be realised that only the recessional phases during historical time (approximately A.D. 1500–1974) are known in detail. The lobes of the Inland Ice as well as those of local glaciers advanced as a result of the climatic deterioration up to the 17th century (Weidick, 1972a) and maintained, though with minor oscillations, an expanded position until the period 1900–1920. The improvement of the climate in this century with a rise in the annual mean temperature by 2° C (Lysgaard, 1949), caused a mean recession of land-based glacier lobes of around 2 km and of up to 20 km or more for calving lobes (fig. 372).

Holocene deposits

The Holocene deposits of Greenland are found as thin, scattered occurrences of moraine on the pre-Quaternary bedrock, and thicker layers of Holocene sediments mainly in valleys and other topographic depressions. Boulder fields occur frequently on the surface of the high plateaux (peneplains) but little attempt has so far been made to map them or to distinguish them from ground moraine cover.

In restricted parts of Greenland where the ice-free strip is at its widest more extensive ground moraine is found on hilly uplands. Other deposits related to the Inland Ice margin include terminal and marginal moraines and associated limnic and fluviatile deposits in ice margin features, such as ice lake deposits, kame terraces, fluviatile plains and valley trains. Eskers of sorted sediments have been found, but so far these glacio-fluviatile features have only been reported from a few localities in central West Greenland (Weidick, 1971). Fine-grained material is scarce in the moraine matrix and the fluviatile sediments. Even the 'varve clay' from former ice lakes usually consists of silt rather than clay; in fact, clay minerals have only been described from Godthåb (Jensen, 1965).

Marine sediments occur in terraces up to 100 m or more above sea level. Transitions from fluviatile to marine terraces occur frequently and most terraces are built up of marine sand silt, overlain by a thin

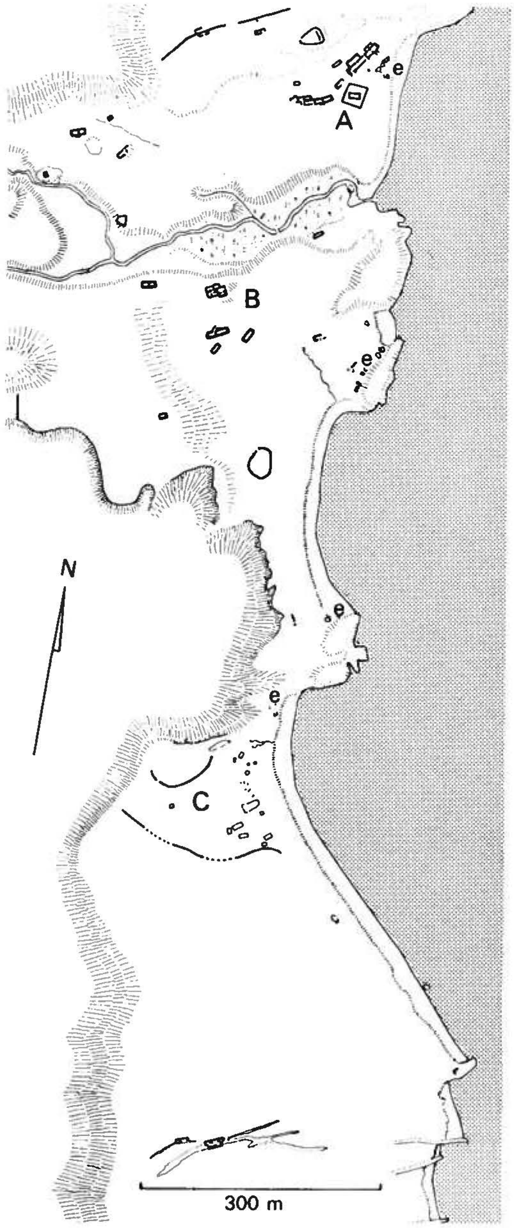


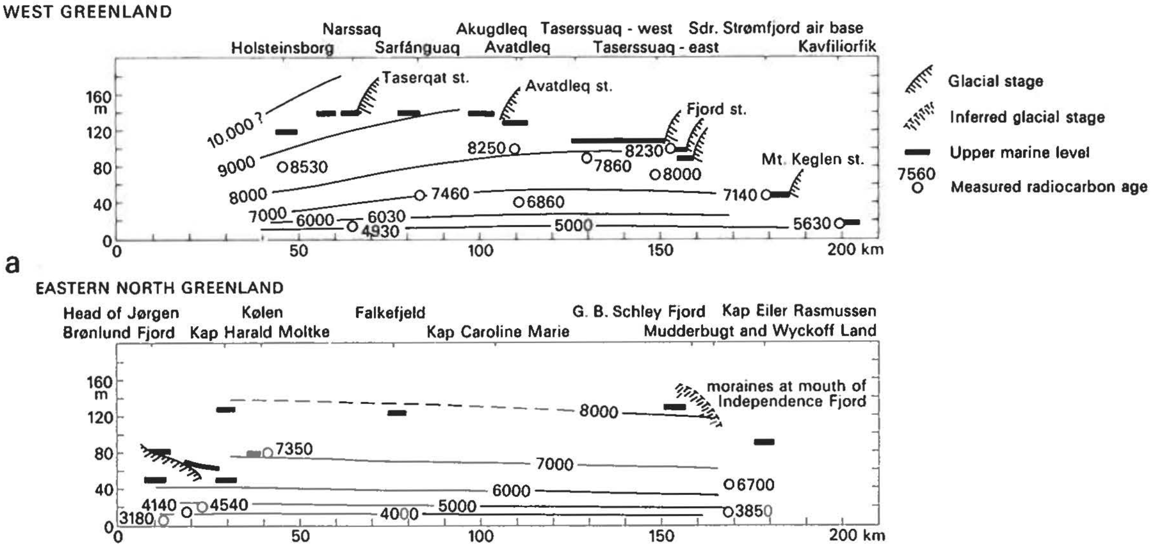
Fig. 373. *Left:* Raised beaches at the village of Qagssiarssuk, South Greenland. The marine limit is here approximately 46 m above sea level and the deglaciation occurred before or at 9000–8500 B.P. (Fredskild, 1973; cf. Table 5). The locality is situated only 11 km from the existing lobe of the Inland Ice at Narssarssuaq airfield. Route 221 A 33 No. 188 (04.08.1958). Copyright Geodetic Institute.

Right: The ruins of the Norsemen's Brattahlid which can be traced on the aerial photograph between the houses of the modern village of Qagssiarssuk. *A* the North farm, *B* the River farm and *C* the Booth place. The place is known to be the locality of the first Norse settlement by Erik the Red in A.D. 985. Eskimo ruins from the time around or after 1700 are marked by *e*. Part of map by Nörlund & Stenberger (1934).

veneer of fluviatile gravel and sand. Beach ridges (fig. 373) with sorted gravel, pebbles and blocks are often found at the outer coast, whereas marine terraces and marine sediments filling valleys are most widespread nearer to the present Inland Ice margin, and occasionally even beneath it.

Aeolian deposits of silt and sand are associated with both fluviatile deposits close to the Inland Ice and to marine deposits closer to the outer coast; in places they cover large areas, though such occurrences are essentially more scattered than other Holocene deposits.

Local glacier deposits are small compared to those from the Inland Ice margin because the local glaciers have not on the whole extended much beyond their present position since the deglaciation of the surrounding country. There are exceptions in southern Greenland where marginal moraines from local



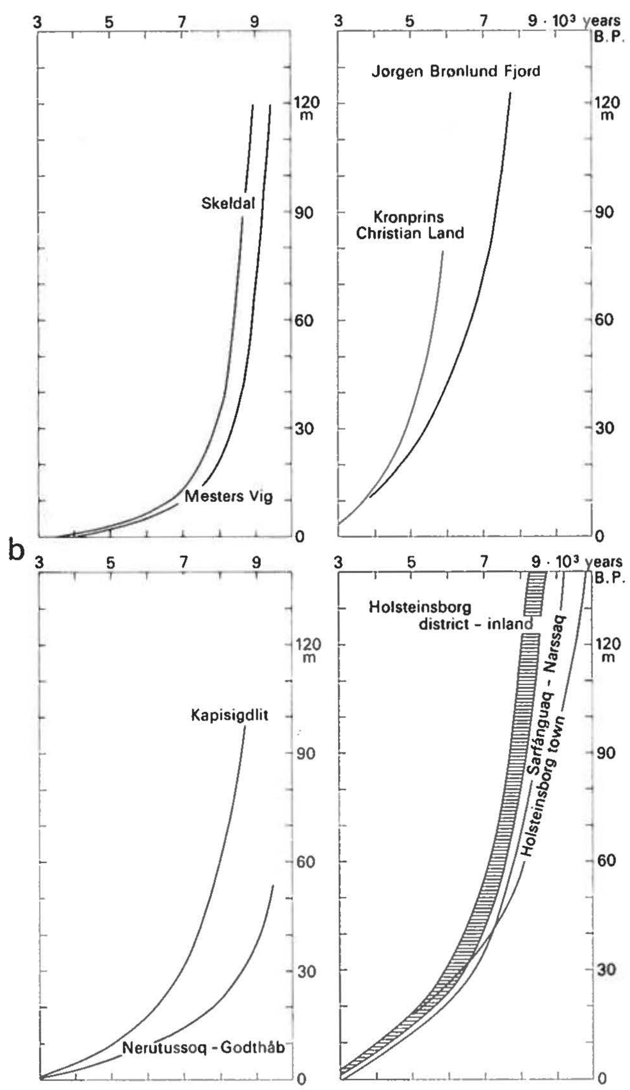


Fig. 374. *a.* Shoreline diagrams from West Greenland (Holsteinsborg – Søndre Strømfjord) and eastern North Greenland (Peary Land; Jørgen Brønlund Fjord – entrance of Independence Fjord). *b.* Emergence curves. Kapisigdlit and Nerutussoq area in the southern part of West Greenland (data from Fredskild, 1973; Kelly, 1974; Weidick, 1972b), Holsteinsborg area of central West Greenland (data from Weidick), Skeldal and Mesters Vig in central East Greenland (data from Lasca, 1969; Washburn & Stuiver, 1962) and Kronprins Christian Land and Jørgen Brønlund Fjord (data from Davies, *in* Weidick, 1972b; Fredskild, 1969; Knuth, 1967).

glaciers indicate that they covered greater areas after the recession of the Inland Ice, and some valleys indicate piedmont glaciation where a total ice cover existed 100–200 years ago but has now disappeared (Weidick, 1968). Esker formation has been observed at several localities associated with the glacier recession in this century.

The slow production of organic material in arctic areas means that soil and peat deposits are small and uncommon; peat layers up to a metre thick occur either connected with guano or human debris (archaeological sites). Turf was used a century ago for heating and as sods for building material.

Sea level changes

Recent sea level changes are closely connected with the isostatic response of the earth's crust to the Holocene deglaciation. The general picture is of an initial fast uplift followed by slower uplift, the rate of which gradually decreased with time. The uplift first began at the outer coast where there was the thinnest ice cover, and the outer coast shows a small earlier uplift compared to the interior parts of the country. Therefore, the old shorelines developed a dome-shaped form, comparable to that recorded in other formerly ice-loaded regions (Donner, 1969; Andrews, 1970). Fig. 374a shows a section through the old shorelines from Holsteinsborg to Søndre Strømfjord air base in central West Greenland. It shows a successively younger age for the upper marine sediments towards the present Inland Ice margin.

The marine limit

Old shorelines and marine sediments indicate the decreasing height of sea level. As shown from the profile of their highest occurrence (fig. 374a) the marine limit is not an isochronous, but a metachronous limit, the height of which seems to increase in accordance with the width of the ice-free strip of land. Thus where the land-strip is 150–200 km wide in West Greenland, the marine limit increases from 110 m at the coast to a maximum of 140 m in the central area. Further south in the Frederikshåb and Julianehåb districts with a land width of scarcely 50 km, the marine limit varies between 40 and 60 m above sea level (fig. 375).

The marine limit in Mesters Vig in East Greenland is reported to be 120 m above sea level (Washburn, 1965) and that of the central part of Scoresby Sund at 120–134 m (Sugden & John, 1965; Funder, 1972). The detailed investigations of Scoresby Sund also reveal an altitude decrease of the marine limit towards the present Inland Ice margin (Funder, 1972) as well as in areas towards the outer coast (Funder & Hjort, 1973); there is a parallel with the West Greenland situation (fig. 374).

In South-East Greenland the land-strip is very narrow and the altitude of the marine limit does not seem to exceed 75 m (Vogt, 1933; Bøgvad, 1940).

In the Thule area of North Greenland the marine limit seems to increase from 26 m at Pitugfik Gletscher near Kap York to more than *c.* 40 m at Saunders Ø (Davies *et al.*, 1963) and to 81 m at the Carey Øer in the waters between Greenland and Canada (Bendix-Almgreen *et al.*, 1967). Further north along the coasts of Canada and Greenland values of 90–100 m are reported (Nichols, 1969; Davies, 1963). This increase of the marine limit from Melville Bugt towards the north may be explained as an effect of the merging Innuitian and Greenland ice sheets over the Kane basin.

In northernmost Greenland, values of up to 130 m are reported for the marine limit of the outer coasts of Peary Land (Troelsen, 1952; Davies, 1963).

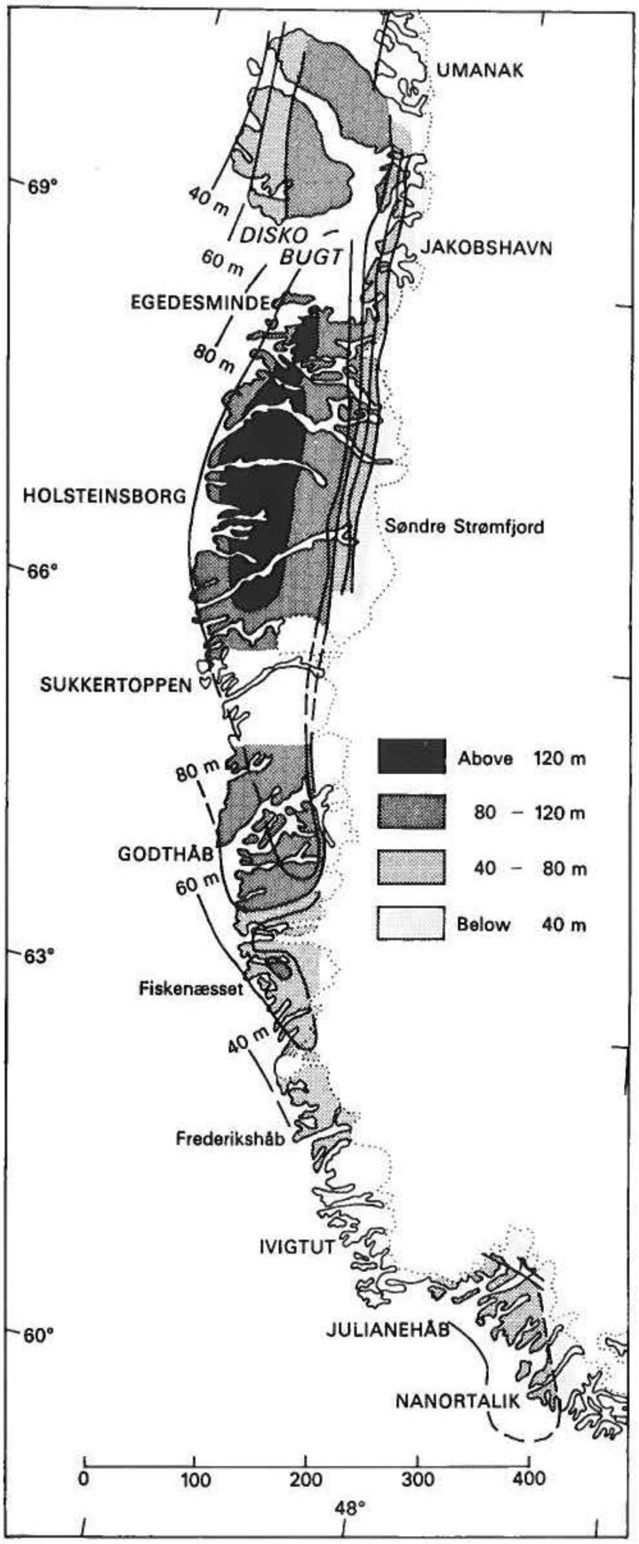
Uplift dates and chronology

In Greenland the glacio-isostatic rate of uplift exceeded the contemporaneous eustatic rise of sea level in the early parts of the Holocene. The decellerating rate of glacio-isostatic uplift and the increased rate of the eustatic rise of sea level during the melting of the Laurentide ice sheet caused long periods of nearly stationary conditions between sea and land in Scandinavia and at times even caused relative submergence. These intervals of stability or transgression resulted in the formation of the marked *Tapes* shorelines in Scandinavia (Andersen, 1965) and possible correlation of this level in Greenland was suggested by Vogt (1933). In Norway, the age of the oldest *Tapes* level is 7860 years B.P., a younger 4820 years B.P. A transgression is stated to have an age of around 5640 B.P. in East Greenland (Hjort, 1973a: the Vega transgression).

In West and East Greenland glacio-isostatic uplift came to an end 4000–5000 years ago, whereas it continued to exceed the eustatic changes in sea level in North Greenland (Weidick, 1972b). The archaeological evidence of the Independence I culture (4100–3700 B.P.) in North Greenland is related to a level 10 m above present sea level and that of Independence II (around 2600 B.P.) to one approximately 4 m above the present (Knuth, 1967). Accordingly, it must be concluded that the period of isostatic uplift in North Greenland first came to an end around 2000 B.P.

More recent secondary transgressions have been demonstrated by ruin sites, now partly drowned. At Qajâ in Disko Bugt (Jakobshavns Isfjord), central West Greenland, the lowermost culture layers (Sarqaq culture? 3500–2700 B.P.) are now below high tide level. A few Norse ruins are now below the present sea level (Roussell, 1941) as shown in fig. 376 and some mediaeval Eskimo ruins in West and East Greenland are also now partly or completely under water (Bøgvad, 1940). The land apparently started to sink between A.D. 1600 and 1700 and has continued to do so until present time (Egedal, 1947; Nielsen, 1952; Saxov, 1958a, 1961). From A.D. 1940 Saxov recorded an uplift. However, all the evidence given for the last millenium points to small sea level fluctuations of scarcely more than 2 m.

Emergence curves (field altitudes of former shorelines plotted against their age, usually in C^{14} years) have now been compiled for several areas in Greenland (fig. 374b). The emergence curves from West Greenland are essentially based on shell dates and in East and North Greenland there are addit-



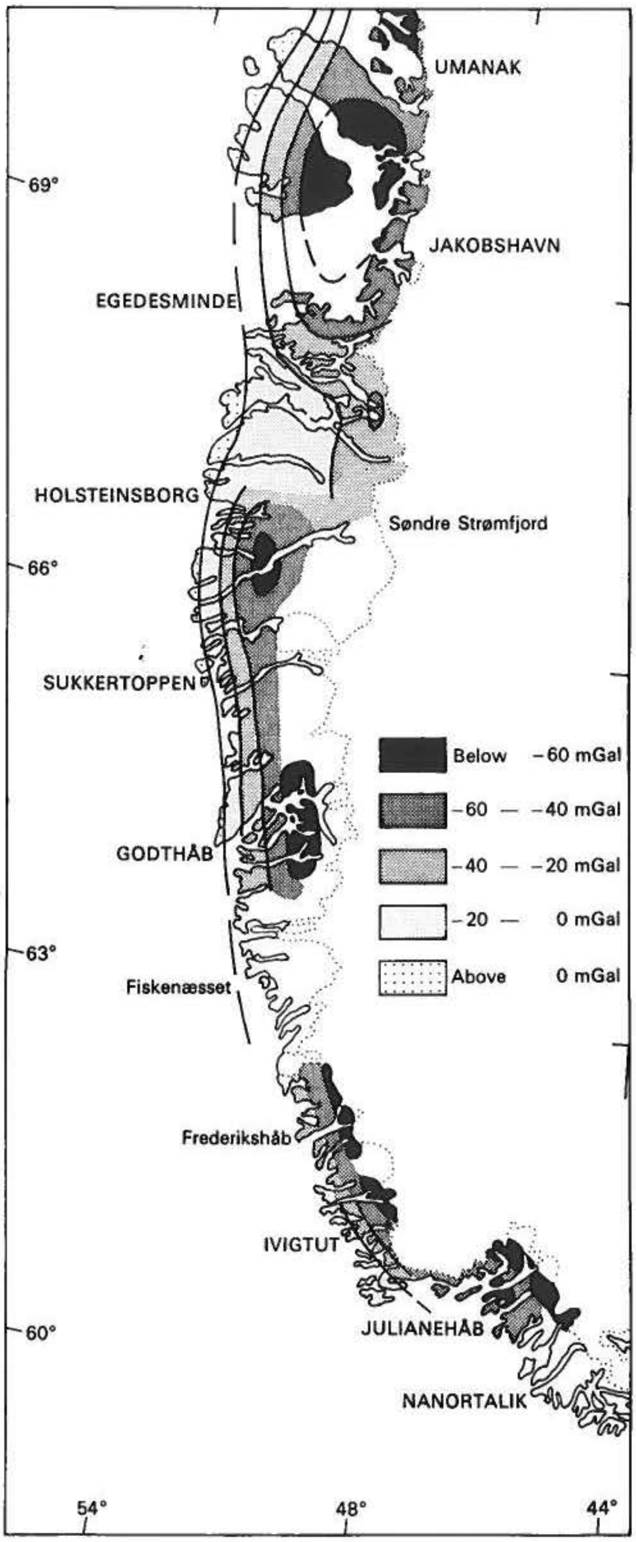


Fig. 375. *Left:* Marine limit in central West Greenland (cf. Fig. 374a; upper marine level).
Right: Bouguer-gravity anomalies in West Greenland (measured close to sea level). Compiled from Kejlsø (1958), Nørgaard (1948), Saxov (1958b) and Svejgaard (1959).

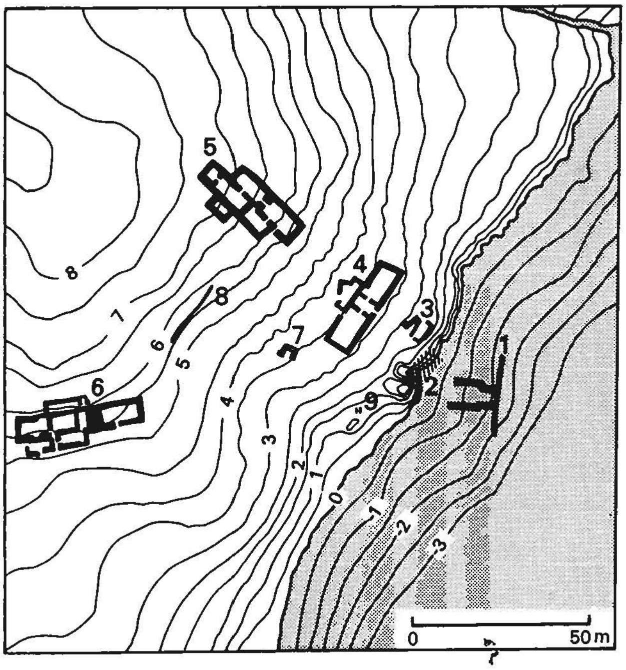


Fig. 376. The Norse settlement Sandnes (now Kilaersarfik) in Ameralik fjord, Godthåb area in central West Greenland. Sandnes was part of the western Norse settlement which, according to the sagas, was abandoned by A.D. 1341. *1* the church; *2* excavated part of the cemetery; *3* earliest site; *4* houses; *5* and *6* stable complexes; *7* smith; *8* dyke (fence); *9* hearth. The 0 m contour line is close to high tide and the tidal range in the area is 3–4 m. Map from Roussell (1941).

ional dates of driftwood. An apparent age of 550 years is used in correction of East Greenland shell dates (Washburn & Stuiver, 1962; Hjort, 1973b) and possibly a correction of 1200 years may be applied on shell material from Peary Land (Weidick, 1972b), whereas such corrections seem unnecessary for West Greenland shell samples. The pattern of this correction points to increasing amounts of upwelling of old sea water towards the north-east along the Greenland coasts, but evidence is still too scarce for final proof.

The glacio-isostatic uplift in Greenland, as in other deglaciated areas, can be approximated by an exponential decay curve such that,

$$U_t = U_0 e^{-kt}$$

where U_0 and U_t are the amounts of uplift remaining to be achieved at zero time and after a subsequent time interval *t*, respectively. In practice, *U* is the present height above sea level of the appropriate shorelines, plus the eustatic correction, plus any future uplift which belongs to this phase of uplift. Several authors using equations derived from this relationship have obtained values for the constant *k*. Ten Brink (1974) was able to assume that uplift in his area was complete and found $k = 0.7 \times 10^{-3}\text{yr}^{-1}$. He considered the geophysical implications of this value and obtained figures for the mantle viscosity and crustal rigidity. Kelly (1973) using different assumptions has suggested *k* values between 0.35 and $0.5 \times 10^{-3}\text{yr}^{-1}$. Analyses of this type are strictly limited in their scope not only by errors in field observations but particularly by the uncertainties over the Holocene eustatic rise in sea level (e.g. the various sea level curves illustrated in Curray *et al.*, 1970). For this reason Donner & Jungner (1975) have preferred to express the uplift of Disko Bugt by an emergence curve, to which they fitted a simple quadratic equation.

The blocked uplift

It has been stressed above that the trend of glacio-isostatic uplift in Greenland shows patterns similar to those from other formerly glaciated regions. However, as the ice recession in the northern hemisphere was complete with the exception of Greenland, where it stopped at an early phase, certain deviations from the general trend can be pointed out.

In Scandinavia and North America, the marine limit as well as the strandlines rise continuously towards the centre of the glaciation, whereas in Greenland, the maximum altitude of the marine zone seems to occur somewhat away from and parallel to the present ice margin, as shown in fig. 375. Evidently, this is at least in part due to a combination of the steady (or even slightly expanded) load of the Inland Ice during the last 5000–6000 years and to the rigidity of the earth's crust at the present Inland Ice margin.

In Scandinavia and North America, the central parts of the formerly ice loaded regions are still under uplift and a decrease in gravity anomalies from the margin of these areas towards the centre of glaciation today is interpreted as evidence of horizontal movement of the subcrustal masses (Heiskanen & Vening Meinetz, 1958), and the estimated residual uplift still to come has been calculated on this basis. However, in West Greenland the gravity anomalies also show a trend so that the greatest negative anomalies (shown as Bouguer anomalies in fig. 375b) occur close to the highest marine limit, though no isostatic uplift can be shown here in the last 4000 years.

Holocene faunal development in Greenland

Marine faunas

Extensive investigations have been made on the faunas in the West Greenland marine sediments. These studies were initiated by Jensen & Harder (1910) and continued by Laursen (1944, 1950). They reveal a development from a high arctic fauna (*Portlandia arctica, Balanus hameri*) through an arctic fauna (*Mytilus edulis, Pecten* (= *Chlamys*) *islandicus*) to a boreal fauna (*Zirphaea crispata, Cyprina islandica*). The boreal fauna characterises the climatic optimum and was followed by the present fauna where the disappearance of *Zirphaea crispata* and *Cyprina islandica* indicates a relatively colder climate.

In East Greenland a former climatic optimum is also indicated by a former more northerly occurrence of *Mytilus edulis* than today (Noe-Nygaard, 1932; Nathorst, 1901). The former extensive occurrence of *Pecten* (= *Chlamys*) *islandicus* is also indicative of a Holocene climatic optimum (Funder, 1972).

Recent radiocarbon dates on West Greenland shell material (Laursen, 1972) show that the marine faunas may indicate additional Holocene climatic fluctuations, presumably of local character, or perhaps related to facies changes in connection with the uplift of land and the related change of currents in fjords and at the coast.

With respect to marine mammals, radiocarbon dates on the Greenland whale (*Balaena mysticetus*) from a raised marine delta 3.6 m beneath a prominent 17 m marine terrace at Dundas, North Greenland, have given an age of 8500 ± 200 years B.P. (Davies *et al.*, 1963). In Holsteinsborg, a whale cranium gave an age of 5845 ± 115 B.P. (Weidick, 1973); however, this sample was embedded in a beach ridge 63 m above sea level, the shells of which gave an age of about 7000–8000 years B.P.

Other information about more recent fauna is discussed below as game associated with the human cultures.

Limnic faunas

Crustacea and Bryozoa in limnic deposits also reflect climatic change. Thus *Lepidurus arcticus* occurs in postglacial layers in South Greenland (60°–61° N), whereas its present distribution is between 66° and 78° N. The bryozoan *Cristatella mucedo* occurs in South Greenland samples from 3600–2600 B.P. indicating a climatic optimum and it is not known from Greenland today. Specific investigations on the development of the crustacean faunas are being undertaken in conjunction with the investigation of pollen from lakes (Fredskild *et al.*, 1975).

Archaeology

The human settlement of Greenland is characterised by the cultures shown in Table 21. The table gives a very simplified picture of the Eskimo immigration which occurred in waves from the Canadian archipelago over the straits between Thule and Inglefield Land into North and West Greenland (or both) and the dates of which are furnished by Larsen & Meldgaard (1958), Knuth (1967) and Fredskild (1967). A review of the investigations of the cultures is given by Gad (1970).

The people of the Independence I culture were musk-ox hunters whereas the primary game for the Sarqaq, Independence II, Dorset and Thule peoples was sea animals (seal, and also whale for the Thule people) and reindeer. The dog was introduced into

Table 21. Eskimo cultures in Greenland

West Greenland	East Greenland	North Greenland
Thule-Inugsuk (from 950 B.P.)	Thule-Inugsuk (from 550 B.P.)	Thule (from 950 B.P.)
	mixed Dorset-Thule (950–750 B.P.)	Dorset
Dorset (2000–1100 B.P.)	Dorset (from south-west Greenland)	Oldest Dorset = Independence II (around 2600 B.P.)
Sarqaq (3500–2700 B.P.)	Sarqaq (from south-west Greenland)	
		Independence I (4100–3700 B.P.)

Table 22. Palynological events in Greenland

South Greenland				Inner West Greenland		Geomorphic events	Jameson Land East Greenland
Kap Farvel		Qagssiarssuk		Kapisigdlit			
2200–present	*Empetrum–Betula glandulosa* phase. Decline in *Juniperus* Climatic conditions become more humid.	1000–present	*Rumex acetosella* phase. Moist followed by dryer period. Vegetational change marked by Norse settling.	1900?–present	*Betula–Alnus–Ericales–Gramineae* phase. Decrease in *Alnus* and increase in Ericales presumed sign of climatic deterioration. Norse settling influenced the vegetation slightly 1000–700 B.P. More moist conditions occur at about 660.	0 Sub-atlantic Stage	Open *Salix arctica* heath or moving soil with *Oxyria digyna* and *Salix arctica*. Cold (humid?) climate.
		1900–1000	*Betula glandulosa–Salix* phase. Decrease of *Juniperus* around 1900. *Betula–Salix* heaths. Continued climatic deterioration; presumably lower precip.				
						2300	
3800–2200	*Betula–Empetrum* phase. End of hypsithermal. Immigration of *Betula glandulosa*. Rather warm-dry climate.	3600–1900	*Betula pubescens–B. glandulosa–Salix* phase. At the beginning of the phase, climatic optimum. Climate possibly more humid.	4000–1900?	*Betula nana–Alnus crispa* phase. Shrubs and dwarf-shrub heaths dominate. *Alnus* and Ericales partly substitute *Juniperus* and *Betula* of the preceding phase: more humid conditions. Climatic temp. max.	Subboreal Stage	Dwarf-shrub heath with *Salix arctica* and *Betula nana* Warm-cold climate (dry?)
						5300	
5300–3800	*Salix–Juniperus–Empetrum* phase. Immigration of *Isoetes lacustris*. The warmest, least oceanic climate in the sequence. Increasing summer temperature.	4500–3600	*Juniperus–Betula glandulosa* phase. Increasing temperature, dry climate. *Juniperus* increase at 5300 possibly marks beginning of climatic change.	7000?–4000	*Betula nana–Juniperus* phase. *Betula nana* dwarf-shrub heaths with *Juniperus* and *Salix glauca*. Climate dry and almost as warm as present.	Atlantic Stage	Rich dwarf-shrub heath with *Betula nana* (and *Thalictrum alpinum*). Climatic optimum (humid?)
(7500)7200–5300	*Salix–Gramineae* phase. Immigration of *Salix glauca*. Beginning of hypsithermal.	6900–4500	*Juniperus–Gramineae* phase. *J.* immigrates around 6900; *Selaginella* emerges 200 years later. Climatic equilibrium for *c.* 500 y.				
8400–(7500)7200	*Sedum–Lycopodiaceae* phase. Increasing dwarf-shrub vegetation. Moist conditions, snow patches.	8000–6900	*Salix–Thalictrum* phase. *Salix* immigrates at 8000 and spreads to present extent immediately.	8000?–7000	*Cyperaceae–Salix–Thalictrum* phase. Dry soil plants as *Artemisia, Rumex acetosella, Gentiana nivalis* and *Campanula* indicate a continental climate. Maximum *Sphagnum* may indicate response to increasing temperature rather than humidity.		Immigration of *Betula nana* around 8000 B.P.
						8200	
9100–8400	*Oxyria–Loiseleuria* phase. *Empetrum, Vaccinium* immigration. Ericaceous dwarf shrubs. Open vegetation with snow-patches.	8700–8000	*Cyperaceae–Plantago maritima* phase. Pioneer vegetation; all the species found occur in present flora.	8700?–8000	*Gramineae–Plantago maritima–Empetrum* phase. Pioneer vegetation. Climate presumed to be like present in Disko Bugt.	Boreal Stage	Pioneer vegetation (grasses) warm, dry climate.
9600–9100	*Oxyria–Koenigia* phase. No indication of plants demanding warmth; supposedly relatively cold.			Source: Fredskild (1973)			Source: Funder (1971)

Greenland with the Dorset people if not earlier.

Europeans of the Norse culture settled the Julianehåb and Godthåb districts of South and southern West Greenland between A.D. 985 and approximately 1500 (figs 373, 376). They introduced many domestic animals such as horses, cattle and sheep.

The disappearance of both the early Eskimo and the Norse cultures is often explained by climatic fluctuations. However, overhunting of game, and for the Norse settlers, the overgrazing and political changes in Europe may also have contributed to their extinction.

Vegetational history

The best guide to the vegetational history of Greenland is found in pollen which often survives in deposits long after other plant material has decomposed. Most investigations of pollen have been carried out in connection with archaeological research and therefore only cover the last millenia.

The first data covering large parts of the Holocene was given by Iversen (1953) from lake cores 100, 50 and 8 m above sea level from the inner parts of Godthåbsfjord, and they presumably cover the last 8500 years. The sites were re-investigated by Fredskild (1967, 1973) and the data correlated with information from cores in South Greenland (Kap Farvel area and Qagssiarssuk).

According to Fredskild (Table 22), the first vegetation at the outer coast at Kap Farvel in South Greenland developed at or shortly after 10 000 years B.P. and can be described as pioneer vegetation with arctic plants confined to snow patches; this persisted until 7200 B.P. This might indicate that the glaciation limit was even lower than today in this area. In the interior parts of South Greenland (Qagssiarssuk) and the Godthåb district (Kapisigdlit, Godthåbsfjord) the vegetation arrived later (8500–8000 B.P.) following the receding Inland Ice. The subsequent development of vegetation in these two more arid inland areas was parallel. A schematic review of the three sites gives the chronology of Table 22.

The most characteristic feature in the West Greenland sequences is the indication of a postglacial hypsithermal (climatic optimum) as late as subboreal time (5300–2500 B.P.). The same has earlier been stated with respect to Kapisigdlit by Iversen. In North Greenland the climatic optimum began before 6000 B.P. since driftwood dated at 5900 B.P. indicates fjords were free of glacier lobes (Fredskild, 1973), and in East Greenland investigations on Jameson Land reveal an immigration of the dwarf birch (*Betula nana*) at 8000 years B.P. (Funder, 1971). Funder is inclined to believe that the climatic optimum in Jameson Land as well as in Kapisigdlit occurred in Atlantic time (7500–5300 B.P.). The discrepancy between these viewpoints is essentially due to differing opinions on the delay of spreading of the species, but they agree that the subboreal stage marks the end of the hypsithermal period and that the subatlantic stage is characterised by climatic deterioration. Possibly the climatic optimum includes the whole of Atlantic as well as subboreal time (Kelly & Funder, 1974).

There are great difficulties in correlating dry and humid periods at the present stage of investigation, and this is even true of the secondary climatic fluctuations in Greenland of the last 2000–3000 years about which much more is known (Fredskild, 1973).

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