The glacial history of the Hans Tausen Iskappe and the last glaciation of Peary Land, North Greenland

By Jon Y. Landvik, Anker Weidick and Anette Hansen

Abstract

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Studies of glacial geology near the Hans Tausen Iskappe, 82°N, suggest that the Greenland Ice Sheet covered Peary Land during the Late Weichselian. The ice sheet experienced successive marginal retreat due to calving along the major fjords, and an extensive thinning in response to the climatic amelioration during the early Holocene. Local ice caps had melted by ca. 8100 cal years ago, a result which is compatible with the studies of an ice core record from the Hans Tausen Iskappe. The present ice caps are thus not relicts from the Weichselian, but have formed during the Holocene. Findings of 5000-6000 year-old driftwood along fjords that connect to the Arctic Ocean indicate that the perennial sea-ice cover was gone during the mid-Holocene.

Keywords: Greenland; glacial history; ice caps; Weichselian; Holocene; climate.

Jon Y. Landvik, The University Courses on Svalbard (UNIS), P.O. Box 156, N-9171 Longyearbyen, Norway (Present address: Agricultural University of Norway, Department of Soil and Water Sciences, P.O. Box 5028, N-1432 Ås, Norway.

Anker Weidick, Geological Survey of Denmark and Greenland (GEUS), Thoravej 8, DK-2000 Copenhagen NV, Denmark.

Anette Hansen, COWI, Thulebakken 34, DK-9000 Aalborg, Denmark.

Introduction

During recent decades, investigations of the Quaternary history of the land areas surrounding the North Atlantic have demonstrated a close coupling between the oceanic conditions and terrestrial climate. The Greenland ice core studies as well as stratigraphic studies (Johnsen *et al.* 1992, Dansgaard *et al.* 1993) in the ice-free areas of Greenland and Svalbard (Funder *et al.* 1998, Mangerud *et al.* 1998) have shown a correlation with the paleoceanographic records of the North Atlantic (Bond *et al.* 1993). Even though our present understanding of the climatic interaction between the North Atlantic and the adjacent land areas has increased, more needs to be learned about the parts of the Arctic that are closer to sources of moisture such as the Arctic Ocean.

This study is a part of the Hans Tausen Iskappe project which was designed to investigate the present and

Fig. 1. Location map.

past climate and glacier dynamics of North Greenland by means of ice-core records (Hammer *et al.* 2001), mass balance studies and the local glacial history on a geologic time scale (this paper).

The research strategy for the geological investigations has been to search for stratigraphic and geomorphologic evidence for fluctuations in volume and extent of the Hans Tausen Iskappe. The northern and northwestern margins of the ice cap were considered as most suitable for new field investigations. In this area, the outlet glacier and associated deposits can be found below the Holocene marine limits. Due to this interaction with Late Quaternary fjord environments, marine fossils enable radiocarbon age control on glacial fluctuations. The paper will also review and discuss previous investigations and unpublished observations from the southern margin of the ice cap and the major fjords systems such as Independence Fjord and J. P. Koch Fjord (Fig. 1).

Methods

The new field studies were based on air photo interpretations succeeded by field investigations in three selected areas, supported by helicopter reconnaissance during camp moves.

The geographic sample positions were determined by use of a GPS receiver, using the WGS-84 datum. Altimetry was

Sample	Lab ref.	Location	Co-ordinates	Elevation	Material	^{14}C Reservoir corrected	Conventional ^{14}C age	Calibrated age	Remarks
Northern margin:									
GGU 417602	T-11769	Nordpasset E	82°53.7'N 36°09.2'W	32	Mt	7260±55	7810±55	8120	In growth position above till.
GGU 417611	T-11773	Nordpasset E	82°54.9'N 36°23.4'W	37	Mt	7245±95	7795±95	8100	Growth position in marine sed.
GGU 417601	T-11768	Nordpasset E	82°53.8'N 36°08.6'W	49	Ha, Mt	7085±95	7635±95	7930	On delta surface
GGU 417609	T-11772	Nordpasset E	82°54.7'N 36°10.5'W	43	Ha, Mt	6910±140	7460±140	7750	Raised shoreline
GGU 417606	T-11771	Nordpasset E	82°54.0'N 36°11.1'W	18	Ha	6450±120	7000±120	7370	Frozen up on delta front
GGU 417605	T-11770	Nordpasset E	82°53.7'N 36°09.1'W	24	*Picea* sp.	–	5855±50	6710 or 6670	
GGU 417620	T-11776	Nordpasset W	82°58.0'N 37°41.3'W	28	Shell frag.	6945±140	7495±140	7790	In redeposited silt.
GGU 417621	TUa-1079	Nordpasset W	82°58.2'N 37°43.9'W	29	Ha	6680±70	7230±70	7560	AMS-date of single valve
GGU 417612	T-11774	Kap Bopa	82°59.4'N 39°40.1'W	33	Ha, Mt	7140±85	7690±85	7970	
GGU 417614	T-11775	Kap Bopa	82°59.4'N 36°40.6'W	13	*Larix?*	-	5470±100	6280	
GGU 417613	TUa-1078	Kap Bopa	82°59.3'N 39°40.2'W	46	Shell frag.	36,870±605	36,980±605		
Southwestern margin:									
GGU 226451	I-9664	E. of J.P. Koch Fj.	82°24'N 40°30'W	30-35	Shells	5550±110	5700±105	6350	Weidick (1977)
GGU 226450	Ua-4586	E. of J.P. Koch Fj.	82°17'N 39°55'W	Ca 25	Ha	5085±60	5635±60	5880	
GGU 226451	Ua-4587	E of J.P. Koch Fj.	82°16'20''N 39°59'W	30-36	Ha	5480±80	6030±80	6280	
GGU 215308	GSC-2279	Adams Gletcher	82°21.2'N 40°51'W	2-3	Shells	4040±150	4590±140	4590	Weidick (1977)
GGU 215308	I-9130	Adams Gletcher	82°21.2'N 40°51'W		Twigs	-	220		Weidick (1977)

Table 1. Radiocarbon dates from the Hans Tausen Iskappe area. The conventional radiocarbon age has been normalised to $\delta^{13}C = -25$ ‰. The dates from Weidick (1977) have been normalised by the authors. All dates on marine material have been corrected for a reservoir age effect of –550 yrs (Funder 1982). A recent survey suggests that the reservoir age could be up to 760 yrs. in North Greenland (Rasmussen & Rahbek). Prefixes TUa- and Ua- are AMS-dates. Calibration to calendar years is obtained using the CALIB programme (Stuiver & Reimer 1993). Ha = Hiatella arctica, Mt = Mya truncata.

carried out through parallel use of two electronic altimeters, and readings were corrected for air pressure changes during the day. The elevations refer to the high tide mark. The tidal range in the area appeared to be small, and in Jørgen Brønlund Fjord (Fig. 1) it is reported to be only 20-30 cm (Høy 1970).

For comparison with the ice core record, all radiocarbon ages (Table 1) have been converted to calendar years according the calibration data-sets by

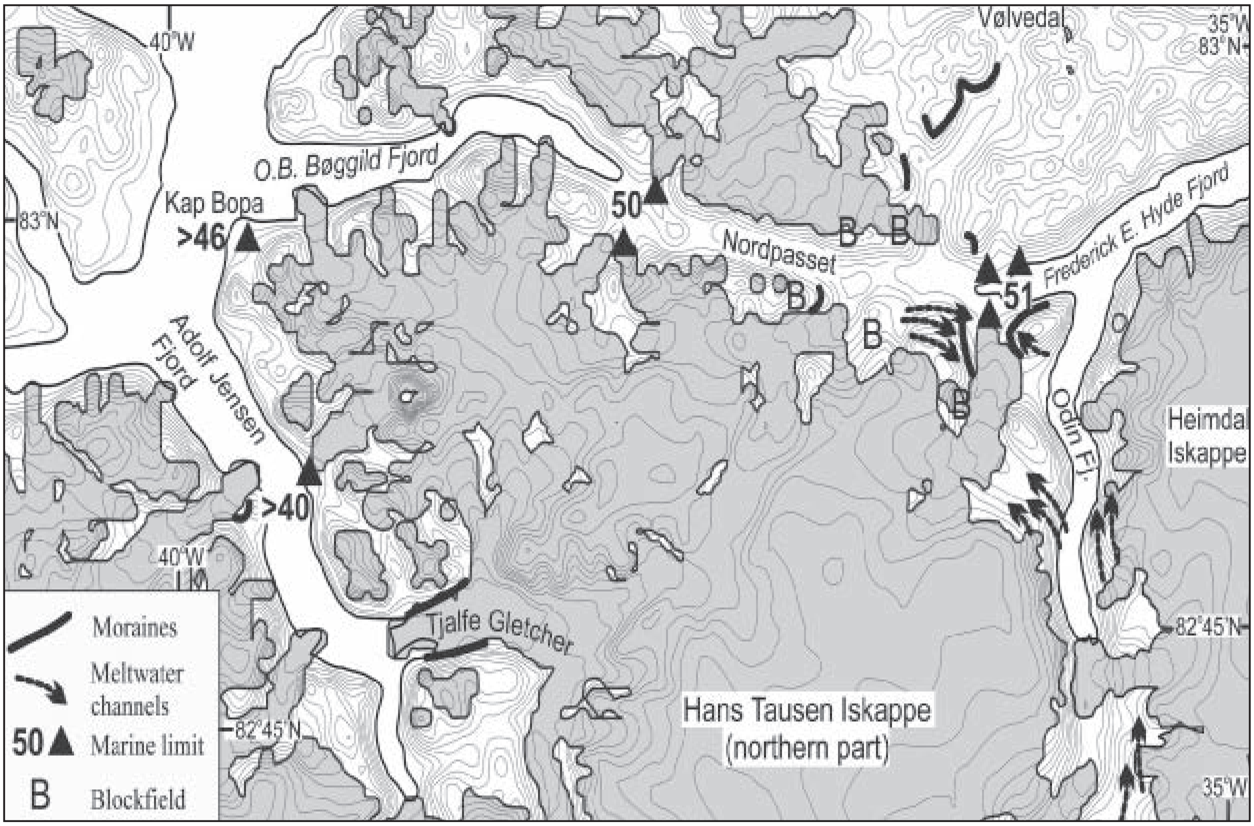

Fig. 2. The glacial geology along the northern margins of Hans Tausen Iskappe and Nordpasset. Moraines and melt-water channels have been mapped in the field. Black triangles: deltas; numerals: elevation of marine limit in metres. The map is extracted from a digital terrain model constructed by Wolfgang Starzer, GEUS.

Stuiver *et al.* (1998a) and Stuiver *et al.* (1998b) using the CALIB 4.1 calibration programme (Stuiver & Reimer 1993).

Physiography

Hans Tausen Iskappe is the second largest ice cap in North Greenland, covering an area of 4208 km^2. It is not only one of the northernmost ice caps in the world, but also one of the few that is located close to the Arctic Ocean. Most of the present day ice cap has an elevation exceeding 1000 m a.s.l. Its southern parts rest on a 800-900 m high plateau, whereas the northern part of the ice cap sits on more rugged bedrock (Reeh 1995, Reeh *et al.* 2001). To the south, the Wandel Dal valley, connecting J. P. Koch Fjord to the west with Independence Fjord to the east, separates the ice cap from the Greenland Ice Sheet. The southern margin of the ice cap (Fig. 1) terminates on dry land at ca. 500 m a.s.l. To the north, however, outlet glaciers reach elevations of ca 100 m a.s.l. in Nordpasset and terminate as tidewater glaciers at the head of Adolf Jensen Fjord to the north-west (Fig. 2).

Nordpasset is a 2-3 km-wide and 25 km long u-shaped, probably glacially eroded valley that connects Frederick E. Hyde Fjord to O.B. Bøggild Fjord. The central part of the pass is 150-200 m a.s.l. and mountains to the north and south reach elevations up to 800 m a.s.l. (Fig. 2). Rock glaciers are developed along the mountain slopes in the central part of the valley, and the valley floor is characterised by vigorous periglacial activity and cryoturbation of the Quaternary deposits. Outlet glaciers from the Hans Tausen Iskappe reach Nordpasset in its eastern parts.

The bedrock of North Greenland is dominated by Late Precambrian to Palaeozoic sedimentary rocks that become successively younger towards the north. They are generally found in east-west trending zones related to the formation of the North Greenland fold belt in the Devonian/Early Carboniferous (Henriksen 1992). The Greenland Ice Sheet probably covers the crystalline basement. The finding of erratics of high-grade metamorphic and plutonic origin along the north coast of Peary Land points to sources under the present ice sheet (Dawes 1986).

The present glaciation limit rises from ca. 200 m along the coast to ca. 800 m a.s.l. in the Hans Tausen Iskappe area (Weidick 2001).

The North Side of Hans Tausen Iskappe

The studies along the northern margin of the Hans Tausen Iskappe were based on air photo interpretation and helicopter reconnaissance, supported by field work in both ends of Nordpasset, and at Kap Bopa where O.B. Bøggild Fjord and Adolf Jensen Fjord meet (Fig. 2). The observations will be discussed from east to west.

Odin Fjord

Odin Fjord cuts the eastern part of the Hans Tausen plateau (Fig. 2). Its southern part is filled by two outlets from the surrounding ice caps: a northern outlet from the Heimdal Iskappe east of Odin Fjord (2KH01013 in Weidick 2001), and a southern outlet called Ymer Gletcher from the Hans Tausen Iskappe west of the fjord (2KH01034 in Weidick 2001). The rims of the plateau on both sides of Odin Fjord and Ymer Gletcher are ca. 600 m a.s.l. These plateaus, particularly east of the fjord, are all characterised by glacial and glaciofluvial sediments and a series of well-developed lateral meltwater channels that were formed by water draining towards the north. The vertical distribution of these channels suggests erosion along the margin of a successively lower fjord glacier in Odin Fjord. A suite of channels drain an over 500 m-high pass towards the northwest, which are today cross-cut by the north-eastward outlet of the Hans Tausen Iskappe (Fig. 2). This relationship shows that the ice cap was less extensive at the time when the meltwater channels were formed.

Eastern Nordpasset

The eastern part of Nordpasset is a key area for the reconstruction of the glacial history of the Hans Tausen Iskappe. The lower parts of the mountain slopes are generally covered by a diamicton of glacial origin. However, they have been subject to extensive downslope movement by solifluction and only a few distinct moraine ridges are found above the marine limit (Fig. 2). Large glaciofluvial deltas, partly covered with beach sediments, are found in front of major meltwater pathways, whereas cryoturbated pebbly silt deposits, interpreted to be of glaciomarine and littoral origin, cover the valley floor below the marine limit.

As for the Odin Fjord area, the eastern part of Nordpasset shows evidence of downwasting of a former glacier. Three sets of geomorphological features are important for the interpretation of the glacial history: a) glacial meltwater channels; b) moraine ridges; c) glaciofluvial deltas.

Meltwater channels

Two transects of the southern slope of Nordpasset were studied in detail. The area between ca. 900 and 600 m a.s.l. is dominated by blocks and a silty matrix derived from weathering of the underlying siltstone. Patches of sub-rounded gravel are found locally in the blockfield. In a horizontal zone at 600 m a.s.l., no blocks are found and only *in situ* weathered bedrock is exposed, probably due to removal of the blocks by glacial

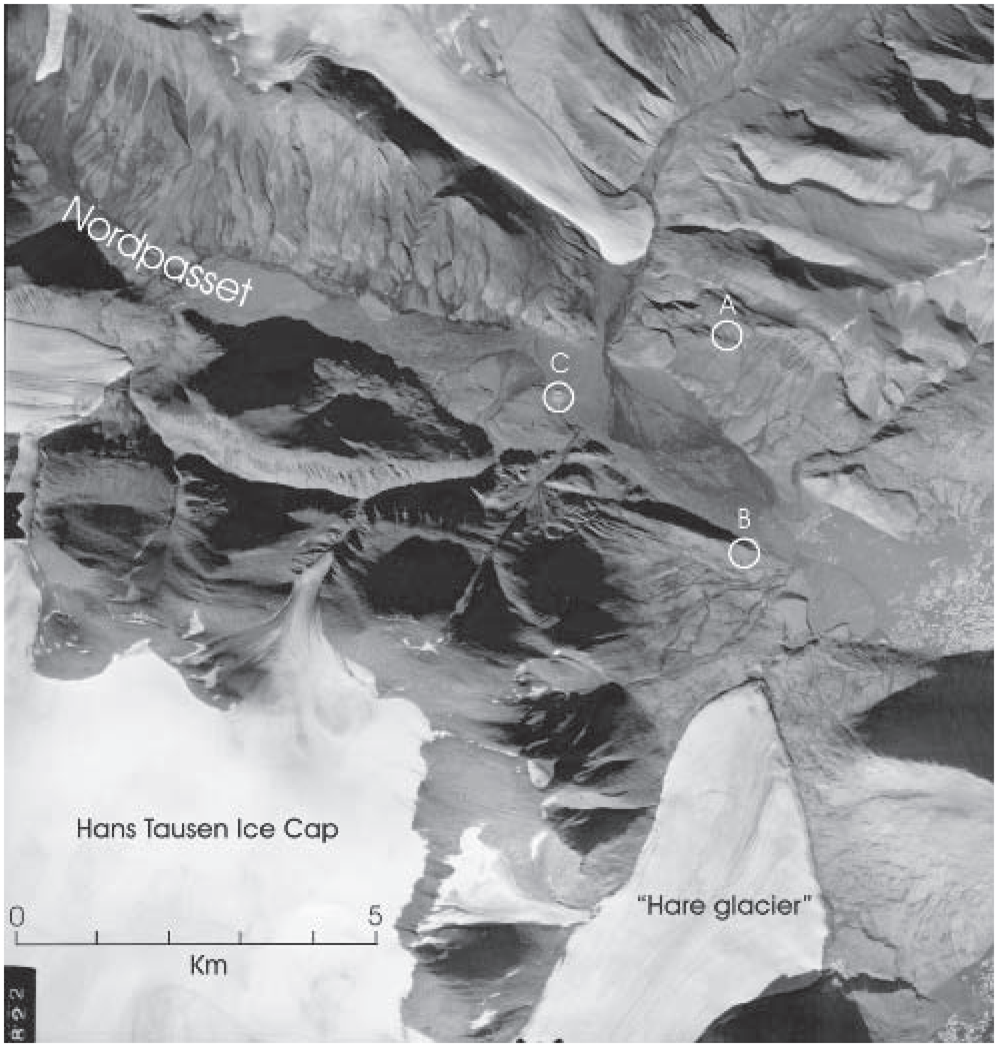

Fig. 3. Vertical air photo of the eastern part of Nordpasset. Localities discussed in the text are marked A to C. Aerial photograph 255K 822, July 1960. Copyright: Kort- og Matrikelstyrelsen, Copenhagen.

meltwater. Below this level, there is a series of distinct ice marginal features, predominately meltwater channels cut into bedrock at successively lower altitudes (Figs. 2 and 3). Several of the channels are 20-30 m deep, dip towards the east, and begin and end in open air (Fig. 4). Their formation requires that meltwater from a valley glacier in Nordpasset turned into ice marginal drainage before it re-entered the glacier either supra- or subglacially.

Moraines

Evidence that a thick glacier filled Nordpasset is also found on the steep northern side of the valley. On a ledge 405 m a.s.l. (Fig. 3, A), there is a ca. 50 m-long moraine ridge comprised of a diamicton with subrounded to rounded boulders in a silty sandy matrix. The ridge has a sharp crest and a fresh-looking appearance (Fig. 5). The lobate shape indicates deposition from a glacier flowing eastwards through Nordpasset, compatible with the dip of the meltwater channels on the south side.

A set of more continuous moraine ridges formed during a later stage of the Nordpasset glacier can be found below ca. 300 m a.s.l. At the eastern end of Nordpasset, a 14 km-long valley connects to the wide east-west running Vølvedal valley to the north. On the western slope of the connecting valley, a

Fig. 4. One of the meltwater channels in the southern slope of Nordpasset.

moraine ridge with declining elevation to the north can be mapped over a distance of 6 to 8 km (Fig. 2). The elevation of the moraine is >600 m a.s.l. in the southern part of the valley, and the northward dip of the moraine shows that it was deposited from a glacier that once filled Nordpasset. The slopes and valley floor below the moraines are characterised by glacial and glaciofluvial deposits.

Another set of moraine ridges is found south of Harebugt at the head of Frederick E. Hyde Fjord. A moraine lobe deposited from the 200 m-high pass between Odin Fjord and Nordpasset crosscuts the lowermost meltwater channels discussed above (Fig. 2). As we infer that no ice existed on the Hans Tausen plateau at his time, it shows that an outlet glacier from Odin Fjord entered the eastern part of Nordpasset during a late stage of the last deglaciation.

Glacial erratics (Fig. 6) are found at

Fig. 5. Moraine ridge on a ledge 450 m a.s.l. on northern side of Nordpasset.

Fig. 6. Erratic boulders 470 m a.s.l. on the mountains south of Nordpasset.

several elevations on the mountains surrounding Nordpasset. The lithologies have not been traced to their area of origin, but the erratics demonstrate the overriding by a glacier prior to the formation of the meltwater channels.

Raised deltas and the marine limit

The marine limit in the area was formed during the last deglaciation. In eastern Nordpasset it can be determined by the elevation of three glaciofluvial deltas (Figs. 2 and 3). The largest one is located 1 km north of the present snout of the outlet from Hans Tausen Iskappe (informally named Hare Glacier by Reeh *et al.* 2001, 2KG01002 by Weidick 2001). The delta surface is ca. 1000 m wide, has its apex close to the present meltwater river from the glacier, and slopes gently with a distinct break towards the delta front. The raised delta is cut by the present river that forms a large modern delta at sea level (Figs. 2 and 3). The distal break of the delta plain is 49 m a.s.l. which is assumed to have formed a few metres below sea level during deposition. Shell fragments of *Hiatella arctica* and *Mya truncata* that were brought to the surface by cryoturbation at the outer part of the delta plain were radiocarbon dated to 7085±95 BP (7930 cal. yrs. BP, T-11768). In a 5 m-high section in the delta front, glaciomarine silty sand is observed above a diamicton interpreted as till. Shells of *Mya truncata* in living position 1.75 m above the diamicton were dated to 7260±55 BP (8120 cal. yrs. BP, T-11769).

A series of raised shorelines where the uppermost beach deposit was hand-leveled to 51 m a.s.l. is found ca 1000 m north-west of the delta (Fig. 3, B). The beach sediments are overrun in places by solifluction deposits suggesting that they represent a minimum estimate for the elevation of the marine limit. When compared to the elevation of the delta plain, however, 51 m seems to be a fair determination of the marine limit in the area. The uppermost occurrence of shell fragments in the beach sediments at 43 m a.s.l is dated to 6910±140 BP (7750 cal. yrs. BP, T-11772).

On the north side of the valley are two smaller raised deltas (Fig. 2). The western one has a steeper fan-delta like surface with a front break at 41 m a.s.l., whereas the eastern one is built up to 50 m a.s.l. Shells brought to the surface by

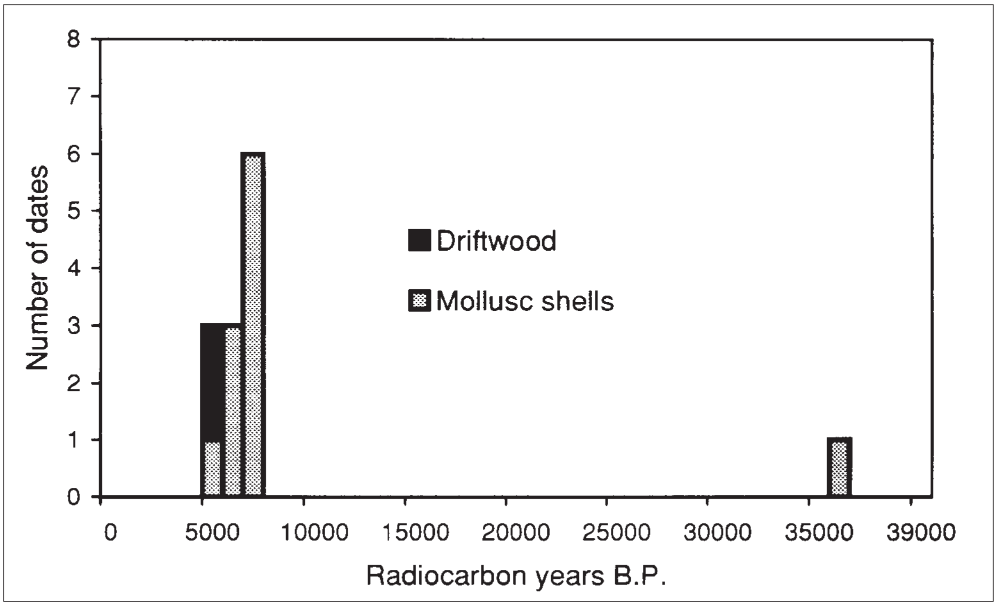

Fig. 7. Distribution of new radiocarbon dates from the Hans Tausen Iskappe area.

cryoturbation in the latter delta front were dated to 6450±120 BP (7370 cal. yrs. BP, T-11771).

Below the marine limit, the valley slopes and floor are covered predominately by solifluction deposits, which comprise a mixture of marine fine-grained sediments, beach gravel and diamicton. In the centre of the valley (Fig. 3, C) there is a mound of marine silt. Redeposited sediments mantle most of the deposit, and the 20 cm-thick active layer only melts the outer part of this mantle. However, the appearance of larger clasts in the lower part of the slopes suggests that the silt deposit is stratigraphically underlain by a diamicton. At the surface, ca. 3 m above the top of the diamicton, numerous valves of *Mya truncata* were found in living position and dated to 7245±95 BP (8100 cal. yrs. BP, T-11773).

Inner Frederick E. Hyde Fjord

As discussed above, the marine limit in western and eastern Nordpasset is 50 and 51 m a.s.l., formed at ca 6900 and 7200 BP (7800 and 8100 cal. yrs. BP), respectively (Fig. 7).

A glaciofluvial delta at the mouth of Vølvedal (83°00’N 34°10’W) is related to the deglaciation of both Vølvedal and Frederick E. Hyde Fjord, and was deposited during a sea level of 53 m a.s.l. However, the exact age of this deposit is not known. A similar height of 55 m a.s.l. was reported by Bennike (1983) from a kame delta at the entrance to the Vølvedal valley. Along the northwestern margin of the Hans Tausen Iskappe, mapping of the marine limit along the eastern coast of Adolf Jensen Fjord gave estimates of >46 m a.s.l.

The drop in the marine limit from 53 m a.s.l. in Frederick E. Hyde Fjord to 51 m in eastern Nordpasset may indicate a slightly earlier deglaciation of the fjord basin, probably due to calving of the glacier. This is also supported by the dip of the moraines from Nordpasset into Vølvedal (see below), which also suggests a deglaciation of the Vølvedal/ Fredrick E. Hyde Fjord valley prior to the final deglaciation of Nordpasset.

Western Nordpasset

Areas below the marine limit are also found in the westernmost 5 km or so of Nordpasset, at the head of O.B. Bøggild Fjord (Fig. 2). The morphology of the glacial deposits is not as well expressed as in the eastern part of the valley, partly due to disturbance by cryogenic processes.

A large raised delta is found at the mouth of a tributary valley entering Nordpasset from the north. The delta is ca. 1 km wide, and reaches halfway into Nordpasset. To the west, it is dissected by river erosion, and terraces are formed at different levels. All terrace surfaces are dissected by up to 1 m-deep ice-wedge polygons. The frontal part of the uppermost terrace surface is slightly offset by a series of front-parallel faults, and the elevation on the ice-proximal side of the innermost fault is 50 m a.s.l. Paired *Hiatella arctica* and *Mya truncata* were found in sediments on the delta front, and a single valve of *Hiatella* was radiocarbon dated to 6680±70 BP (7560 cal. yrs. BP, TUa-1079).

A slightly higher minimum age for the deglaciation is recorded from the southern side of the valley, where shell fragments in a silt subjected to solifluction were dated to 6945±140 BP (7790 cal. yrs. BP, T-11776).

Three km from the coast, a fan delta deposited from Nordpasset sits in the middle of the valley. The lowermost mapable river channels on the fan surface were found at 55 m, an elevation which is compatible with a marine limit of 50 m a.s.l., as discussed above.

Valleys north of Nordpasset

An air photo study and helicopter reconnaissance of the Vølvedal and Nornegæst Dal valleys were carried out in order to track the northward distribution of the glacial deposits mapped in Nordpasset. The moraines and associated glacial sediments that were mapped in the valley connecting Nordpasset and Vølvedal (see above) continue over the 400 m-high pass into Vølvedal. The valley floor of the inner part of Vølvedal is 200-300 m a.s.l. and characterised by thick glacial and glaciofluvial deposits dissected by meltwater channels. Several large meltwater channels are also cut into bedrock, and there has been a stage of meltwater overflow from Nordpasset into Vølvedal over a pass ca. 600 m a.s.l.

Also in Nornegæst Dal north of Vølvedal, a similar distribution of sediments is found. Here, the glacial and glaciofluvial sediments cover the valley floor and continue up along the slopes. However, the upper limits of glacial sediments or any marginal moraines could not be determined during our reconnaissance. Outlet glaciers from Roosevelt Fjelde apparently overrun the glacial deposits along the northern side of the valley. Several large meltwater channels and kame terraces, especially along the southern side of Nornegæst Dal and Frigg Fjord, show that considerable meltwater drainage entered Frederick E. Hyde Fjord through the Frigg Fjord tributary during the last deglaciation.

The South Side of Hans Tausen Iskappe

The present information on the Holocene deglaciation of the region from J.P. Koch Fjord to Independence Fjord is illustrated in the map of Fig. 8. The simplified trend lines are based on detailed mapping of recessional ice marginal features such as moraines, meltwater channels, etc., by W. Davies (in GEUS files, unpublished). A more comprehensive account of these maps is given by Weidick & Dawes (1999). The trend lines indicate a deglaciation pattern in which rapid deglaciation along Independence Fjord and the lowland along Jørgen Brønlund Fjord and Wandel Dal resulted in ice remains over the land areas presently covered by Hans Tausen Iskappe and Heinrich Wild Iskappe.

The chronology of this recessional pattern is fragmentary. In the J. P. Koch Fjord basin, the recession of the Green-

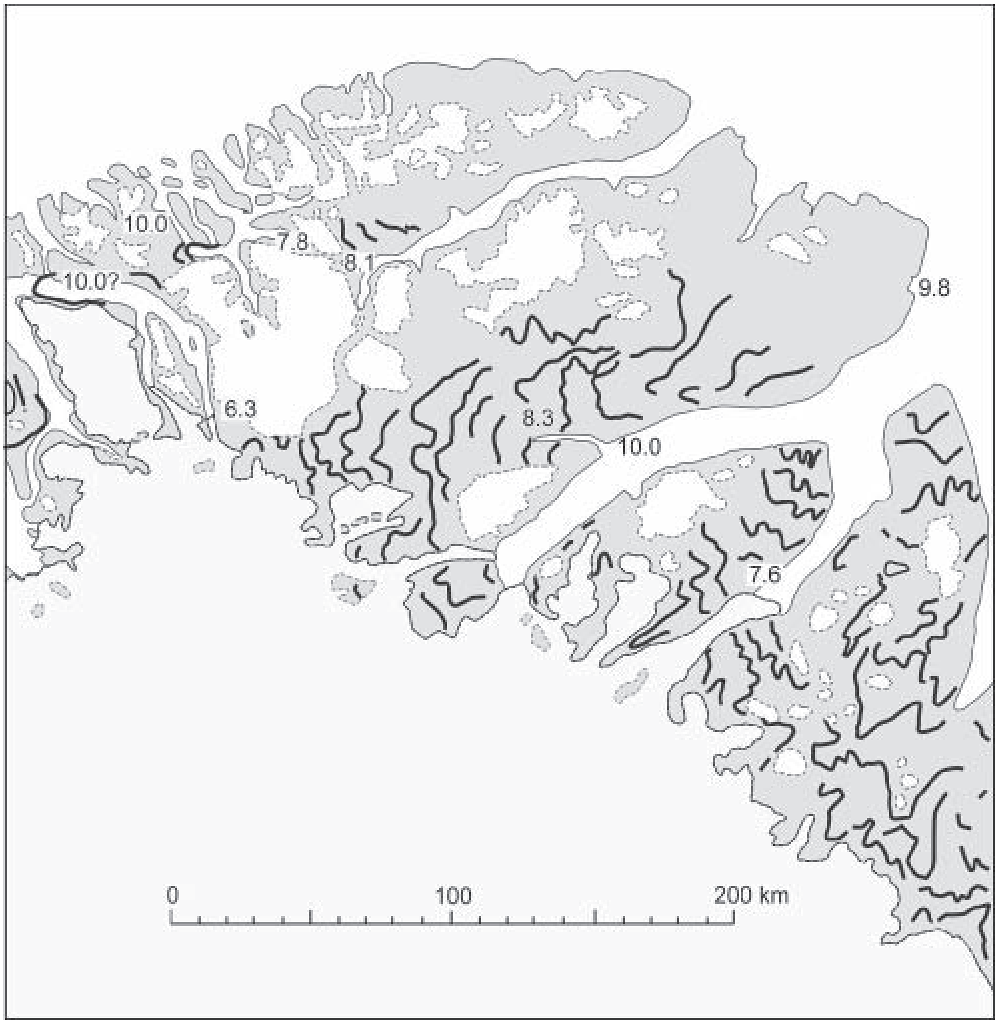

Fig. 8. Trend lines of early and mid-Holocene deglaciation simplified from morphological maps of North Greenland made by W. Davies, U.S. Geological Survey (Weidick & Dawes 1999). The maps are not published, but can be found in the files of GEUS. The age of deglaciation expressed in cal. yrs. BP.

land Ice Sheet margin is not well dated. The Warming Land stade, represented by moraines in the central part of J. P. Koch Fjord (Fig. 8), were assumed by Kelly & Bennike (1992) to date from 9500-8000 BP (ca 11,200-8900 cal. yrs BP). West of this fjord, in Wulff Land, Warming Land and Nyeboe Land, this recession was related to extensive downwasting of the ice, resulting in the formation of dead ice terrain at these localities (Kelly & Bennike 1992). We also know that the ice retreat in Independence Fjord had reached the mouth of the tributary Jørgen Brønlund Fjord by 8000 BP (ca. 8900 cal. yrs. BP)(Bennike 1987).

Inner J.P. Koch Fjord

Extensive deposits of Holocene age are located around the head of J. P. Koch Fjord and in the east-west trending valley between the fjord head (terminus of Adam Gletcher, an outlet of the Greenland Ice Sheet) and the outlets of the Hans Tausen Iskappe. Marine terraces are found up to the marine limit, estimated to be ca. 42 m a.s.l.

During fieldwork in 1976, shells were collected from silt in a terrace 30-36 m a.s.l. The silt is overlain by gravel that can be mapped up to the marine limit of ca. 42 m a.s.l. The shells yielded ages of 5550±105 BP (6350 cal. yrs. BP, I-9664) and 5480±80 (6280 cal. yrs. BP, Ua- 4587)

(Table 1). Shells embedded in fine-grained sand and laminated silt in a slightly lower terrace, 1.5 km downstream, were dated to 5085±60 BP (5880cal. yrs. BP, Ua-4586) (Table 1), whereas shells and twigs from the outermost silty neoglacial moraines of Adams Gletcher were dated to 4040±150 BP (4590 cal. yrs. BP, GSC-2279) and 220 BP (I-9130).

These dates from the southwestern margin of the Hans Tausen Iskappe suggest that the ice cover was close to or less than present prior to 5500 BP (6300 cal. yrs. BP), and that the Adams Gletcher had a readvance to its present position some time during the last 300 years, i.e. at the end of the Little Ice Age.

Independence Fjord

A reconstruction of the marine limit distribution in Northern Greenland has emerged from the increasing number of field observations in recent decades (Funder & Hjort 1980, Bennike 1987, Kelly & Bennike 1992, see also compilation by Funder & Hansen 1996). Relative sea-level curves (Fig. 9) for the inner part of Independence Fjord have been supplied from Jørgen Brønlund Fjord and from the mouth of the fjord at Kap København and Prinsesse Ingeborg Halvø, respectively (Funder & Abrahamsen 1988).

For the outer part of Independence Fjord, a marine limit of 65-100 m a.s.l. was determined by Funder & Hjort (1980). Based on the shoreline diagram (Fig. 10), the initial deglaciation of the area probably took place at about 9000 BP (10,200 cal. yrs. BP), which complies with earlier age estimates of an ice margin in the area at 9000-10,000 yrs. BP (10,200-11,500 cal. yrs. BP).

Based on the relative sea level curve (Fig. 9) and a marine limit of 80 m a.s.l. (Bennike 1987), we assume that the deglaciation of Independence Fjord reached Jørgen Brønlund Fjord by 8000 BP (8900 cal. yrs. BP), and about 7500 BP (8350 cal. yrs. BP) at the head of Jørgen Brønlund Fjord. This is somewhat younger than the age of 9500 ± 500 cal. yrs. BP as suggested by Bennike (1987).

The emergence curve for the inner part of the Independence Fjord shows that the relative sea level reached present level around 1000 years ago. The information is based on a Thule eskimo

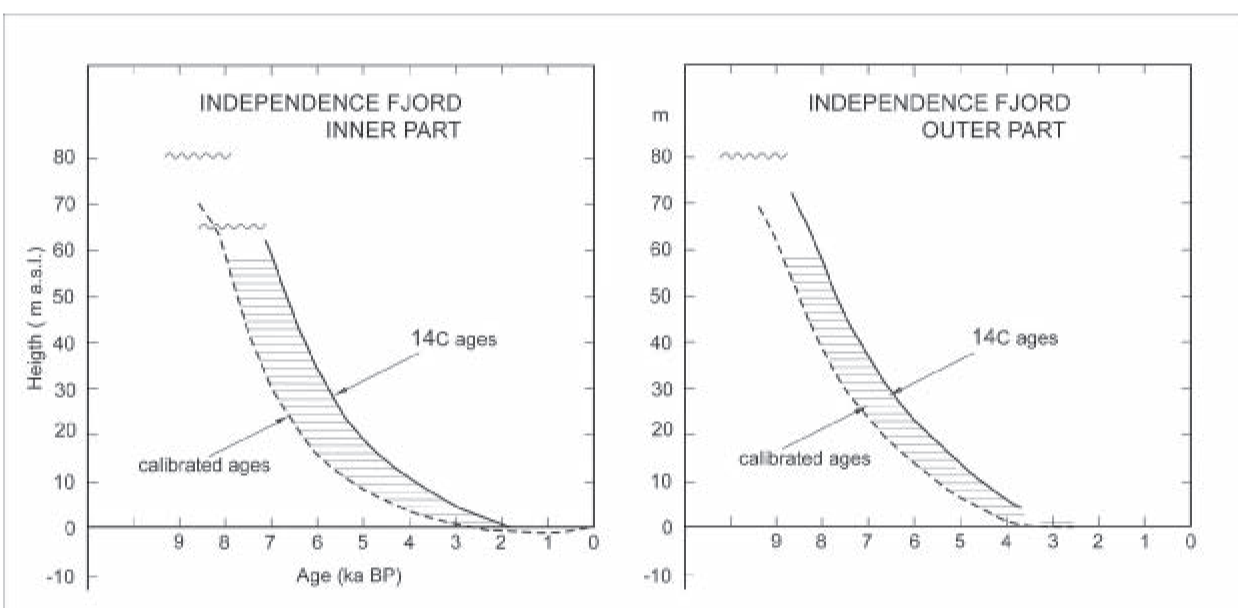

Fig. 9. Emergence curves of inner and outer part of Independence Fjord. Inner part covers the region around Jørgen Brønlund Fjord, the outer part the region around the mouth of Independence Fjord between Kap København and Prinsesse Ingeborg Halvø (see Fig. 10). Curves drawn by full lines are based on compilation of ^{14}C-datings, for the interior region by Bennike (1987) and for the outer region by Funder & Abrahamsen (1988). The broken lines are emergence curves based on calendar years.

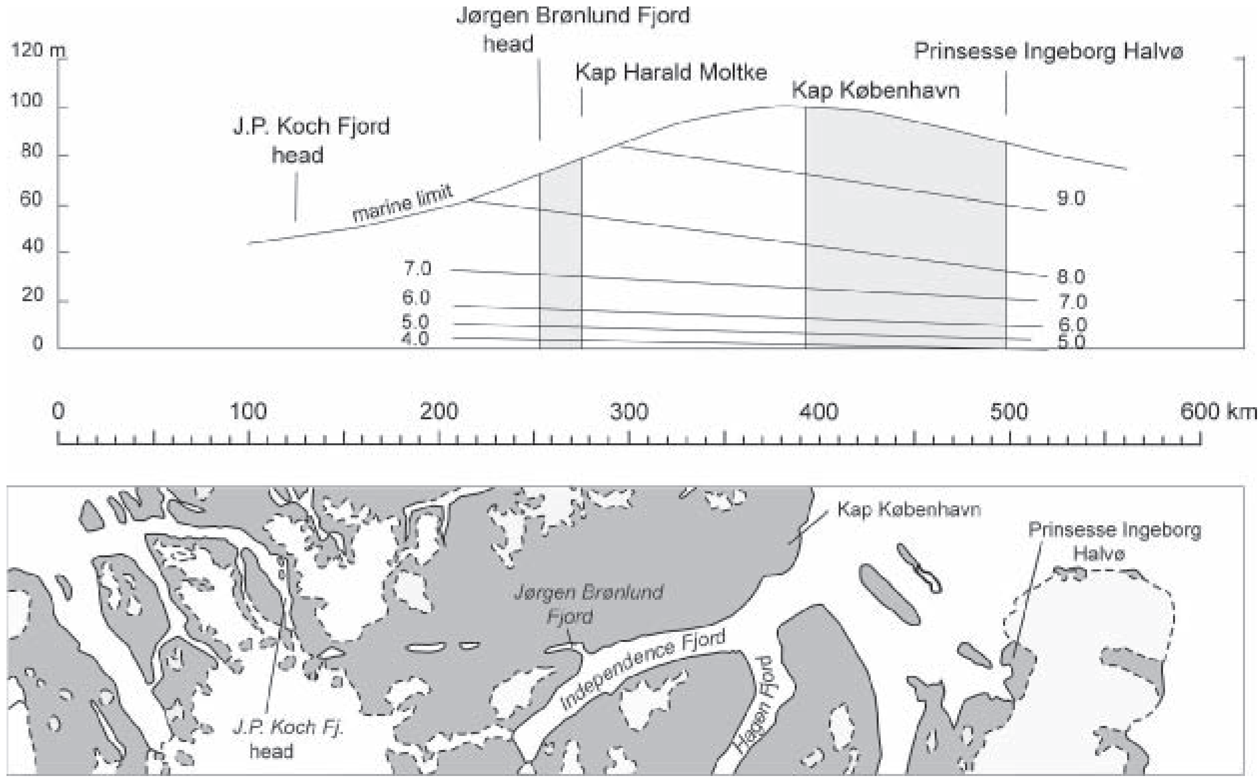

Fig. 10. Simplified outline of the marine limit and the conceptual trends of strandlines in a profile from the head of J.P. Koch Fjord in the west to the mouth of Independence Fjord in the east. Large figures: Estimated time for deglaciation expressed in calibrated ka BP. Hatched areas of the curves: areas covered by the two emergence curves of Fig. 9.

ruin on the south coast of Jørgen Brønlund Fjord which is transgressed by the sea (Bennike 1987). The emergence curves of the outer Independence Fjord do not extend to this period. However, drowned Thule culture ruins from this region (Kronprins Christian Land) were also reported by Hjort (1997). In both cases the sea level related to the Thule culture is only slightly (1 m or so) under the present one.

The calendar ages of the emergence curves of Fig. 9 are applied in the construction of the conceptual equidistant shoreline diagram along Independence Fjord (Fig. 10). A shoreline dip towards the outer coast can be seen, even if the gradient of the highest shorelines (8000 to 9000 cal. yrs BP) is only ca 10 cm/km. This is significantly lower than the 50 cm/km of a 9400 cal. yrs BP shoreline at Germania Land 600 km further to the south (Landvik 1994, Weidick *et al.* 1996). However, this could be an artifact due to limited control of the exact direction of the isobases in the area. The relatively low marine limits close to the Hans Tausen Iskappe indicate a late local deglaciation compared to the middle and outer parts of Independence Fjord.

Glacial History of the Hans Tausen Iskappe – Discussion

The extent of the last ice sheet

Our studies in Nordpasset show that an ice sheet that also filled adjacent fjords and valleys during the last glaciation inundated the 900-1000 m high bedrock plateau, which is covered by the northern dome of the Hans Tausen Iskappe. There is also strong evidence for a confluence between this ice sheet and the Greenland Ice Sheet to the south. Evidence for such a continuous ice cover during the Late Weichselian has also been found in the Jørgen Brønlund Fjord area, at the eastern end of the

Wandel Dal ice-free corridor, by Bennike (1987). He showed that outlets from both the Greenland Ice Sheet and a "Peary Land ice cap" reached Independence Fjord through the Wandel Dal, and that there was a halt in ice recession at the mouth of Jørgen Brønlund Fjord ca. 9000 to 7600 BP. Such an easterly ice flow in Jørgen Brønlund Fjord is not possible without a full confluence between the Late Weichselian Greenland Ice Sheet and the ice over the Hans Tausen plateau.

The extensive glaciofluvial deposits in the large valleys north of Nordpasset suggest further confluence with ice over the mountains of northern Peary Land (namely, Johannes V. Jensen Land). Weidick (1976) pointed out that only a slight depression of the present glaciation limit would lead to the formation of an ice cap over Peary Land. As reviewed by Funder & Hansen (1996), there are still large uncertainties whether large ice shelves existed along the coast of Peary Land (Funder & Larsen 1982, Dawes 1986), or only restricted piedmont glaciers reached the western parts of the coastline.

The last deglaciation

Northern margin of Hans Tausen Iskappe

The successive lowering of the glacier surface in Nordpasset shows that the whole ice mass must have experienced large surface melting during the deglaciation. Even the highest areas were brought under the equilibrium line due to the climatic amelioration. Such a style of deglaciation over the area implies that the whole Hans Tausen plateau must have been ice-free before the formation of the ice marginal features we have reported from Odin Fjord and Nordpasset. The higher ground, including the Hans Tausen plateau, must have been deglaciated prior to ca. 7200 BP (8100 cal. yrs. BP), as indicated by the deglaciation dates in eastern Nordpasset.

The consequence of this deglaciation model is that the present-day Hans Tausen Iskappe formed after 8100 cal. yrs. BP as a result of an equilibrium line lowering during the Middle or Late Holocene. This conclusion is supported by the results of the ice core from the ice cap. The age estimates of the 345 m-long core suggests that the entire ice cap formed after 3500-4000 years BP (Hammer *et al.* 2001).

The margins of several outlet glaciers along the northern margins of the Hans Tausen Iskappe were studied in order to reconstruct any Late Holocene glacier fluctuations. There are fresh-looking moraine ridges in close contact with the glacier snout. In front of most of the outlet glaciers there is a clear morphological contrast between these ridges and the old landscape which lacks pronounced moraine ridges. This maximum position in the area was generally reached ca AD1900, with a subsequent recession or still-stand (Weidick 2001). The morphological contrast and lack of older moraine ridges suggest that the present glacier margins were at their outermost position during the whole Holocene. The terminal moraines of two outlet glaciers from the northern dome, the one in eastern Nordpasset and in Tjalfe Gletcher at the head of Adolf Jensen Fjord, were visited in 1994. Despite a thorough search for radiocarbon-datable material, neither of the advances to the present day position could be dated.

The southern margin and the Independence Fjord basin

The onset of the Holocene ice retreat in the area between the Hans Tausen Iskappe and the Greenland Ice Sheet cannot be settled exactly. With allowance for the preceding readvances of the Warming Land stade (Kelly & Bennike 1992) in North Greenland (around J. P. Koch Fjord) and the dated events of Independence Fjord, the retreat might be ascribed to around 9000 BP (10,200 cal. yrs. BP)(Bennike 1987). The ice margin retreat must have occurred as a fast break-up due to calving in the fjords. This is supported by the annual recession rate of 100 m/year or more that we have calculated in Independence Fjord, which contrasts with only 50 m/year in the narrower Jørgen Brønlund Fjord. Here, as elsewhere, topographic conditions (sills, narrowings of the fjords) might have caused halts of the recession through the fjords (cf. Mercer 1961). However, with exception of the 85 m-deep Jørgen Brønlund Fjord (Høy 1970), the bathymetry of the North Greenland fjords is unknown.

The approximate rate of ice recession over the land area can be calculated from the deglaciation of Jørgen Brønlund Fjord, i.e. 50 m/year. This is similar to a recession rate of 47 m/year over Germania Land further south (Weidick *et al.* 1996). The recession to the present extent of most larger ice caps in the area north of Independence Fjord then would have taken 1-2 ka.

The fast recession of the outlet glaciers occurred at the same time as the first openings in the permanent fjord ice cover (see below) which is probably related to the early Holocene temperature increase of the surface waters in the Greenland Sea (Koç *et al.* 1993, Koç & Jansen 1994). During this change, the glaciers presumably shifted their frontal characteristics to a "temperate mode". The recession rates were dependent on fjord width and calving rate. We conclude that the ice in the larger fjords (e.g., Independence Fjord) attained its present position at about 8000-9000 cal. yrs. BP, whereas this was reached at about 6300 cal. yrs. BP in the narrower fjord systems such as J. P. Koch Fjord.

Here, the ice recession to a position behind the present margin occurred shortly before 5600 BP (6350 cal. yrs. BP). This development is comparable to other dated sites in North Greenland. Reworked shells in the present moraines are dated to 5000 BP (5700 cal. yrs. BP) at C.H. Ostenfeldt Gletscher, ca. 100 km WSW of Adams Gletscher, and to 4700 BP (5400 cal. yrs. BP) at Steensby Gletscher, nearly 250 km WSW of Adams Gletscher at 12 km behind its front (Kelly & Bennike 1992). Similar evidence is obtained from the Inland Ice outlet Nioghalvfjerdsfjorden Gletscher at 79(N in Northeast Greenland. The fjord was deglaciated to 80 km behind its present front shortly after 8000 cal. yrs. BP and became glacier-filled to its present extent after ca 4500 cal. yrs. BP (Bennike & Weidick 1999).

As shown by our studies from Hans Tausen Iskappe, the glacier recession in the fjords occurred at the same time as surface ablation and thinning of the ice cap. It must be presumed that the apparently thin ice caps such as Storm Iskappe and Chr. Erichsen Iskappe, as well as those to the south of Independence Fjord, disappeared. The trend lines of the deglaciation around these ice caps are cross-cut by the present ice cap margins (Fig. 8), which indicates a substantial regrowth of the ice caps after the initial deglaciation of the region. Some of the largest ice caps (e.g., Hans Tausen, Nordkronen, Heinrich Wild) conceal a subglacial alpine topography, covered by several hundred metres of glacier ice, as shown for the Hans Tausen Iskappe.

Post-Glacial Climate

The fjords in this part of north Greenland today experience a semi-permanent sea-ice cover where the fjord ice melts only at rare intervals (>30 years)(Higgins 1990). The finding of driftwood both at the eastern end of Nordpasset and Kap Bopa (Table 1) shows that the fjords were seasonally ice-free 5850-5500

BP (6700-6300 cal. yrs. BP). The two driftwood samples have been identified as either *Picea* sp. or *Larix* sp., but could not be identified to species level due to poor preservation (L.M. Paulssen, written comm. 1995). Thus, their geographical growth area could not be determined.

There are several reports of driftwood from North Greenland. From the Jørgen Brønlund Fjord area, Bennike (1987) compiled radiocarbon dates from different sources, including 17 samples of wood, both driftwood and charcoal. These dates cover a time-span from 5900 to 2500 BP (6600-2550 cal. yrs. BP), which is in agreement with our dates showing that the fjords were seasonally free from fjord ice shortly after 6000 BP. Several pieces of driftwood below the 2500 BP date at 6 m a.s.l. (Bennike 1987) suggest that predominantly seasonally open conditions prevailed until at least 1000-1500 BP, when sea-level dropped below present (Fig. 9). However, there seems to be a contrast in sea-ice survival between the fjords facing the Arctic Ocean, as discussed by Higgins (1990), and the fjords facing the northern Fram Strait, such as Independence Fjord and the tributary Jørgen Brønlund Fjord.

Hjort (1997) suggested ice-free fjords during the early and mid-Holocene warming in Northeast Greenland, followed by intermittent periods of ice-covered and ice-free fjords during the subsequent neoglacial climate decline after ca. 5700 cal. yrs. BP. It is unknown to what degree such periods of ice-free fjords increased the regional precipitation, especially in summer months, but its importance for the acceleration of calving from outlet glaciers is substantiated by the descriptions by Higgins (1989, 1990).

The vegetation records give some other constraints on the Holocene climatic optimum in North Greenland. In a study of the vegetation history from Klaresø on the south coast of Jørgen Brønlund Fjord, Fredskild (1969, 1973) showed that the period from 5000 to 3300 BP (5700 to 3500 cal. yrs. BP) was characterised by richer vegetation than today, probably caused by enhanced precipitation due to locally open fjords. Distinct drops in the local sedimentation rate and pollen influx occurred at 3300 BP (3500 cal. yrs. BP) and 2100 BP (2100 cal. yrs. BP). A slightly earlier transition to colder summers at 3900 BP (4400 cal. yrs. BP) was concluded for lake Sommersø close to Station Nord (Funder & Abrahamsen 1988). The fact that this lake has a more oceanic setting than the lake in the Jørgen Brønlund Fjord area suggests that a cooling of regional significance occurred, as is also suggested by Funder & Abrahamsen (1988).

These intervals also coincide with the interpretation of the ice core records from the Hans Tausen Iskappe, which suggests that the build-up of the ice cap started after ca. 3500-4000 cal. yrs. BP, probably as a response to increased precipitation during summer (Hammer *et al.* 2001).

The period of ice-free conditions in the fjords also encompasses the date range of the paleo-eskimo cultures in Peary Land. They have been dated to 4400-4000 BP (4900-4500 cal. yrs. BP) (Independence I) and 2800-2400 BP (2900-2400 cal. yrs BP) (Independence II) (Knuth 1984, Andreasen 1996). The end of Independence II occurs at the same time as the driftwood stranding ceased in the fjord region.

Conclusions

The plateau covered by the Hans Tausen Iskappe, and adjacent areas, were fully glaciated during the last glaciation (Late Weichselian).

The ice cap was probably confluent with the Greenland Ice Sheet to the south, and with ice caps over northern Peary Land to the north.

The last deglaciation occurred as a

downwasting of the ice sheet and ice caps, combined with a rapid recession by calving along the large fjords, leaving the Hans Tausen plateau ice-free by 7200 BP (8100 calendar years BP).

There was a rapid ice recession along the large Independence Fjord, which was essentially completed by 9000 cal. yrs. BP.

Dates of driftwood show that the presently sea-ice covered fjords were seasonally ice-free during the Mid-Holocene (6800-2500 cal. yrs. BP). This correlates with the presence of the paleo-eskimo cultures in North Greenland.

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