DANMARKS OG GRØNLANDS GEOLOGISKE UNDERSØGELSE RAPPORT 2007/25

Previous glaciological activities related to hydropower at Paakitsoq, Ilulissat, West Greenland

Andreas Peter Ahlstrøm

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MINISTRY OF THE ENVIRONMENT

GEUS

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G E U S

Contents

Introduction

The current report constitutes a brief summary of the previous glaciological activities carried out in the Paakitsup Akuliarusersua catchment located *c.* 45 km northeast of Ilulissat in West Greenland (69°27’N, 50°19’W). Emphasis is given to the efforts of the Geological Survey of Denmark and Greenland, GEUS (previously the Geological Survey of Greenland, GGU), but some activities of other institutions are also mentioned. The report has been produced at the request of Nukissiorfiit (the Greenland Energy Authority) under contract to ASIAQ (the Greenland Survey).

Historical overview of research activities

The margin of the Greenland ice sheet near Paakitsup Akuliarusersua, often referred to as Paakitsoq (old spelling: Pâkitsoq) after the nearby bay, has been the subject of a large number of glaciological studies and is still an area of international research activities. The early work in the region was mainly due to the relatively easy access onto the ice sheet margin and the proximity to the central part of the ice sheet. This made it possible to establish measurements along a transect covering the complete elevation range of the Greenland ice sheet. Gravimetric and seismic measurements were carried out on such a transect by the Expéditions Polaires Francaises, Missions P.-E. Victor (EPF) and reported in 1950 (Munck, 1950; Joset, 1950; Holtzscherer & Bauer, 1954). The EPF was later followed by the Expeditions Glaciologiques Internationale au Groenland (EGIG) which carried out geodetic and mass balance measurements.

The early scientific campaigns provided a baseline of measurements that made it an attractive region to return to. Apart from this the main reasons for scientific interest has been:

- the proximity of the Jakobshavn Isbræ which has a significant influence of the total mass loss of the Greenland ice sheet
- the long meteorological record available from Ilulissat dating back from 1873
- the occurrence of pre-holocene ice near the margin for paleoclimatic studies

Additionally, the hydropower pre-feasibility studies in the region starting around 1980 brought researchers in, eventually paving the way for a semi-permanent research station near the equilibrium-line altitude on the ice sheet. The station was established and operated in the early 1990s by the ETH Greenland Expeditions (ETH = Eidgenössische Technische Hochschule, Zürich, Switzerland) and was labelled ETH Camp. In the mid-1990s, the responsibility for the station was taken over by CIRES/CU (Cooperative Institute for Research in Environmental Sciences/Colorado University) and has since been referred to as ETH/CU Camp or Swiss Camp. The Swiss Camp played an important role in the large-scale PARCA project (Program for Regional Climate Assessment) funded by NASA. Since the mid-1990s, researchers from CIRES have also operated a number of automatic weather stations (AWS) positioned in a transect within the Paakitsup Akuliarusersua catchment up to Swiss Camp. This line of automatic weather stations has been labelled the JAR-transect (Jakobshavn Ablation Region). These AWS constitutes a part of the USfinanced GC-Net of weather stations on the Greenland ice sheet. The Swiss Camp has also hosted a number of other ice-sheet research activities.

Hydropower pre-feasibility studies

The oil crisis in the early 1970s spurred an interest in becoming self-sufficient with energy in Greenland, with hydropower as the most obvious source that was locally available. A large-scale effort was initiated, mainly through the Greenland Technical Organisation (GTO), to locate and quantify the hydropower potential in Greenland. The hydropower investigation responsibilities was continued by Nukissiorfiit and the Greenland Survey (ASIAQ) as former parts of the now defunct GTO. The Greenland Homerule has decided on a long-term plan to switch from diesel-powered plants to hydro-electric power plants, most recently outlined in the report Energiplan 2020 (2005).

Investigations were carried out by GTO in Paakitsup Akuliarusersua area since 1976 (GTO, 1979). The first campaigns aimed at utilizing the entire Paakitsup Ilordlia fjord for hydropower by damming towards the Paakitsoq bay at the outlet. However, since 1980 the investigations concentrated on the smaller catchment at Paakitsup Akuliarusersua, draining towards the lakes 326, 233 and 187 (GTO, 1981). GTO and later the Greenland Survey (ASIAQ) has since then collected data in the catchment, including parameters such as water discharge, air temperature and precipitation (e.g. GTO, 1982, 1983, 1984).

The Geological Survey of Greenland (GGU, later GEUS) got involved with related glaciological investigations in 1982 with support from the European Fund for Regional Development of the EEC. These activities had a direct hydropower focus until the late 1980s when Greenland left the EEC. The annual glaciological investigations of GGU in the area continued into the beginning of the 1990s, but with emphasis shifted towards research with an increasing participation from international partners. In the field campaign of 1988 included researchers from Switzerland, the Netherlands, West Germany and USA, looking for a site to study the future effects of global warming. Eventually, GGU shifted the attention away from the annual involvement in the Paakitsup Akuliarusersua region, while the international research activities increased significantly. The activities of the GGU in the period 1982-1995 comprised of:

- Mass balance observations
- Ice-dynamic modelling
- Photogrammetric and satellite mapping
- Ice-penetrating radar surveys
- Hot-water drilling
- Ice-temperature measurements
- Dye-tracing studies
- Basin delineation modelling
- Run-off simulations
- Meltwater retention studies

The long and deep involvement is reflected in the broad scope of the glaciological work that was carried out. All activities were either directly carried out as hydropower pre-feasibility studies or alternatively as research of immediate importance for the hydropower considerations. The following sections will briefly present these efforts, starting with a characterization of the Paakitsup Akuliarusersua basin.

Characterization of the catchment

The Paakitsup Akuliarusersua drainage basin, hereafter referred to as Paakitsoq for the sake of brevity, lies between 69°25'N to 69°32'N and 50°05'W to 50°20'W. Excluding the ice-sheet part, the basin covers an area of 33.6 km^2 within an altitude range of *c.* 200–600 m asl. The majority (> 90%) of the basin runoff is meltwater from the ice sheet, draining through three lakes labelled 326, 233 and 187. The basin contains four outlet glaciers (or sectors) from the inland ice sheet, numbered 1GE07001–2 and 1GE04001–2 (Weidick et al., 1992). The central outlet glacier, 1GE07001, terminates in lake 187 which contains the outlet from the entire catchment. According to Weidick (1968), the glaciers have experienced thinning since 1880. The meltwater from lake 233 presently drains along the northern edge of glacier 1GE07001 to lake 187. This drainage configuration has previously been blocked by glacier 1GE07001, a situation which caused the water to drain out of the westernmost part of lake 233 instead. Likewise, the meltwater from lake 326 to lake 187 previously drained over a bedrock threshold in the northwestern corner of lake 326. This configuration had been stable for at least 41 years as determined from aerial photography studies, but changed during January-May 1989 to a drainage underneath the ice sheet margin instead, although still from lake 326 to lake 187 (Thomsen & Olesen, 1990). As a result, the water level of lake 326 had dropped 14.5 m by July 30, 1989, corresponding to a water volume loss of 30×10^6 m^3. These changes demonstrate that even the short-term stability of the drainage configuration in the ice-free part of the Paakitsup Akuliarusersua basin is dependent on the behaviour of the ice-sheet margin.

The part of the cathcment covered by the inland ice-sheet is even harder to pin down. Meltwater may form underneath the ice sheet, but the majority is formed by melting of the ice sheet surface (ablation). The newly formed meltwater trickles along as innumerable small surface streams, eventually forming large supraglacial river systems. The rivers follow the surface topography until it reaches a crevassed area or disappears into moulins. Surface drainage depends largely on local surface undulations, reflecting subglacial conditions and structural features such as shear bands, healed crevasses and ridges indicating the ice flow pattern (Thomsen & Reeh, 1986). The rivers may flow in canyons up to 6 m deep, cutting through the surface undulations. The river bed in these canyons sometimes inclines inland, opposite the main inclinination of the ice surface, directing the meltwater away from the ice margin. Such drainage conditions complicate mapping of surface drainage, if solely based on topographic information (Thomsen & Reeh, 1986).

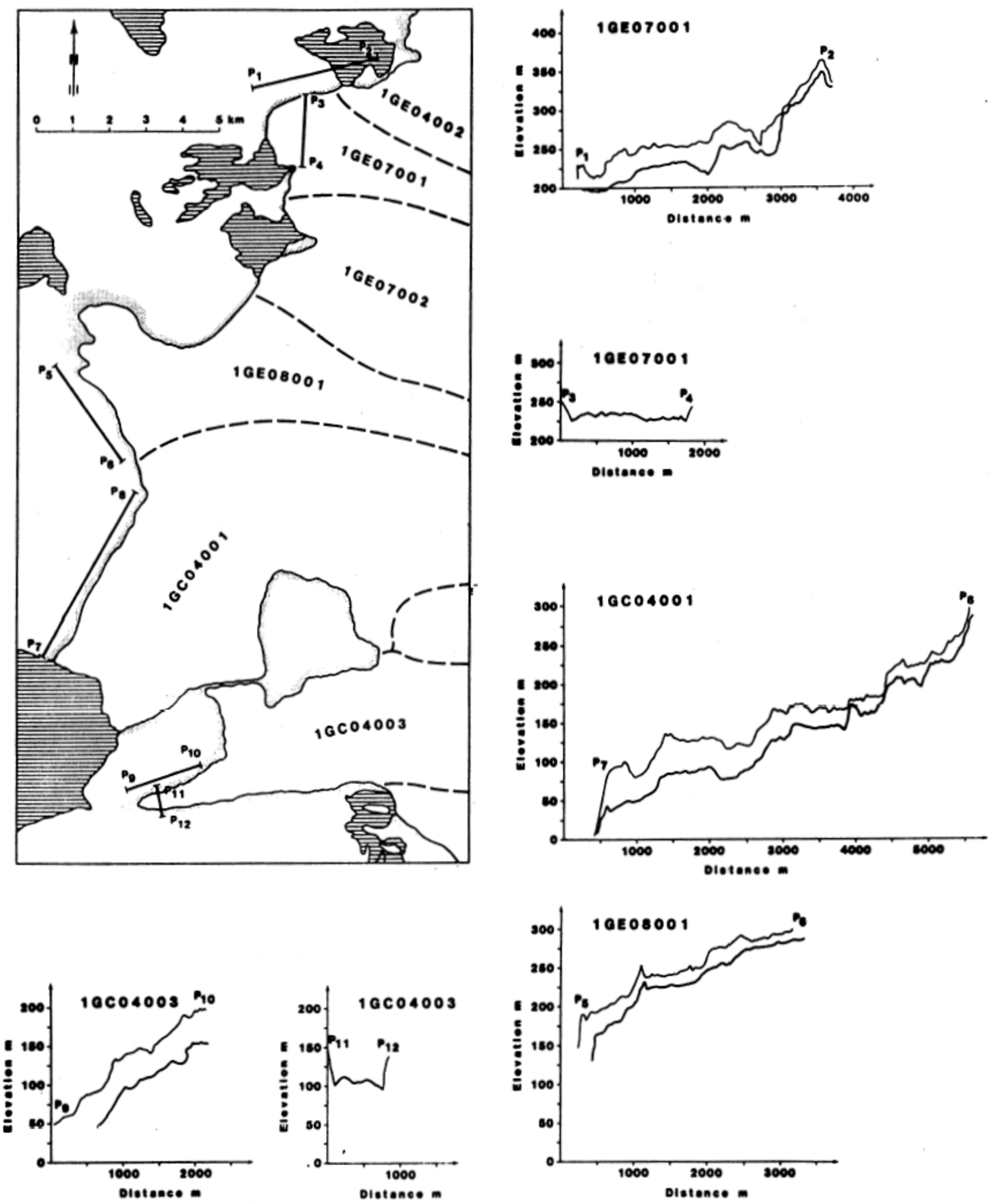

Figure 1. *Profiles and positions of profiles. Upper thin profiles show the height of the trim line zone, corresponding to the extend of the glaciers around 1880. Lower thick profiles show the height of the ice margin in 1959 (Fig. 6.2 in Thomsen, 1983b).*

Glacier moulins provide efficient pathways for the meltwater to enter the englacial system and may penetrate to the bottom of the glacier, even through 'cold' ice (ie. ice of sub-zero temperature). Moulins are initially formed by meltwater entering into a crevasse, eventually melting a rounded conduit pathway that often persists even when the original crevasse has closed. Beneath the ice sheet, the subglacial hydrological system carries the meltwater towards regions of lower pressure, ie. towards the ice margin. The conduits carrying the meltwater inside and underneath the ice sheet will be shaped by a continuous struggle between melting of the conduit walls by the passing meltwater and conduit closure by ice-flow. As the amount of meltwater entering the system is highly variable on many time scales, the capacity of the ice-sheet plumbing will never reach a steady state. This means that the water pressure in the subglacial drainage system at the base of the ice sheet will vary with the season as well as diurnally, due to the changing input of surface meltwater. Thus the natural variations in surface meltwater production may potentially cause variations in the subglacial drainage configuration that leads the meltwater to the ice-free part of the basin. In other words, the size of the catchment (and thus the amount of available water for hydropower) may depend on ice-sheet geometry as well as the amount of surface meltwater entering into the englacial system.

Evidently, a catchment containing an ice-sheet sector requires a glaciological analysis in order to assess the future availability of water for hydropower purposes. Such an analysis ideally comprises of surface mass-balance measurements to determine the amount of meltwater forming, ice-dynamic modelling to assess the evolution of ice-sheet geometry and temperature over time, photogrammetric and satellite mapping to determine the surface topography and drainage patterns, mapping of the bedrock topography by ice-penetrating radar, hot-water drilling to determine the subglacial water pressure and the englacial temperature, dye-tracing studies to help the basin delineation, modelling of future changes in the basin delineation, simulations of meltwater run-off for a range of scenarios and meltwater retention in the cold snowpack. All these aspects of a glaciological analysis have been carried out by the GGU in the Paakitsup Akuliarusersua catchment as summarized in the following sections.

Mass balance observations

An important part of the glaciological programme of the GGU in Paakitsoq was the mass balance measurements. These point data are vital for calibration of distributed catchment-scale melt-models. Modern mass balance measurements are often carried out by automatic weather stations (AWS) and GEUS has developed a new system called automatic mass-balance stations (AMS). Both station types have temporally continuous mass balance data. The traditional mass-balance measurements that were carried out in the Paakitsoq basin, were done by insertion of long aluminum stakes into the ice sheet surface. These stakes were then revisited in the spring to measure the snow accumulation and density and in late summer to measure the snow and ice melt (ablation). A stake transect was established and visited by GGU in the spring and late summer every year from August 1982 to 1990. During 1990–1992 a stake line was established at higher elevations (above ETH Camp) in the basin to study firn densification and meltwater refreezing (Braithwaite et al., 1992, 1993; Braithwaite, 1993). Mass balance data from this transect is available for the period August 1990 – August 1991 (Braithwaite et al., 1992).

The individual stakes in a transect are ideally supposed to represent an elevation interval with a large area. However, measurements may vary with up to 17% between stakes within a few metres due to local variations and measurement inaccuracies (Braithwaite, 1983; Clement, 1983). Thus, it must be emphasized that the stake measurements of mass balance reported here represents individual point measurements at different elevations, and that care should be taken when attempting to extrapolate these measurements to catchment scale. The increasing homogeneity of the ice sheet surface with elevation means that the representability of a stake improves at higher elevations.

1982

The initial stake deployment in 1982 was designed to cover the elevation range from 300 m asl. at the ice margin and up to 1500 m asl. The stake transect included seven stakes, positioned with elevation intervals of roughly 200 m, with the lower stakes ending on glacier lobe 1GE07001 in Lake 187. The field work was carried out by helicopter. It was not possible to insert stakes at higher elevations due to limitations in the helicopter navigation system and the great distances involved (Thomsen, 1983).

1983

The stakes deployed in 1982 were visited again by helicopter on May 12 and on August 11, 1983. The positions of the stakes had to be revised and stake 13 was not found. It was not possible to reestablish the stake as the motorized drill was rather inefficient at strongly negative snow/ice temperatures. The snow cover experienced in the spring visit was very uneven, probably due to wind redistribution, up to an elevation of 500 m asl., but was more even from 700 m asl. Snow at low elevation only occurred in local depressions or crevasses and there was no snow at stake 3.0 and 5.0. At the other sites, snow pits were dug yielding rather similar snow densities between 330 kg/m^3 and 380 kg/m^3 with a mean of 360 kg/m^3. The temperature of the snow was between −9°C at the surface to −12°C at the bottom of the snow pits. At the summer visit, stakes 13.0 and 15.0 from 1982 were not found. Four new stakes (8.0, 10.0, 12.0 and 13.0) were inserted at elevations of 800, 1000, 1200 and 1300 m asl., respectively. The snow cover was melted away up to an elevation of 900 m asl. and refreezing of meltwater in the snowpack was registered at 1100 m asl. The annual equilibrium line altitude (ELA) was determined to be around 1030 m asl. (Thomsen, 1984c). This relatively very low value, when comparing to average conditions, and is based on the general air temperature depression following the eruption of the Mexican volcano, El Chichón, in 1982. The influence of this event is also seen in other mass balance records from Greenland (Ahlstrøm et al., in press).

Højde i m over havniveau	Masseændringer (mm vandækvivalent) 15/8 82 - 12/5 83	Balance (mm vandækvivalent) 15/8 82 - 12/5 83	Masseændringer (mm vandækvivalent) 12/5 83 - 11/8 83	Balance (mm vandækvivalent) 12/5 83 - 11/8 83	Balance (mm vandækvivalent) 15/8 82 - 11/8 83	Ablation mm vand-ækviva-lent 15/8/82-11/8/83
300	-423	-423	-2322	-2322	-2745	-2745
500	-306	-306	-1710	-1710	-2016	-2016
700	(+300) -306	-6	(-300) -567	-867	-873	-1173
900	(+429) -90	+339	(-429) -228	-657	-318	-747
1100	(+577) -108	+469	(-351) (+50)+	-301	+168	-409
1300	÷	÷	÷	÷	÷	÷
1500	(+502) -36	+466	÷	÷	÷	÷

(300) Tal refererer til sne
300 Tal refererer til is
+ Genfrysning i snepakke (usikker bestemmelse)
÷ Stage ikke genfundet

Table 1. *Mass changes and balance on the Inland Ice at Paakitsoq. Figures in brackets refer to snow and those without brackets refer to ice. Units are mm w.e. (water equivalent) and elevations are given in m asl. The mark '+' implies refreezing in the snow (uncertain measurement) and the mark '÷' that no stake was found.*

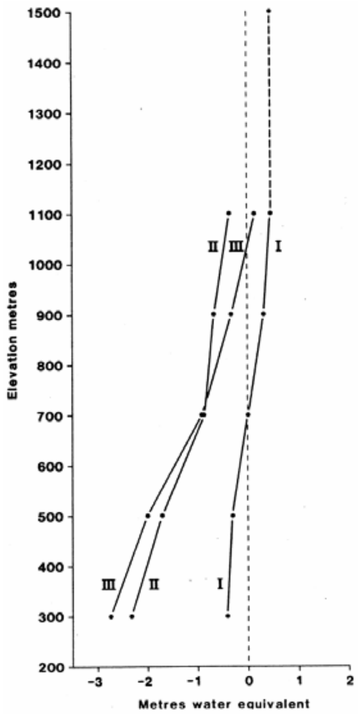

Figure 2. *Mass balance in relation to elevation on the Inland Ice at Paakitsoq. (I) Transient balance: August 15, 1982 – May 12, 1983. (II) Transient balance: May 12, 1983 – August 11, 1983. (III) Annual balance: August 15, 1982 – August 11, 1983. (Fig. 4.2 in Thomsen, 1984c).*

1984

The stake transect was visited by helicopter on May 15 and August 24, 1984. Three stakes (9, 12, 13 and 15) could not be found in May, and four stakes (9, 12, 13, 15) could not be found in August. The winter snow was very patchy, and was confined mainly to drifts in gullies and crevasses up to stake 7, while it was continous at higher elevations. The transient balance was measured in snow pits and by depth soundings at the stakes. The winter snow was probably redistributed by wind drifting as there were no signs of heavy winter melting. The data was reported as a plot of the mass balance with elevation, showing that the annual equilibrium line altitude would have to be higher than 1100 m asl. and probably lower than 1200 m asl. (Thomsen, 1985a).

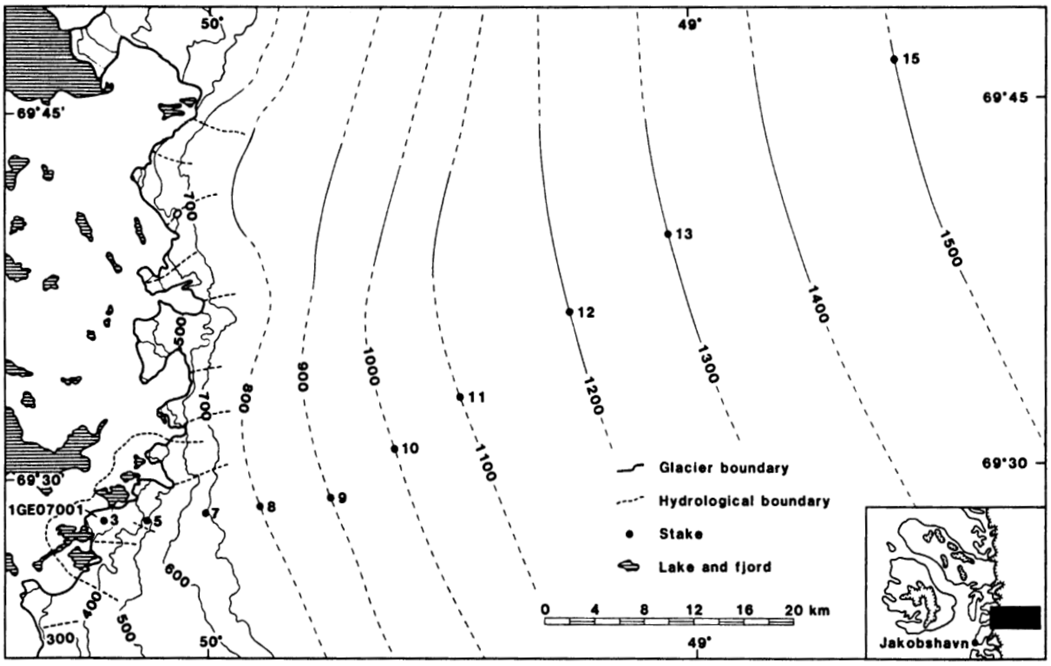

Figure 3. *Location of stakes at Paakitsoq in 1984. (Fig. 35 in Thomsen, 1985a).*

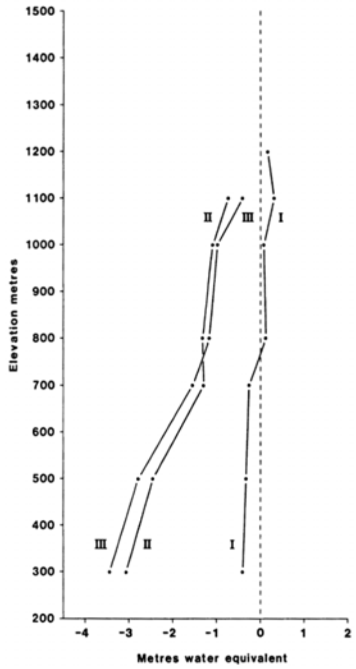

Figure 4. *Mass balance in relation to elevation on the Inland Ice at Paakitsoq. (I) Transient balance: August 11, 1983 – May 15, 1984. (II) Transient balance: May 15, 1984 – August 24, 1984. (III) Annual balance: August 11, 1983 – August 24, 1984. (Fig. 36 in Thomsen, 1985a).*

1985

The stakes were visited by helicopter on May 9, July 25 and August 7, 1985. There were again some difficulties in locating the stakes. In May stake 11.0 was not found. A new stake 4.0 was established. In July, stakes 8.0 and 10.0 were not found and stake 11.0 was recovered. Six new stakes (stakes 6.0/0785, 7.5/0785, 8.0/0785, 10.0/0785, 11.0/0785 and 12.0/0785) were established. In August stake 8.0 was recovered. Repeated altimeter readings revealed deviations from the elevations previously assumed (Thomsen, 1984c, 1985a). Generally, the stake elevations were lower than first believed. The stakes were surveyed in August with microwave line-of-sight, radio-location equipment belonging to GTO, with improving the positioning. Observations of the winter snow disrtibution were similar to the previous two years. The ablation was high compared to that of earlier years. On August 7, firn was still covering the surface at stake 12.0/0785. The firn cover was continuous at higher elevations and the transition between firn covered surface and bare ice was between stake 11.5/0785 and stake 12.0/0785. From these observations the annual equilibrium line altitude was approximately 1100 m asl. (Thomsen and Reeh, 1986).

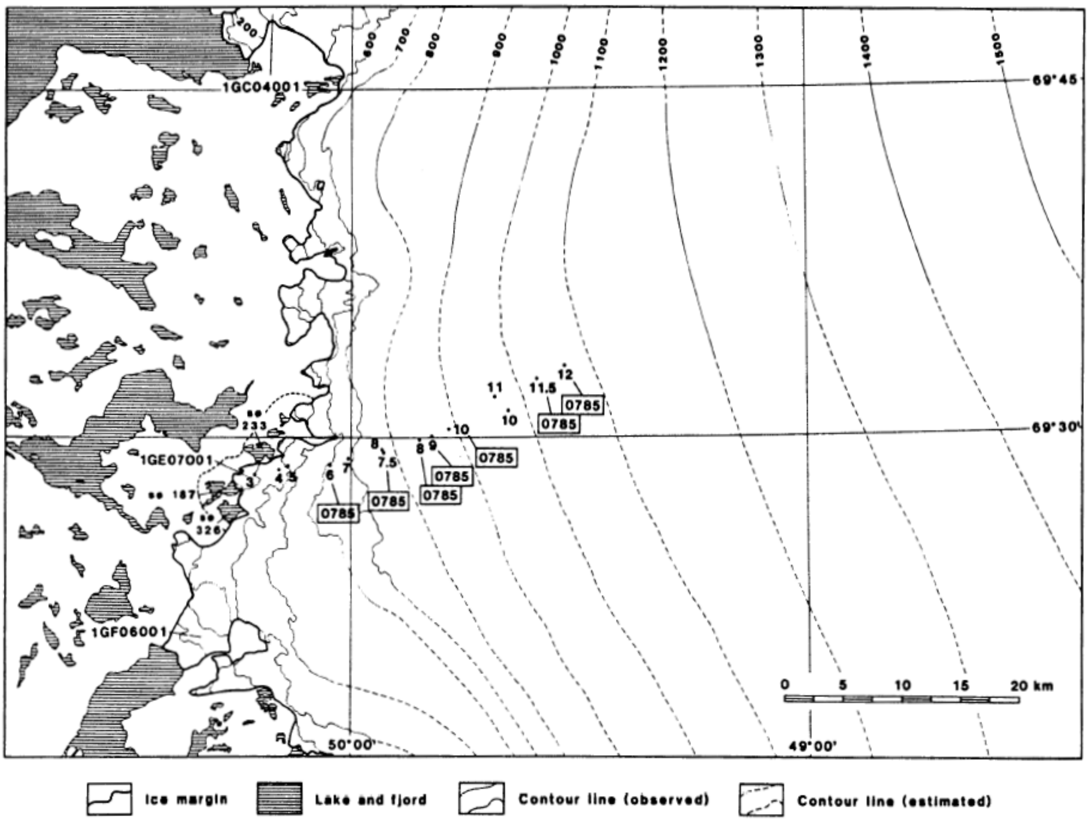

Figure 5. *Stakes at Paakitsoq in 1985 and 1986. Position of stake 10.0 is estimated as no survey has been carried out. Contours in metres. (Fig. 1 in Thomsen & Reeh, 1986).*

Stake	Approximate elevation m a.s.l.	24th Aug 84 9th May 85	9th May 85 7th Aug 85	25th Jul 85 7th Aug 85	24th Aug 84 7th Aug 85
3.0	245	−482	−3168	−513	−3650
4.0	395	–	−2475	−423	–
5.0	440	−396	−2655	−432	−3051
6.0/0785	570	–	–	−279	–
7.0	630	−102	−1862	−294	−1964
7.5/0785	735	–	–	−270	–
8.0	735	−63	−1476	–	−1539
8.0/0785	780	–	–	−333	–
9.0	870	3	−1619	−275	−1616
10.0/0785	910	–	–	−230	–
10.0	970	106	–	–	–
11.0	995	–	–	−342	−1428
11.5/0785	1055	–	–	−261	–
12.0/0785	1120	–	–	−(48)	–

– Stake not found or not established ( ) Icy firn estimated density 0.6 g/cm³

Table 2. *Transient balance and annual balance of the Inland Ice at Paakitsoq in millimetres of water (Table 1 in Thomsen & Reeh, 1986).*

1986

The stakes were visited by helicopter on May 13 and August 25, 1986. Six new stakes (stakes 2, 21, 22, 23, 24 and 25) were established in May. The six new stakes were all located near the ice margin ending in Lake 187 in the vicinity of stake 2. Winter snow was patchy (as described for 1984) up to 600 m asl. and continuous at higher elevations. A higher snow accumulation was registered compared to the two previous years, whereas the summer ablation was smaller. The annual equilibrium line altitude was between 1000 and 1100 m asl. (Thomsen, 1987a).

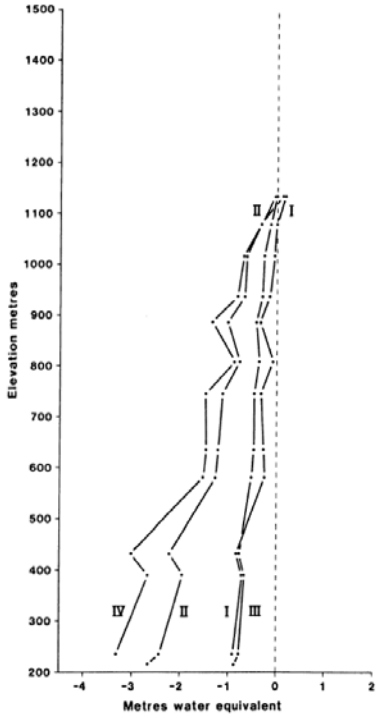

Figure 6. *Mass balance in relation to elevation on the Inland Ice at Paakitsoq. (I) Transient balance: August 7, 1985 – May 13, 1986. (II) Transient balance: May 13, 1986 – August 25, 1986. (III) Transient balance: July 31, 1986 – August 25, 1986. (IV) Annual balance: August 7, 1985 – August 25, 1986. (Fig. 2 in Thomsen, 1987a).*

1987

The stakes were visited by helicopter on May 13 and August 13, 1987. Winter snow was patchy (as described for 1984) up to 500 m asl. and continuous at higher elevations. At elevations of about 500 m asl. the 1987 summer ablation is high compared to earlier years. At lower elevations the stakes were melted out, confirming strong melting during the summer. For these stakes only a minimum ablation given by the length of the stake still in the ice by the time of the spring visit (Thomsen, 1988a).

millimetres of water

Stake	Elevation m a.s.l.	25th Aug 86 13th May 87	13th May 87 13th Aug 87	25th Aug 86 13th Aug 87
2.0	210	−580	>(−2900)	>(−3500)
2.1	205	−850	>(−2900)	>(−3800)
2.2	205	−800	>(−2900)	>(−3700)
2.3	200	−740	>(−2800)	>(−3600)
2.4	200	−810	>(−2800)	>(−3600)
2.5	200	−560	>(−2900)	>(−3500)
3.0	235	−630	>(−2900)	>(−3500)
4.0	380	−570	>(−2700)	>(−3300)
5.0	415	−560	>(−3200)	>(−3800)
6.0	560	−100	−2080	−2180
7.0	615	−180	−1980	−2160
7.5	720	− 70	−2000	−2080
8.0	780	140	−2370	−2230
9.0	850	60	−2270	−2210
10.0	890	130	−1560	−1430
11.0	965	140	−1750	−1620
11.5	1020	140	−1580	−1430
12.0	1070	340	− 480*	− 130*

* estimated.

Table 3. *Transient and annual balance for the Inland Ice at Paakitsoq. Numbers in brackets are minimum values given by the length of a stake still in the ice on May 13, 1987, that has melted out of the ice.*

1988

The stakes were visited by helicopter on May 19 and August 18, 1988. Further stake readings were made with irregular intervals in August. Winter snow was patchy (as described for 1984) up to 700 m asl. and continuous at higher elevations. The snow surface showed signs of melting in May with water-soaked snow occupying depressions on the surface. However, little or no ablation occurred as there were no signs of drainage. Stake readings on August 18 show an annual equilibrium line at approximately 1050 m asl. (Thomsen et al., 1989a).

1989

The stakes were visited by helicopter on May 12 and August 14, 1989. Winter snow was patchy (as described for 1984) up to 500 m asl. and continuous at higher elevations, except at stake 8, where no snow was recorded. At elevations of about 600 m asl. the 1988/89 annual balance was close to the mean annual balance for the measurement period 1982–1989. Stake readings on August 14 show an annual equilibrium line at approximately 1100 m asl. (Thomsen and Olesen, 1990).

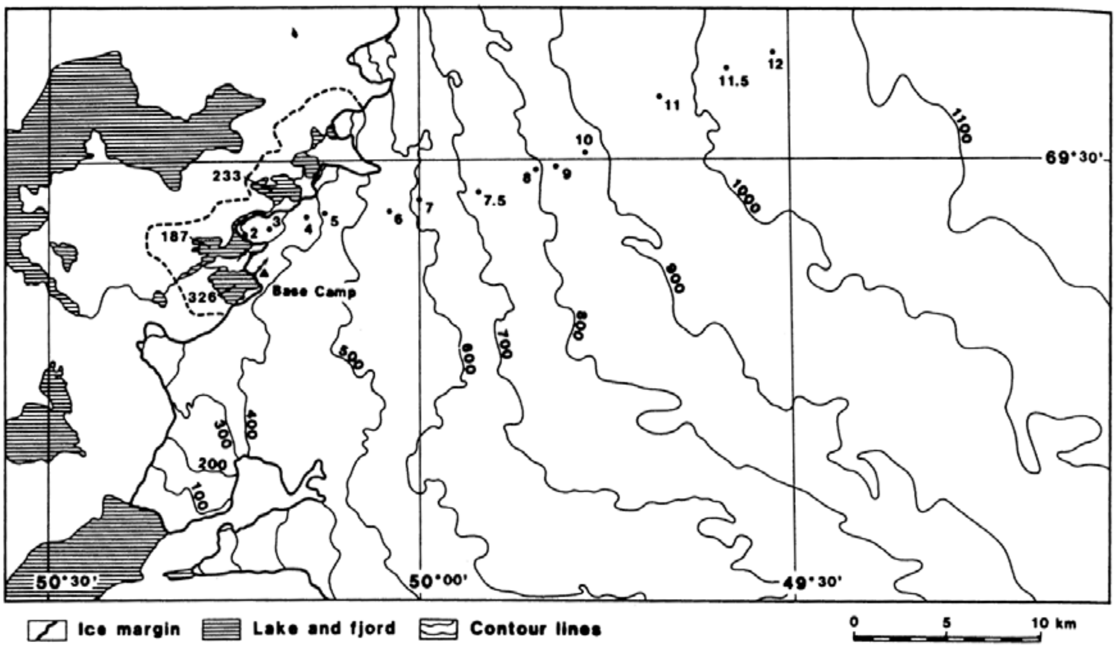

Figure 7. *Stakes at Paakitsoq in 1987, 1988 and 1989. Contours in metres. Ice sample profile near base camp given by line of arrows (Fig. 1 in Thomsen & Olesen, 1990).*

1990

The stakes were visited by helicopter on May 10 and August 15, 1990. Winter snow was patchy (as described for 1984) up to 500 m asl. and continuous at higher elevations. The measurements of mass balance from previous years showed the annual equilibrium line altitude to be at about 1100 m asl. The uppermost stake (stake 12), at an elevation of 1070 m asl., showed an annual balance of -621 mm for the period August 14, 1989, to August 15, 1990. This is the highest ablation recorded at this elevation since 1985 when measurements were started at this stake location. Extrapolation of the mass balance data to higher elevations indicates an annual equilibrium line altitude at about 1160 m asl. A reconnaissance was made for a future study of mass balance conditions in the lower accumulation zone at Paakitsoq. A stake net was established from 1100 to 1600 m asl., extending the existing stake line which has been measured since 1982 (Thomsen et al., 1991).

1991

The stake transect that had been maintained since 1982 was visited for the last time by GGU this year, but the results were not published. The entire set of mass balance data from all the stakes over the years are compiled at GEUS. The new stake transect established in 1990 in the lower accumulation zone was measured during May and August, 1991 (no specific dates given). Emphasis was on determining density profiles by digging snow pits and taking samples with a SIPRE corer. The density profiling, mainly using the SIPRE corer, was repeated in August 1991 but comparisons between May and August suggested that density determinations in August were too high. The reason stated as the most probable was compaction of the firn during cutting of the core or removal from the drill, which makes sample volumes too small. However, cores containing ice layers gave similar densities to those found by Braithwaite et al. (1982) on Nordbogletscher, South Greenland, and the mass balances were estimated using densities from only these cores (Braithwaite et al., 1992b).

Stake	Elevation m a.s.l.	Winter balance	Summer balance	Annual balance
Swiss camp	1160	+0.62	−0.57	+0.05
122	1190	+0.30	−0.66	−0.36
128	1250	(+0.4)	(−0.3)	(+0.1)
130	1260	(+0.6)	(−0.3)	(+0.3)
139	1350	+0.61	−0.26	+0.35
141	1380	(+0.5)	(−0.2)	(+0.3)
151	1450	(+0.6)	+ive	+ive
157	1500	(+0.5)	+ive	+ive
161	1550	+0.48	+ive	+ive
163	1620	+0.52	(+0.1)	(+0.6)
165	1640	+0.62	(+0.1)	(+0.7)

( ) = estimated balance.
+ive = positive balance.

Table 4. *Measured and estimated mass balances for 1990/91 in the lower accumulation area at Paakitsoq. Units are m water a-1 (Braithwaite et al., 1992).*

Ice-dynamic modelling

The only way to assess the future behaviour of the Greenland ice sheet in any detail is by setting up an ice-dynamic model, which can then be forced with past climate records and predicted scenarios for the future climate. Ice-sheet models have developed parallel with the computational possibilities available, but few models a well-suited to assess the behaviour of the ice-sheet margin in three dimensions. At Paakitsoq, the modelling is complicated by the fact that there is a relatively large calving outlet glacier (1GC04001) neighbouring the southern flank of the basin and on a larger scale by the position between the two icestreams Jakobshavn Isbræ and Eqip Sermia, which may exhibit even more unpredictable behaviour.

The only ice-dynamic modelling study done specifically to assess marginal changes on a time-scale of interest in hydropower investigations is by Reeh (1983). In this study, Reeh devised a non-stationary ice-dynamic model, specifically for calculating the advance and retreat of the ice margin on a time scale of 10 to 500 years, based on theory developed by Nye (1960, 1965a, 1965b). The various parts of the computer code was developed by N. Reeh, D. A. Fisher and D. Dahl-Jensen. The model was a two-dimensional flow-model that was applied to each individual sector of the ice sheet margin in the Paakitsoq basin. Each flow-line calculation was considered to be uncoupled from the results obtained from its neighbouring flow-lines. The bedrock topography was unknown apart from a single transect measured at the EGIG line just north of the Paakitsoq basin. To circumvent this problem, the bedrock topography had to estimated from a steady-state flow-line calculation carried out for the EGIG line where the bedrock topography was known and combine this with the known surface topography and calculated mass balance along each flow-line. The data sources available for the study of Reeh (1983) were rather sparse. Although some ice cores had been retrieved (Dye 3, Crête and Milcent) to estimate accumulation back in time, the various measurements of surface velocity, elevation changes and other relevant parameters were not abundant. Notably the lack of data on the bedrock topography would be a major limitation to any ice-dynamical modelling study.

A summary of the most important conclusions of Reeh (1983) is given below:

1. The various glaciers react differently on the same mass-balance history
2. Narrow glacier tongues terminating at low elevations react more strongly and with greater delay that broad glacier margins at higher elevations.
3. Large differences in the reaction exist also between the actual glacier tongues. As an example, glacier 1GE08001 reacts faster but not as strongly as glacier 1GC04003.
4. The calculated retreat of the glacier fronts since about 1880 is between 100 and 800 m and largest for the actual glacier tongues terminating at low elevations.
5. Within the last 1000 years, the glaciers had their maximum extent in the period 1650–1700 AD at which point the glacier fronts were advanced between 100 and 900 m compared to present day.
6. In Norse times between *c.* 1100 and 1450, the glacier fronts had retreated back to positions approximately as present day or significantly further back.
7. The response of glacier fronts and low-lying parts of the glaciers are dominated by changes in ablation rather than accumulation changes. This conclusion is limited to the time span considered here, ie. 10–500 years.
8. The influence of accumulation changes rises and becomes dominant in the upper part of the accumulation zone.
9. The model correctly predicts the elevation changes observed along the EGIG line in the period 1958–1968.
10. The glacier fronts must be expected to continue their retreat in the coming decades unless the climate suddenly becomes as cold as during the “little ice age” a few hundred years back.

Ice velocity measurements

In 1987 the ice velocity was measured on the glacier tongue ending in Lake 187. Ice velocity was measured by theodolite survey at stakes drilled into the ice from fixed points established on the ground. The position of the stakes is found in the section on hot-water drilling under *1987*. Ice movement were highest at stake 3 located at the foot of a small icefall. For all stakes there were a marked seasonal variation in ice movement, with mean summer velocities up to twice the mean winter velocity. Variations in sliding velocity could be an explanation which in turn implies that the basal ice is at the melting point. From depth soundings in Lake 187 and ice thickness measurements with radar, it was reasonable to believe that the glacier tongue was floating. Water level recordings in Lake 187 showed that the mean water level was 2 m higher in summer (GTO, 1983). The possibility that the movement pattern was connected with a floating ice tongue and thus was of local origin could not be excluded (Thomsen, 1988a).

metres per day

Stake	31st May 86 15th Sep 86	15th Sep 86 15th May 87	31st May 86 15th May 87
2.0	0.10	0.05	0.06
2.1	0.15	0.08	0.09
2.2	0.09	0.04	0.06
2.3	0.13	0.08	0.09
2.4	0.10	0.06	0.07
2.5	0.18	0.09	0.12
3	0.30	0.19	0.23
4	0.14	–	–

Table 5. *Ice velocities on the margin of the Inland Ice at Paakitsoq (Table 2 in Thomsen, 1988a).*

Photogrammetric and satellite mapping

According to Weidick (1968), the glaciers have experienced thinning since 1880. Thinning of glacier 1GE07001 between 1880 and 1959 has been estimated to 40 m near margin to about 20 m at an elevation of 350 m asl. (Thomsen, 1983b). Comparisons with surface topography maps based on aerial photography from later periods, namely 1959 and 1985, show a relatively constant thinning rate of approx. 1 m per year in the outermost margin. This trend has been confirmed by unpublished elevation measurements on the ice margin in 2003, 2004 and 2005 (Reeh, unpublished). The retreat of the ice margin is important, not so much because of the reduced glacierized area in the basin, but rather because the change ice thickness over a given bedrock topography will change the basin delineation and thus the run-off to a hydropower facility.

It is also desirable to obtain a detailed knowledge of the surface drainage on the ice sheet, as this determines where the meltwater actually enters the ice. It is entirely possible for meltwater generated within the delineated basin to be transported through a surface river system to a part of the ice sheet which lies outside the calculated basin delineation. For this reason, an aerial photography was acquired specifically for glaciological mapping purposes, ie. the photo was taken late in day with sunshine to produce a clear shadow marking of surface features. The result was a comprehensive glaciological map of the ice sheet in the Paakitsoq basin, published by Thomsen, Thorning and Braithwaite (1988), showing all surface drainage features, such as meltwater streams, moulins, full and emptied meltwater lakes, surface lineations, debris cover and crevasse fields. A separate map showing only the delineation of individual drainage cells draining into moulins was also produced to address the question of surface routing of meltwater and is shown in Fig. 8.

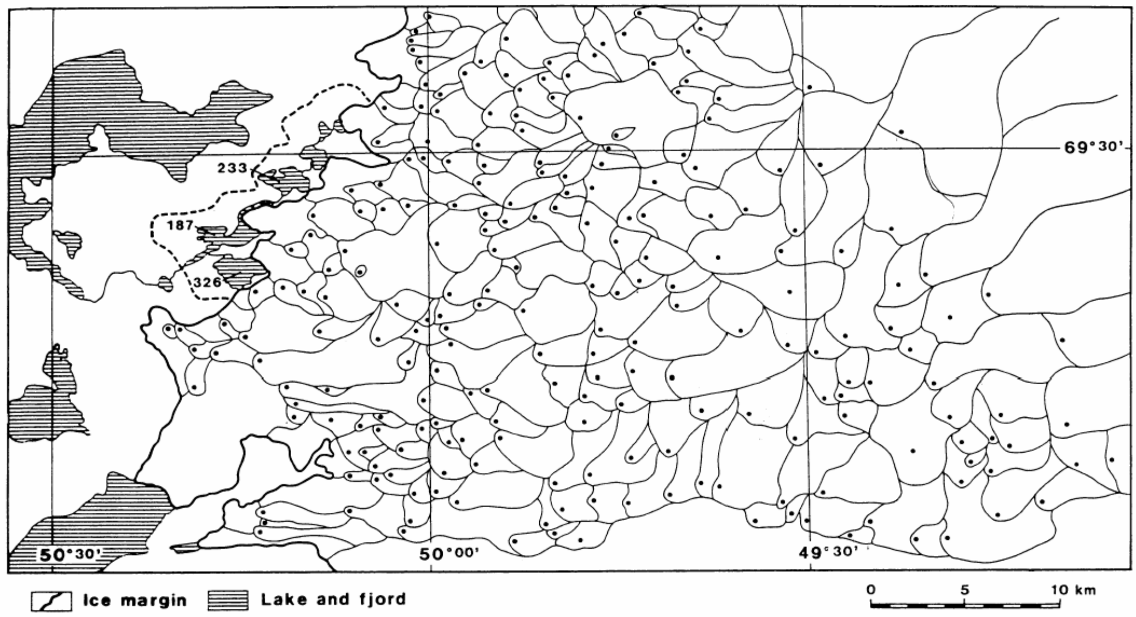

Figure 8. *Drainage cells, each draining into either a moulin or a moulin complex, whose positions are given by dots (Fig. 2 in Thomsen et al., 1989b).*

Ice-penetrating radar surveys

Knowing the bedrock topography beneath the ice sheet is crucial for two of the most important questions related to hydropower pre-feasibility studies:

1. Delineation of the hydrological catchment within and beneath the ice sheet
2. Ice-dynamic modelling of the ice-sheet response to climatic forcing

The ice-dynamic modelling carried out by Reeh (1983) was done from an extremely limited amount of information as to the bedrock topography, namely a single transect north of the Paakitsoq basin (the EGIG line). Instead, bedrock topography was derived from an assumed mass balance profile and other data sources. The calculated bedrock topography from this analysis relied on a number of assumptions and it was never even attempted to use this bi-product for delineation purposes. As basin delineation in ice-covered regions relies heavily on knowledge of the bedrock topography, it was decided to develop an ice-penetrating radar system at GGU in collaboration with the Electromagnetics Institute (EMI), Technical University of Denmark (DTU) (the later Ørsted-DTU, now Danish National Space Center, DTU). The first attempt was made in 1984, using a 300 MHz dipole antenna, mounted beneath a small twin engine fixed-wing aircraft (Twin Otter) with extra fuel tanks. The 1984 work was not successful, partly because of a component failure in the radar system, but possibly also due to the antenna configuration as described in Thomsen & Madsen (1985).

For the following year, 1985, the radar system was improved by the use of a new antenna, a two element fed cylindrical parabola installed in a helicopter (Bell 206 Jet Ranger) rather than a fixed-wing aircraft. The construction was carried out at EMI. Plans to use an alternative 60 MHz radar system had to be discarded due to the incompatibiity of the large antenna with the small helicopter. Flights were carried out during a five week period in July–August 1985. The tests had shown the altitude above the ice to be a decisive factor for the penetration of the radar, and thus the measurements were carried out approximately 10 m over the surface at an airspeed of 80–90 knots. The only means of navigation was visual sighting supported by standard flight instrumentation. This, together with the low altitude, severly limited the size of the area that could be surveyed and constrained the possible configuration of the lines, yielding a somewhat unorthodox flight pattern. Altogether 21 flight hours were used including time for transport of fuel into the area. The operation was very successful. Good refelctions were obtained along most lines, although difficulties were encountered over water-logged areas, large crevasses, and the most broken up glaciers. It was suggested that following measurement campaigns should be carried out earlier in the year to minimize the impact of meltwater on and within the ice sheet on the radar performance (Thorning et al., 1986).

Frequency	300 Mhz
Pulse effect	800 W
Repetition	0-15.5 kHz
Pulse length	0.25 microsec
Range resolution	24 m
Effect	max 1100 W
Weight	*c*. 70 kg

Table 6. *Radar system details (Table 1 in Thorning et al., 1986).*

To extend the data coverage towards the eastern parts of the ice-sheet basin and to obtain reflections in difficult areas still missing, a spring campaign was carried out over three weeks in April 1986 again using a Bell 206 Jet Ranger helicopter. A major improvement was the use of a Del Norte line-of-sight local navigation system borrowed from GTO, making it possible to operate further away from the ice margin. The low flying altitude of 10 m above the surface reduced the amount of accurate positions recorded to 80% of the data, but the remaining positions could be interpolated fairly accurately. For most of the data the reflections were more clearly defined and more continuous than in 1985 (Thorning & Hansen, 1987).

The data from the radar campaigns in July–August 1985 and April 1986 were digitized and migrated. A simple geometric migration was used assuming a constant wave velocity of 169 m/μs through the ice sheet. After a qualitative evaluation of the migrated data, the accepted data were averaged within 100 m windows along the profile and if possible the standard deviation was calculated. The averaged, migrated ice thickness data were finally gridded to a 100 by 100 m grid, including points along the actual ice margin with zero ice thickness. The gridding method was based on the minimum total curvature principle. Two-dimensional Fourier filtering was then applied to the gridded data. To produce a topographic map of the subglacial relief, a surface elevation grid was produced from a contour map based on aerial stereophotography from 1985 and added to the ice thickness grid (Thorning & Hansen, 1987).

In May 1987, yet another radar survey was carried out in key areas of the Paakitsoq basin, this time with an improved antenna and a new video recorder. The flight elevation above the ice surface was 10 m as in the previous years, and last years navigational system was also used again although with the four remote stations in a different configuration. The objective was to obtain data from two specific areas where problems in the hydrological interpretation existed (Thorning & Hansen, 1988a). In total, good reflections had been obtained over 70–80% of the distance flown in Paakitsoq in the period 1985–1987 (Thorning & Hansen, 1989).

In addition to the airborne radar survey, a new mono-pulse radar was tested on the ice surface with several sets of antennas for peak frequencies 40, 20, 10.5, 2.5 and 1.25 MHz. Each antenna was a flexible resistively-loaded, centre-fed dipole, identical and symmetric about the feed point. Radio wave velocities were chosen at 300 m/μs in air and 168 m/μs in ice. The mono-pulse radar was used in three areas. The first was near the first hole drilled with the hot-water drill as described in Olesen & Clausen (1988) yielding a depth of *c.* 300 m corresponding well to the airborne radar survey and the borehole logging. The second area was where the hot-water drilling ran into problems at a depth of 270 m, some 60 m above bedrock. Presumably, the material causing difficulties for the drilling also scattered the radar waves, as no reflected pulse could be recognized. The third area was on the glacier lobe of sector 1GE07001 extending into Lake 187. A cross-sectional profile and a center flow-line profile was obtained using the 5 MHz antenna, giving a total of roughly 2 km of ice thickness data with depths ranging from 120 to 190 m along the flow line. These depths were read from Fig. 5 in Thorning & Hansen (1988b) and do not correspond to the 70 to 50 m claimed in the text of the same paper (Thorning & Hansen, 1988b).

Site	Drilling m	EMR m	Difference m	%
8	298	300	2	0.7
3	305	300	5	1.7
2	270	240	30	11.1
2 A	278	240	38	13.7
1	382	400	18	4.7

Table 7. *Comparison of ice thicknesses derived from airborne radar survey depths and hot-water drilled bore hole (Table 2 in Olesen & Clausen, 1988)*

During five days in August 1988, the mon-pulse radar was applied to the glacier lobes ending in Lake 326 and 187. Most positions could be reached by foot and only a few helicopter lifts were necessary for the more distant working areas. All lines were measured with a 5 MHz antenna and with a transmitter-receiver distance of 90 m or, in one case, 100 m (Thorning & Hansen, 1989).

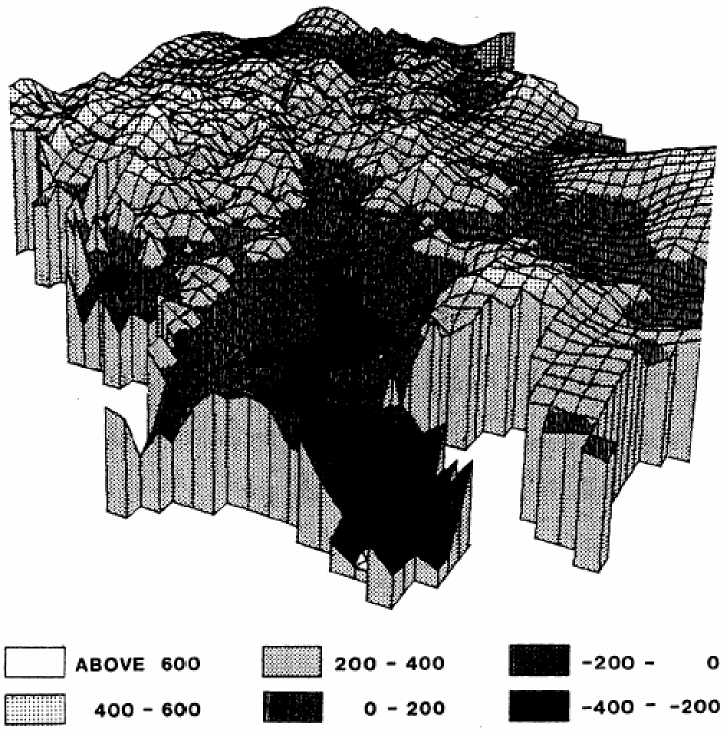

Figure 9. *A three-dimensional representation of the subglacial topography seen from the southwest (Fig. 3 in Thomsen et al., 1989b).*

Hot-water drilling – subglacial water pressure

Drilling a hole through the ice sheet to the bedrock makes it possible to measure directly a range of important quantities: An exact point measurement of ice thickness, the englacial temperature from frozen-in thermistor strings, and finally the subglacial water pressure if the drilled hole connects to the drainage system. This section deals with the latter of these, as the first was mentioned in the previous section and the second will be discussed in the following section.

The subglacial water pressure is an important factor for the drainage configuration and thus also for basin delineation. To facilitate direct measurements of the subglacial water pressure, a new hot-water drill was developed at GGU and tested on the ice-sheet margin at Paakitsoq (Olesen & Clausen, 1988). The first drill was destroyed in the 1988 season, because it was accidentally dropped from a helicopter sling load, but was replaced by a new, improved drill the following year (Olesen, 1989). The drill was modelled over the Swiss system used by the Swiss Federal Institute of Technology (ETH) in Zürich (Iken et al., 1977). The Swiss group at ETH Zürich drilled through Jakobshavn Isbræ just south of Paakitsoq (Iken, 1990). Hot-water drilling through ice is a fast and reliable and widely used method if a constant borehole diameter is not important and no ice core is required. The basic principle consists of pumping water through a heating system and into a hose with a rigid drill tip with a nozzle. The hot water from the nozzle melts the ice in front of it and flows back up the drill hole.

The total weight of the first complete GGU hot-water drilling system with tools and spare parts was 435 kg. This included a power/piston-pumping unit, a heating unit and 500 m of heat resistant high pressure hose, a 2 m long 25 mm diameter stainless steel drill tip with interchangeable nozzles and clinometer on top, a lightweight tripod with winch and pulley, and a low pressure pumping unit for use when water had to be drawn from farther away. The heat loss was about 2% per 10 m of hose (Olesen & Clausen, 1988).

During the seasons 1988, 1989 and 1990, holes were drilled through the ice to the bottom and water pressure was measured, either sporadically or as continous records over short time spans (less than a week). The records generally show high water pressure with maximum and minimum recordings in the holes ranging from 105% to 79% of the ice overburden pressure. Furthermore, the data indicate that water pressure expressed as a percentage of the ice overburden pressure decreases with increasing ice thickness (Thomsen and Olesen, 1991). A brief summary of the drilling campaigns carried out by GGU is given below for each year.

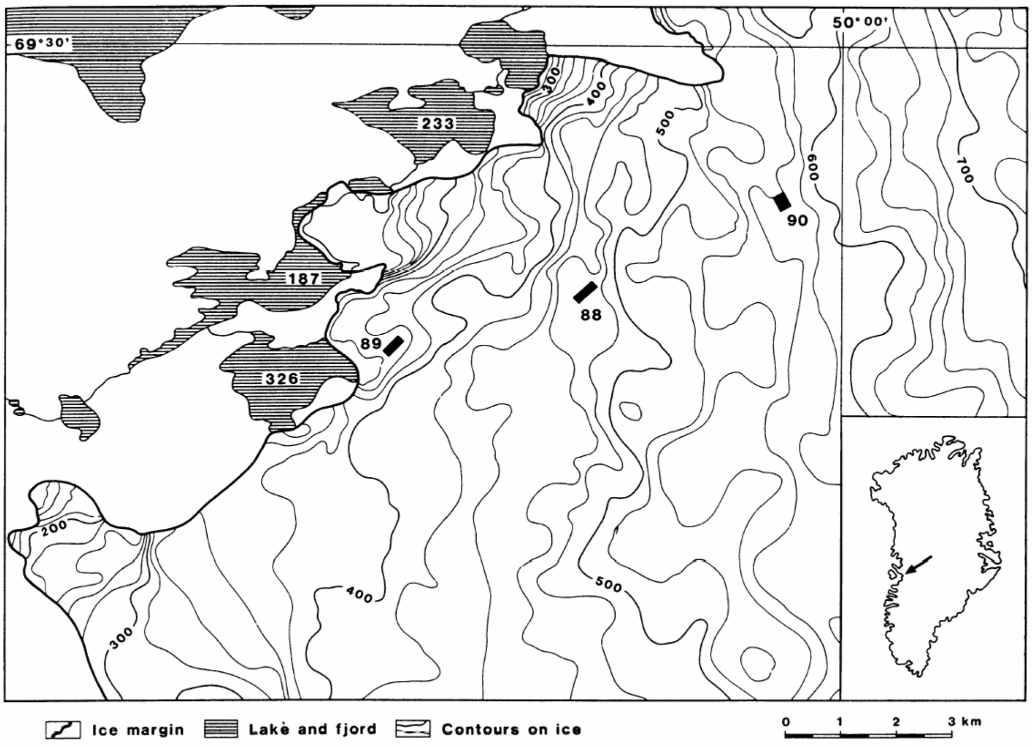

Figure 10. *Part of ice sheet sector at Paakitsoq, West Greenland. Lakes shown with horizontal lines. Contour interval on ice 20 m; no contours shown on land areas. Locations on the ice are given where measurements of subglacial water pressure were obtained; numbers (88, 89, 90) indicate year of measurements (1988, 1989, 1990). (Fig. 1 in Thomsen & Olesen, 1991).*

1987

The first hot-water drill was tested on the ice sheet margin, Eight deep holes were made to depths between 270 and 383 m totalling 2436 m and the ice-sheet thickness was measured on five of these positions where bedrock was reached (Olesen & Clausen, 1988). Drill positions were all near the base camp, approx. 4 km East of Lake 326 and Lake 187 (Thomsen, 1988a).

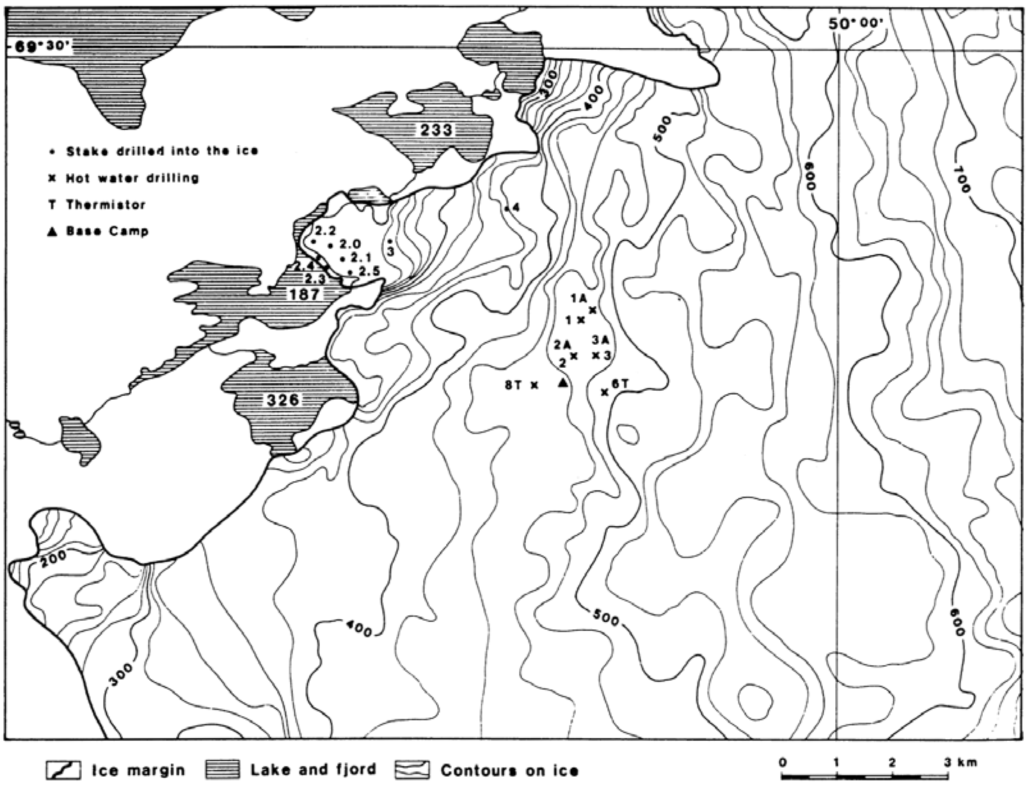

Figure 11. *Part of the Paakitsoq drainage basin showing locations of stakes, drill sites and thermistor strings. (Fig. 2 in Thomsen, 1988a).*

1988

Nine holes drilled, totalling 3220 m. All holes were drilled near base camp 4.2 km east of Lake 187, along a profile crossing a subglacial valley leading to Lake 187. An improved inclinometer and a new device for weighing the load on the hose were tested. Penetration to bed was difficult due to debris layers. Sudden drops of water level were observed in the drill holes during drilling, indicating the existence of voids or englacial conduits. Explosives were used for blasting the bottom of the boreholes to force a possible connection to the hydraulic system at the bed. Small meltwater streams were directed into the drilled holes to prevent them from freezing, but this did not work too well. Few measurements of water levels were achieved as the pressure sensors failed and the boreholes froze up. The readings obtained, pointed at a high water pressure. During the field work, the drill was dropped from the helicopter sling and destroyed (Thomsen et al., 1989a).

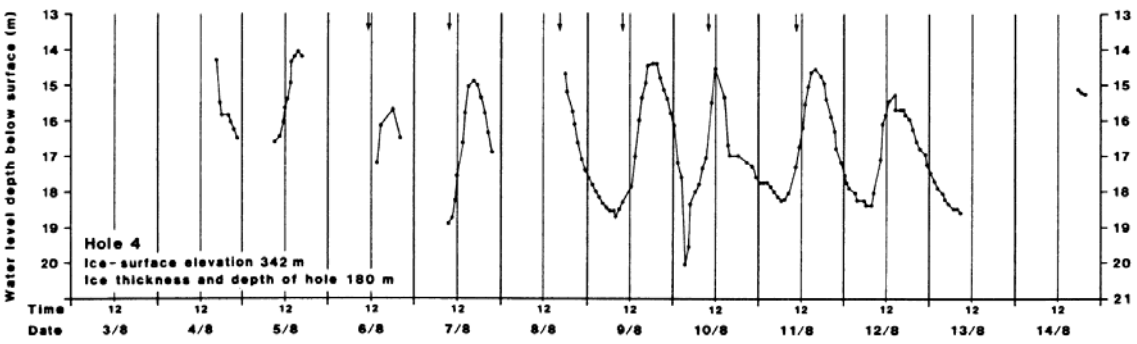

Figure 12. *Example of water-level recording in a borehole, obtained in 1988. Vertical arrows show times when water was pumped into the hole to prevent freezing (Fig. 2 in Thomsen & Olesen, 1991).*

1989

A new drill was taken into the field. Despite minor technical problems with the new drill, nine holes were drilled with a total length of 1872 m, the deepest being 520 m. Eight holes were drilled near the base camp along two profiles on the glacier tongue leading to Lake 326. The base camp this year was 600 m from Lake 326 on the glacier tongue. More exact positions of drill sites were not given. All holes seemed to connect to the englacial or subglacial drainage system. Water level was recorded in all holes either manually or by dataloggers connected to pressure sensors. In four of the holes continous readings of water level fluctuations were obtained, while the remaining holes yielded only scattered data due to problems with freezing. The continous recordings showed a marked diurnal fluctuation with higher water levels around 16.00–19.00 hours and minima at 7.00–9.00 hours local time. The measurements show a subglacial water pressure close to the ice-overburden pressure (Thomsen & Olesen, 1990).

1990

Eight deep holes were drilled, totalling 2963 m. Two of these holes were drilled at higher elevations (935 and 1140 m asl.) to depths of 600 m and 500 m, respectively, without reaching bedrock. The remaining six holes were all drilled near the base camp, which was situated on the ice sheet at an elevation of 560 m asl. about 8 km from the margin east of Lake 233 and Lake 187. Three of these holes were drilled on a local ridge on the ice surface to depths of 354 m, 254 m and 217 m. The first was believed to reach the bottom, whereas the two others probably ended in a debris layer. There was no draining in the boreholes, not even after detonation of explosives in the 354 m hole. The last three holes were drilled in a nearby local depression in the ice surface, all reaching the bottom at depths of 341 m, 347 m and 350 m, respectively. In all cases, the water level dropped several metres when the drill stopped advancing at the bottom of the ice. Despite leading surface streams into the holes, only the 350 m hole remained open. The water level in the 350 m hole was recorded after stopping the water supply, both manually from the surface and by a pressure sensor installed in the hole and connected to a data logger. Air temperature was recorded at the same site two metres above the ice surface. Slightly more than three complete diurnal cycles were recorded, showing a marked diurnal oscillation in water levels, with higher levels at 20.00–22.00 hours and minima at 12.00–13.00 hours. Maximum and minimum registrations in the continous record correspond to a basal water pressure of 94% and 85% of the ice overburden pressure. This was comparable with the measurements made further downstream in 1988 and 1989. The water level variation seemed to follow the variation in temperature but with a delay of 2 to 5 hours between maximum air temperature and maximum water level. Two single measurements in the borehole, made while a stream was flowing into the hole, oddly enough showed the lowest water levels recorded, without any obvious explanation (Thomsen et al., 1991).

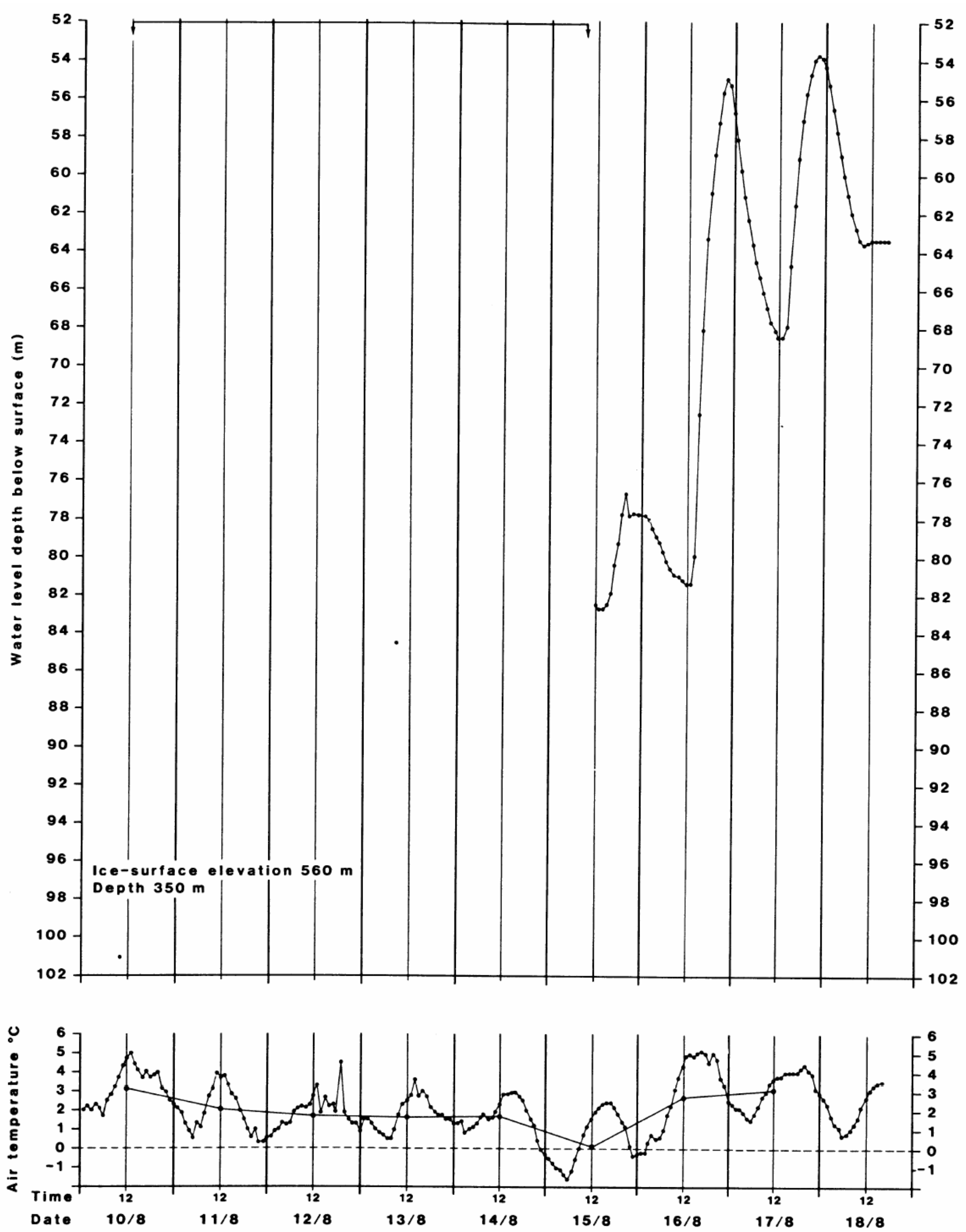

Figure 13. *Upper diagram shows water level recordings in borehole, August 1990. Vertical arrows connected with a bar indicate period when surface water was led to the borehole. Lower diagram shows measured air temperature at two metres of elevation. Straight lines are daily mean temperatures. (Fig. 2 in Thomsen et al., 1991).*

Ice, firn and lake water temperature

It is desirable to know the temperature of the ice down from the surface to the bottom for several reasons. The most shallow part of the ice sheet (*c.* 0–20 m) plays an important role in determining the amount of meltwater that is allowed to run off the surface. The surface layer, whether it is snow/firn or bare ice interacts with the atmosphere through the surface energy balance and thus remembers the cold temperatures it experienced during the winter. This cold content has to be elimininated before run-off can occur. Cold firn or cold surface ice (ie. below the freezing-point) also causes refreezing of meltwater percolating through a snowpack, forming super-imposed ice. This refreezing complicates mass balance measurements and calculations, and raises the temperature of the underlying ice considerably.

The temperature of the whole body of ice is essential for ice-dynamic modelling. The ice temperature has a strong non-linear effect on ice flow, which is most pronounced in the range −10°C to 0°C. The 'warmer' the ice, the more readily it flows. It is common to set up a coupled set of equations to describe the evolution of the ice-dynamics in parallel with the temperature evolution. Directly measuring the ice temperature in the ice column contrains the solution of the thermo-mechanical ice-dynamic model, making it considerably more reliable.

Finally the temperature at the base of the ice sheet is important to determine the degree of basal sliding in the ice flow as well as the existence and type of subglacial drainage system. If the temperature is well below the freezing point, the ice sheet will be cold-based, ie. frozen to the bed, implying no-sliding conditions and also that a widely distributed subglacial drainage system is unlikely. Large meltwater rivers might possibly still carry enough energy to keep a large conduit open, but this issue is still debated.

A simple method to obtain the englacial temperature is to insert thermistor strings into boreholes and then leave them to freeze in. After a period of a few weeks, the ice temperature has regained equilibrium with the surrounding ice, allowing measurement of the ice temperature.

1987

Two sets of thermistor strings were drilled down to depths of 202 m and 300 m in the ablation area (TD1 and TD2 in Fig. 14). The 300 m hole which extended to the bottom was situated 3.2 km upstream from the ice margin terminating in Lake 326, at an elevation of 455 m asl. The 202 m hole was situated 4.4 km upstream from the ice margin along the same flow line, at an elevation of 490 m asl. Ice thickness at this site was expected to be *c.* 300 m from a previous EMR survey (Thorning & Hansen, 1987) but later proved to be 470 m (Thomsen et al., 1991). Thermistors were mounted at every 25 m on the string except for the lower end where the distance between the three lower thermistors were 10 m and 15 m, respectively. The accuracy was given as ±0.2°C. Temperature readings were made several times during the two weeks duration of the field programme and were read five weeks later by a visiting team from GTO. Experience shows that an equilibrium temperature should be reached after 2 to 3 weeks (Blatter, 1985). The temperature readings revealed negative temperatures throughout the ice body with a minimum temperature of −2.1°C and a maximum temperature of −0.6°C. The temperaute measured at the bottom of the ice was −0.9°C (Thomsen, 1988a).

1988

The temperature measurements made in 1987 were repeated and confirmed to be in equilibrium. An additional thermistor string was installed higher up on the ice sheet at stake 7 at an elevation of 615 m asl. (Thomsen et al., 1991), 9.5 km upstream of the ice margin terminating in Lake 326 (labelled TD3 in Fig. 14). More installations were planned further inland (stake 11), but the ice drill was lost during helicopter operations as described in the previous section (Thomsen et al., 1989a).

1989

Temperature readings were made with the thermistor string installed at stake 7 in 1988. The temperature readings revealed slightly negative temperatures from −0.1°C to −0.3°C through the ice body from the bottom, decreasing to −2.1°C in the upper fifty metres. In continuation of the 1988 work it was planned to install further thermistor strings at higher elevations on the ice, one at stake 11 and one at 1150 m asl., respectively 23 km and 41 km upstream from the ice margin. The 1150 m asl. position had to be given up due to half a metre of slush on the surface. At stake 11, a 520 m deep hole was drilled, but a layer at 175–250 m causing difficulties for the drilling also caused the thermistor string to stick, finally making it necessary to cut the cable (Thomsen et al., 1990).

1990

Thermistor strings installed in deep holes in previous years were remeasured, two new deep thermistor strings were installed (TD4 and TD5 in Fig. 14) and three new thermistor strings were installed in shallow holes (TS1, TS2 and TS3 in Fig. 14). The shallow thermistor strings were installed to detect the penetration depth of the zero-degree isotherm. Three deep holes were already equipped with thermistor strings from previous years; TD1 and TD2 installed in 1987 and TD3 installed in 1988. The ice temperature readings corresponded closely to earlier readings, revealing slightly negative temperatures throughout the ice body with a minimum temperature of −2.1°C. A new thermistor string was installed at TD4 at an elevation of 965 m asl. and to a depth of 500 m into the ice (drilling was continued to 600 m depth without reaching the bottom). Similarly, TD5 was installed at 1140 m asl. at the ETH Camp/Swiss station, reaching a depth of 600 m. Furthermore a 50 m thermistor string was inserted here for the Swiss team for mapping of the shallow depth ice temperatures. In addition to the deeper installations, three shallow thermistor strings were installed. TS1 was drilled to 10 m depth close to the ice margin near Lake 326. Ice temperatures were recorded every six hours in the period between May 12 and August 15. Taking into account ablation during the recording period and the accuracy of the temperature sensors (±0.2°C), the zero-degree isotherm penetrated to a depth between 0.7 and 1.2 m below the surface by August 15, 1990. TS2 and TS3 were drilled to a depth of 14 m at elevations of 615 m asl. and 890 m asl., but temperature readings were postponed to 1991 (Thomsen et al., 1991). Apparently, the readings planned for 1991 were never carried out.

However, the activities of the ETH Greenland Expeditions 1990–1991 may include some information on englacial temperatures at ETH Camp/Swiss station (Ohmura et al., 1991).

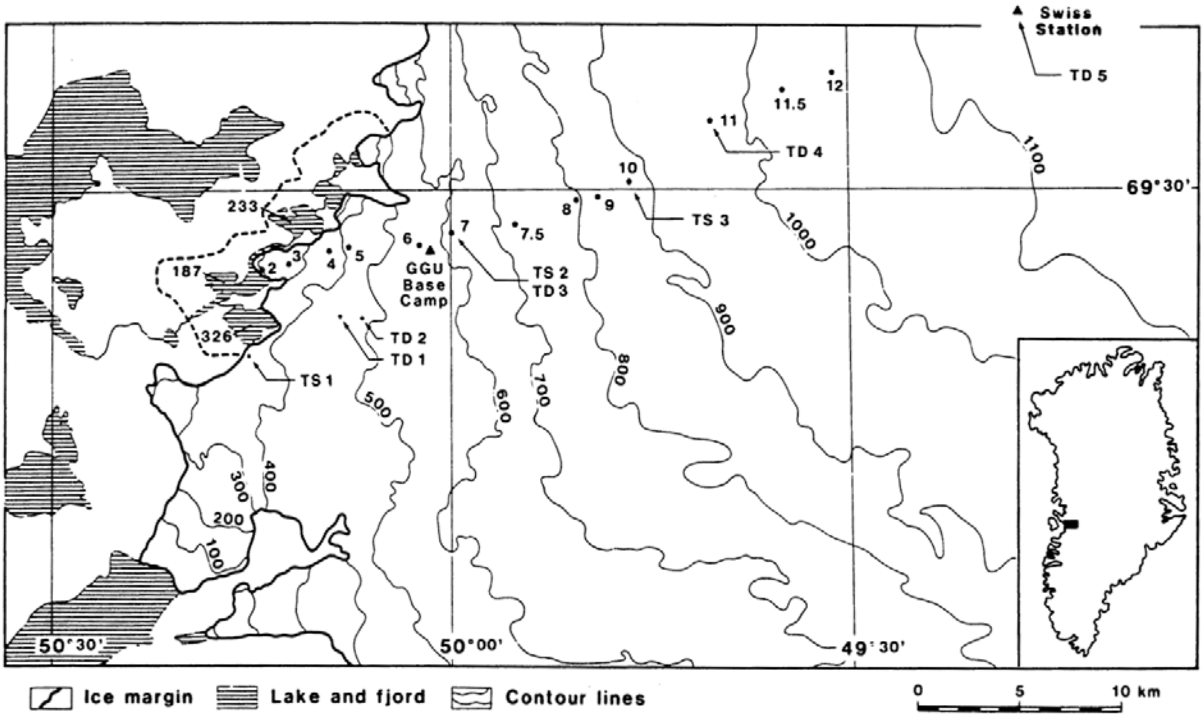

Figure 14. *Drainage basin at Paakitsoq showing stakes for measuring mass balance as well as locations of shallow depth (TS) and deep (TD) thermistor strings used for measuring englacial temperatures. Contours are in metres. (Fig. 1 in Thomsen et al., 1991).*

1991-1992

These field seasons consisted of a dedicated programme to estimate refreezing of melt-water in the firn (mulityear snow). As part of this programme, firn temperatures were measured at a the stakes shown in Fig. 15. The temperature profiles obtained at these sites a listed in Table 8. Mass balance for 1990/1991 was also measured at these stakes but is presented in the section on mass balance.

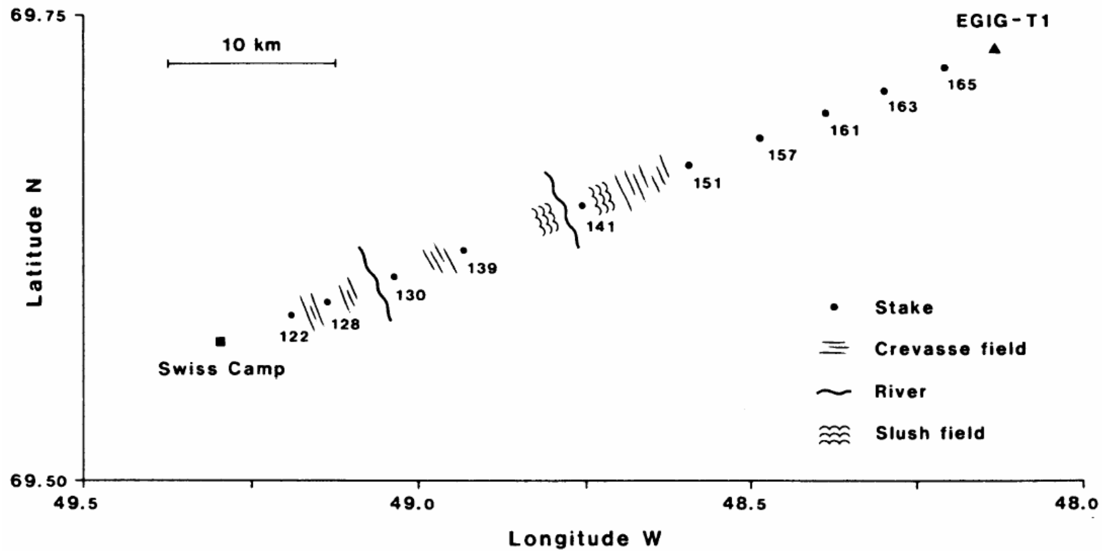

Figure 15. *Stake locations, hydrological features and visible crevasses in the lower accumulation area, Paakitsoq, as seen in August 1990. (Fig. 2 in Braithwaite, 1993).*

Stake	Date	m a.s.l.	Depth below surface – metres 1	2	4	6	8	10
122	30/7/92	1180	m	m	−9.4	−10.9	−10.5	−10.0
151	1/8/92	1440	m	m	−8.5	−9.7	−9.3	−9.0
157	19/8/91	1510	−0.1	−0.5	−3.2	−7.5	−9.6	−9.9
	14/5/92		−14.7	−13.8	−11.8	−9.8	−9.1	−9.0
	1/8/92		−8.1	−10.2	−11.3	−10.7	−10.1	−9.6
161	15/5/92	1530	−15.6	−14.7	−13.8	−11.7	m	−9.2
	2/8/92		−7.0	−9.4	−11.4	−11.2	m	−10.2
163	19/8/91	1620	0.0	−0.1	−4.0	−9.7	−12.4	−12.8
	11/5/92		−16.1	−15.2	−13.2	−11.8	−11.3	−11.5
	2/8/92		−10.7	−12.3	−13.0	−12.8	−12.2	−12.1
165	13/5/92	1620	−16.3	−15.5	−13.7	−12.4	−11.8	−12.0
	2/8/92		−10.9	−12.5	−13.5	−13.2	−12.7	−12.5

Table 8. *Temperature profiles in the lower accumulation area, Paakitsoq. Units are °C. (Table 2 in Braithwaite, 1993).*

1985 Lake temperatures

The temperature profiles in meltwater lakes at various elevations were measured in 1985 and reported by Thomsen & Reeh (1986). The estimated size of the lakes ranged from about 0.1 × 0.1 km to about 1 × 1 km with depths varying from 1.2 m to 4.2 m. No variation in temperature with depth was found and all lake temperatures were found to be above 0°C. The temperatures increase with decreasing elevation from 0.1–0.2°C at about 1000 m asl. To 0.8°C at about 350 m asl. As no obvious relation between water volume in the lake and temperature exists, the above trend can probably be related to length of melting season and the temperature of the underlying ice.

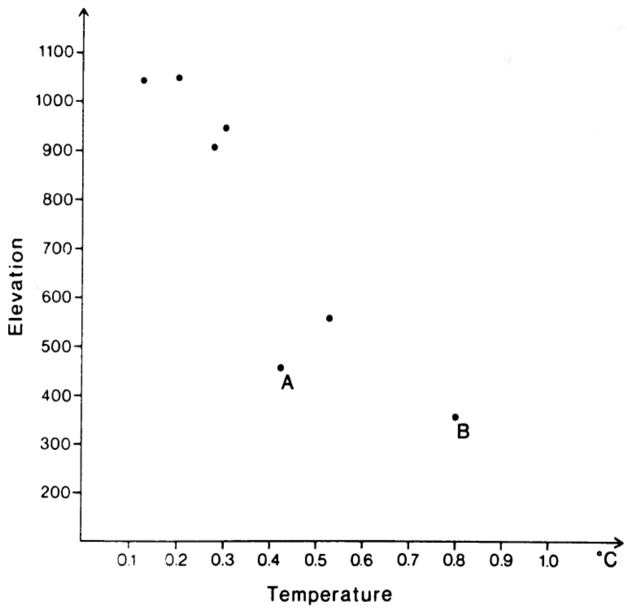

Figure 16. *Measured lake temperatures versus elevation on the Inland Ice at Paakitsoq. A – Meltwater flow through lake. B – Weak meltwater flow through lake. (Fig. 3 in Thomsen & Reeh, 1986).*

Basin delineation

Delineation of a hydrological basin within an ice sheet is a quite complicated task. Even if major drainage features of the ice sheet margin could be expected to be semi-permanent, or at least recurring, the change in capacity and through-flow of meltwater will in unison cause variations in the subglacial water pressure, which in turn affects the hydrological potential within the ice body and thus the desired direction of the water within and beneath the ice. The idea of a hydrological potential driving the water within a uniform and pervasive englacial/subglacial drainage system was formulated by Shreve (1972) and is still commonly used. To calculate the potential, it is necessary to provide the relationship between the subglacial water pressure and the overburden pressure exterted by the ice, expressed as the hydraulic parameter k. The value of k may vary between $k = 0$, corresponding to a situation where the subglacial water pressure is atmospheric, and $k = 1$, corresponding to flotation pressure potentially causing the ice sheet to disconnect from the bedrock.

As the hydrological potential depends so much on the subglacial water pressure, expressed through k, then so does the basin delineation and thus ultimately the amount of water available for hydropower purposes. For the Paakitsoq basin, a range of basins were calculated using the theory of Shreve (1972) and published in Thomsen et al. (1988) and more comprehensively in Thomsen et al. (1986) along with maps of the hydrological potential. An example of a hydrological potential and the derived basin delineation is shown in Fig. 17.

The initial basin delineation identified regions within the ice sheet basin, which could shift depending on the subglacial water pressure. It was attempted to reveal the pathway of the water by injecting dye into moulins on the ice sheet surface, and then subsequently identify the outlet by analysing meltwater appearing at the ice margin for traces of dye. These attempts were not successful, mainly because the meltwater would in most places appear underneath the ice directly into large lakes adjoining the ice margin.

The possibility of using a natural tracer, namely the ratio between the heavy and the light oxygen isotope in the meltwater, was also investigated. This ratio depends on a range of factors connected to the transport and deposition altitude of the precipitation, but is also dependent on the temperature at the time of the snow deposition. The bottom line is that the ratio varies in a predictable way for both snow and ice within the ice sheet basin as confirmed by Reeh & Thomsen (1986). This knowledge can theoretically be used to reveal information about the origin within the basin of the meltwater appearing at the ice margin. However, the problems with the lakes at the margin were similar to those from using injected dye. As a side-effect it was discovered that the ice margin at Paakitsoq had preserved the layering of annual snow fall apparent in ice cores from the central parts of the ice sheet. The relatively well-preserved ice marginal record has lately been used for taking out large samples (tonnes) for paleaoclimatological analysis.

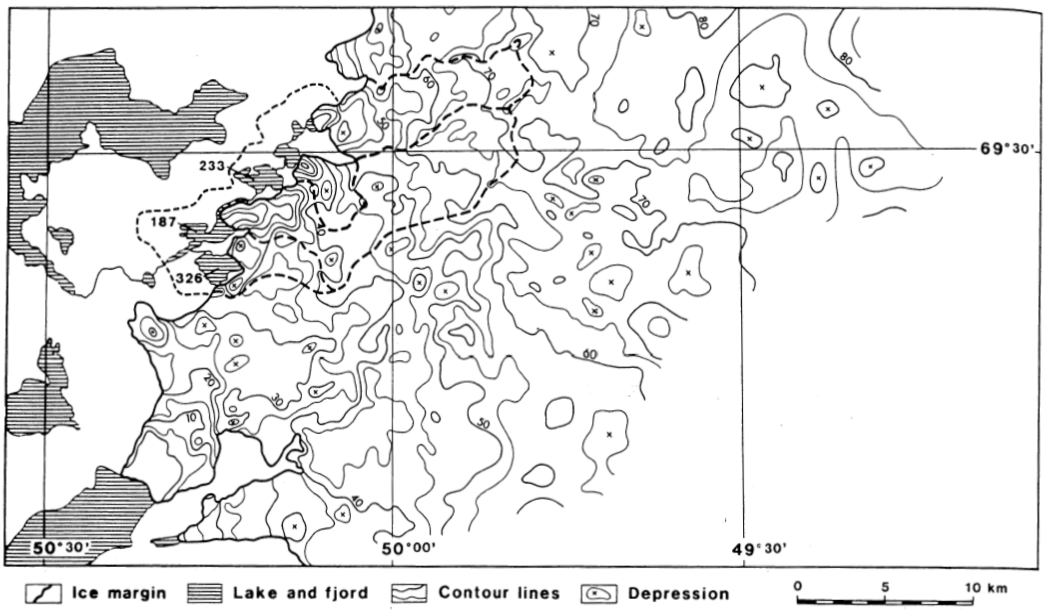

Figure 17. *Calculated subglacial water potential for k=0.7. Units are 105 N/m2. Subglacial water divides are shown by heavy dashed lines (Fig. 1 in Thomsen, 1988b).*

Run-off simulations

The run-off from the ice-sheet provides the vast majority (>90%) of the water available for hydropower in the Paakitsoq basin and it is therefore of primary interest to be able to predict future variations in the meltwater production. The melt modelling carried out for the

Paakitsoq basin by the GGU was done using the degree-day approach which exploits the strong correlation between the amount of time with positive air temperature and the surface melting. Results from these efforts, carried out in 1986, were published in Braithwaite & Thomsen (1989). Mass balance and specific run-off was calculated as functions of elevation within the basin, using extrapolated monthly data for air temperature and precipitation from Ilulissat. Monthly totals for the different balance elements were summed to give annual values at each elevation for an assumed hydrological year from September to August. The specific run-off was calculated as the sum of annual ablation (of ice and snow) and rainfall after allowing for possible effects of refreezing. Temperatures recorded at Ilulissat were extrapolated to obtain values for the ice sheet by taking account of the inland heating effect between coast and ice sheet, the vertical temperature lapse, and the cooling effects between ice-free and ice-covered areas. Monthly ablation of ice and snow at each elevation was assumed to be proportional to the corresponding monthly positive temperature sum, which was calculated from the monthly mean temperature value. The relationship, illustrated in Fig. 18, strictly only valid for ice was applied due to the sparsity of the snow cover. Either snow or rain was added to the ice sheet surface, with the choice depending on the probability of sub-freezing temperatures in the month considered. Simulated meltwater and rainfall was assumed to be absorbed by refreezing into any snow cover until the snow density reaches a specified critical value. The simulated run-off line was found to occur at about 1500 m asl.

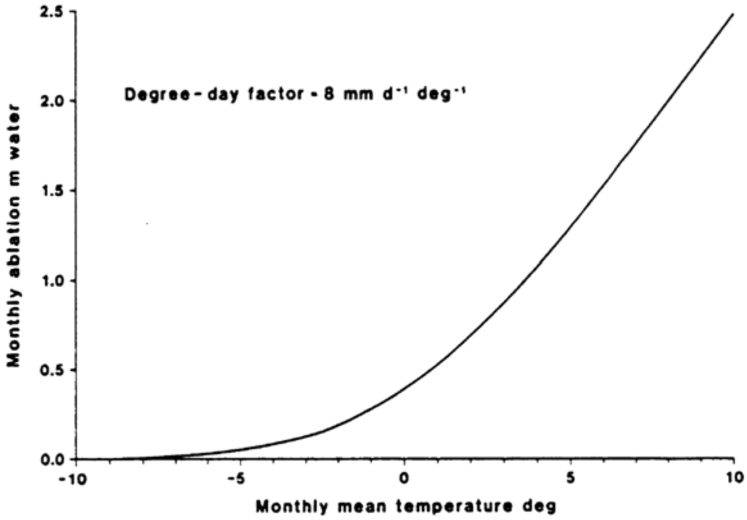

Figure 18. *Monthly ablation as a function (assumed) of monthly mean temperature (Fig. 3 in Braithwaite & Thomsen, 1989).*

The calculated specific run-off from the mass balance calculation was integrated over the assumed glacier area and added to the small contribution from the ice-free part of the basin to provide the combined basin discharge. The run-off was calculated for 25 different basin configurations, corresponding to different values of the hydraulic parameter k in the basin delineation (where k represents the relationship between basal water pressure and ice-overburden pressure at the base of the ice sheet). The calculated annual run-off volumes for 1980–1985 ranged from 337×10^6 m^3, for k = 1.0, to 227×10^6 m^3, for the most pessimistic interpretation with k = 0.0. This corresponds to a range of +17% to −21% of the mean observed run-off of 289×10^6 m^3. the best agreement between run-off observations and model was for k = 0.7. The change in run-off due to changes in the basal water pressure, expressed as variations in k, were thus found to be considerable.

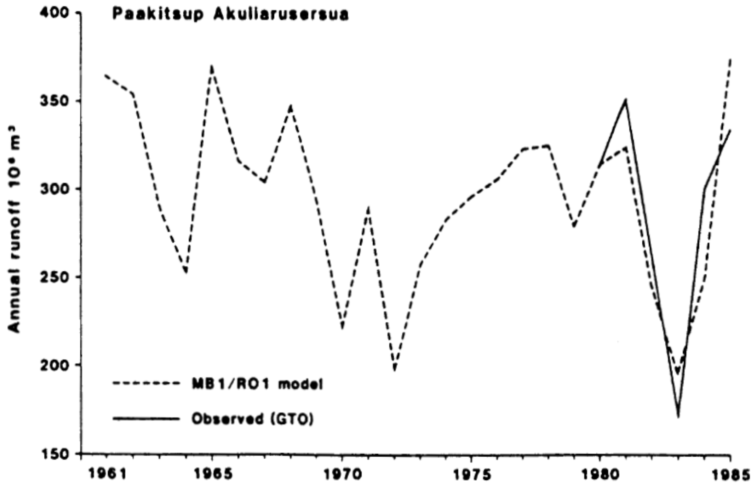

Figure 19. *Simulated run-off variations in the Paakitsoq basin for a 25 year period (Fig. 7 in Braithwaite & Thomsen, 1989).*

Conclusion and future perspectives

It is important to state that some of the most important glaciological problems pertaining to hydropower planning have not been solved for the Paakitsoq basin. Notably the future evolution of the basin size and meltwater availability have not been adequately predicted. This was simply not possible with the modelling tools available at the time of the previous investigations of the GGU. However, the extensive amount of data collected through this and following efforts in the Paakitsoq basin, greatly facilitates a re-evaluation of the future hydropower potential using recently developed ice-dynamic and melt models as well as the latest knowledge in basin delineation. The following paragraphs will describe the various new sources of knowledge and data that could be utilized in a re-evaluation of the hydropower potentail of the Paakitsoq basin:

New glaciological and climatological data

Since the GGU ceased to be active in the Paakitsoq basin roughly 15 years ago, there has been a continued effort to collect glaciological and climatological data, initially by the Swiss and shortly after by the US-based research groups. These data are available through the National Snow and Ice Data Center (NSIDC) in Colorado, USA. With some care, the datasets of GGU and NSIDC can be combined to form a 25 year long unique time series of ice-marginal mass balance in the Paakitsoq basin. Together with the discharge and climate data collected by GTO and later ASIAQ at the outlet of the Paakitsoq basin since 1980, these data represent a strong tool for calibration of the latest generation of melt models, developed specifically for estimating runoff from highly glaciated basins (Hock and Holmgren, 2005; Hock, 2005).

New surface and bedrock topographical data

In 2005 the Commission for Scientific Research in Greenland supported the collection of a new dataset in Paakitsoq basin consisting of surface topography from laser altimetry and ice thickness from ice-penetrating radar. The data was collected from a Twin Otter fixed-wing aircraft and benefitted from being an add-on project to the EU project 'Space borne measurements of Arctic Glaciers and implications for Sea Level'. This dataset is available in raw form for interpretation and would complement the existing data collected in the 1980s in two ways: In terms of bedrock topography, it would extend the area surveyed and by increasing the level of detail within the critical parts of the basin and in terms of surface topography, it would provide crucial information on the ice-marginal thinning during the last 22 years (since the last map of 1985).

New satellite data

The last 20 years has witnessed significant progress with respect to satellite data useful for, and available to, earth observation issues. A number of these would be useful in a revised analysis of the Paakitsoq basin. Notably information about the albedo of the ice-sheet surface in the period since 1985 from the US AVHRR Polar Pathfinder project, and later on the MODIS sensor 16-day average albedo product, both readily available for inclusion in melt models. The methodology of including this type of data in melt modelling has been matured during the last decade (Ahlstrøm, 2003). Another useful data source is the high-resolution imagery from e.g. Landsat, SPOT, ASTER and IKONOS sensors, ranging from 30 m to 1 m pixel-resolution. This sort of data makes a temporal mapping of the ice-marginal retreat/advance feasible and approaches the quality of aerial photography. It is also possible to generate elevation models from the latter two of the sensors mentioned, with quality depending on the users ability to provide accurate ground-control points (GCPs). It is possible to generate virtual flights through the landscape by combining visual and elevation data, providing a strong visualization tool for presentation purposes. Another interesting remote sensing product is available from passive microwave sensors such as SSM/I and AMSR-E, which can be used to map the temporal evolution of the melt extent above the snow line. Finally, it has become feasible to derive the surface velocity and elevation of entire regions of the ice sheet from active microwave (radar) sensors such as ERS-1/-2 and RADARSAT-1 by repeat-track synthetic aperture interferometry, providing a very strong boundary condition for ice-dynamic modelling.

New melt and climate models

The field of mass-balance modelling as experienced a huge development in parallel with the increasing computational power available (Ahlstrøm, 2003). The intensified interest in the melt processes of the Greenland ice sheet and other ice masses in connection with global climate change has led to an better understanding of the particular problems applying to the assessment of future runoff. Similarly, the regional climate modelling has now reached a level of detail where the distributed model run output can be used as input for the more specialized melt models on catchment scale. Specifically, the Danish Meteorological Institute has recently completed a regional climate simulation with a grid resolution of 10 km for the next 100 years. Even if there is a number of scaling issues to be dealt with before the results can be incorporated in melt models, both the temporal and the spatial scales are for the first time applicable to hydropower pre-feasibility studies as in the Paakitsoq basin.

New ice-dynamic models

The ice-sheet margin poses a difficult problem for ice-dynamic models, due to the increasing relative importance of bedrock topography on the increasingly shallow ice body. This causes a number of commonly used simplifying assumptions to break down. The ice margin optimally requires a full solution of the Stokes equations, which has only recently been possible in the so-called 3D higher-order ice-dynamic models. Alternatively, one can apply full-solution 2D-flowline models and then combine the results to spatial picture of ice-marginal change. These tools are now available and experiments to do nested modelling of a higher-order model in a more general ice-sheet scale model are underway, although this field is not yet well developed. A new ice-dynamic modelling effort at Paakitsoq will first of all be able to make use of the extensive bedrock topographical information available, as the last study was done before any bedrock elevation data wsa available (Reeh, 1983). Additionally, the availability of a detailed climate simulation extending into the next century, distributed surface velocity data, 25 years of mass-balance data, englacial temperature data, and surface topographical data from 1948, 1959, 1985 and 2005 would make a new modelling effort extremely useful for hydropower planning purposes.

In addition to the mentioned sources of new knowledge and data, the latest development in the understanding of englacial and subglacial drainage will also contribute to a renewed effort in the Paakitsoq basin. This field of research is highly active with a lot of attention being devoted to the Greenland ice sheet due to its recently increased contribution to global sea-level rise. Parallel to the concern-driven research in the effects of climate change on the Greenland ice sheet, there is an increase in specifically Nordic glaciological research pertaining to the impact of climate change on renewable energy sources, with hydropower as the most important of those. Thus, scenarios for the future Greenland hydropower potential have already been part of such research efforts and the Paakitsoq basin is planned to be the next catchment entering into this collaboration. The important side-effect is that the expertise of researchers and specialists from other Nordic countries will be applied to Greenland conditions.

In conclusion, it is strongly recommended that the future hydropower potential of the Paakitsoq basin is re-evaluated using the extensive amounts of new data and knowledge available. Such an investigation should provide a highly useful tool in the further planning of hydropower development, not only considering the immediate application to the town of Ilulissat, but also through supporting the development of expertise related to utilization of the large hydropower potentials of the basins adjoining the Greenland ice sheet in West Greenland. These basins are currently under evaluation for supporting energy-intensive aluminium industry and represent a huge unexploited potential for economical development in Greenland. The suggested re-evaluation of the Paakitsoq basin will thus provide the necessary tools and methods to investigate the uncertainties in the future potential of these larger basins.

References

Ahlstrøm, A. P. 2003: Ice Sheet Ablation Assessed by Observation, Remote Sensing and Modelling. PhD Thesis, University of Copenhagen. Geological Survey of Denmark and Greenland Report 2003/49.

Ambach, W. 1963: Untersuchungen zum Energieumsatz in der Ablationszone des grönlandischen Inlandeises (Camp IV-EGIG, 69°40’05” N, 49°37’58” W). *Meddelelser om Grønland*, 174(4). *(in German)*

Ambach, W. and M. Kuhn. 1985: Accumulation gradients in Greenland and mass balance response to climatic changes. *Z. Gletscherk. Glazialgeol.*, 21, 311–317.

Blatter, H. 1985: On the thermal regime of Arctic valley glaciers; a study of the White Glacier, Axel Heiberg Island, and the Laika Glacier, Coburg Island, N.W.T., Canada. Zürich: Eidgenössische Technische Hochschule, Geographisches Institut (*Züricher Geographische Schriften* **22**).

Braithwaite, R. J., Clement, P., Clausen, H. 1982: Inferences from a 19 m firn core, Nordbogletscher, South Greenland. *Rapp. Grønlands geol. Unders.* **120**, 96–98.

Braithwaite, R. J. 1983: Glaciological investigations at Qamanârssûp sermia. Interim report 1982 and Appendix tables. *Grønlands geol. Unders. Gletscher-hydrol. Meddr*. 83/4, 34 pp.

Braithwaite, R. J. & Thomsen, H. H. 1983: Afstrømningen fra Paakitsup Akuliarusersua, Jakobshavn. *Grønlands geol. Unders. Arbejdsnotat*, 7 pp. (unpublished, *in Danish)*

Braithwaite, R. J. & Thomsen, H. H. 1984: Runoff conditions at Paakitsup Akuliarusersua, Jakobshavn, estimated by modelling. *Grønlands geol. Unders. Gletscher-hydrol. Meddr.* **84/3**, 22 pp.

Braithwaite, R. J. & Thomsen, H. H. 1989: Simulation of runoff from the Greenland ice sheet for planning hydro-electric power, Jakobshavn/Ilulissat, West Greenland. *Annals of Glaciology* **13**, 12–15.

Braithwaite, R. J., Olesen, O. B. & Thomsen, H. H. 1992a: Calculated variations of annual ice ablation at the margin of the Greenland ice sheet, West Greenland, 1961–1990. *Journal of Glaciology* **38**(129), 226–272.

Braithwaite, R. J., Pfeffer, W. T., Blatter, H., Humphrey, N. F. 1992b: Meltwater refreezing in the accumulation area of the Greenland ice sheet: Pâkitsoq, summer 1991. *Rapp. Grønlands geol. Unders.* **155**, 13–17.

Braithwaite, R. J. and Laternser, M. 1993: Measurements of firn density in the lower accumulation area of the Greenland ice sheet: EPOCH 1992. *Rapp. Grønlands geol. Unders.* **159**, 62–65.

Braithwaite, R. J., Laternser, M. and Pfeffer, W. T. 1994: Variations of near-surface firn density in the lower accumulation area of the Greenland ice sheet, Pâkitsoq, West Greenland. *Journal of Glaciology* 40(136), 477–485.

Clement, P. 1983: Glacial-hydrologiske forhold i Nordbosø bassinet, Johan Dahl Land. *Grønlands geol. Unders. Gletscher-hydrol. Meddr.* **83/9**, 54 pp.

GTO. 1979: Forundersøgelsesrapport vandkraft. Bynære bassiner 1979, 133 pp. *(in Danish)*

GTO. 1981: Forundersøgelse vandkraft 1980. Egedesminde – Christianshåb, Jakobshavn, Godhavn, 9 pp. *(in Danish)*

GTO. 1982: Forundersøgelse vandkraft 1981. Ilulissat/Jakobshavn, 43 pp. *(in Danish)*

GTO. 1983: Forundersøgelse vandkraft 1982. Anlægsteknik – Hydrologi. Paakitsup Akuliarusersua. Ilulissat/Jakobshavn, 53 pp. *(in Danish)*

GTO. 1984: Forundersøgelse vandkraft 1983. Anlægsteknik – Hydrologi. Paakitsup Akuliarusersua. Ilulissat/Jakobshavn, 74 pp. *(in Danish)*

GTO. 1985: Vandkraftundersøgelser hydrologi. Paakitsup Akuliarusersua, 60 pp. *(in Danish)*

GTO. 1986a: Vandkraftværk Paakitsup Akuliarusersua. Ilulissat/Jakobshavn. Projektforslag, 123 pp. *(in Danish)*

GTO. 1986b: Generelle hydrologiske bassin-informationer. Bynære bassiner 1986. Supplement 1, 161 pp. *(in Danish)*

Hock, R., 1999: A distributed temperature index ice and snow melt model including potential direct solar radiation. *Journal of Glaciology* 45(149), 101-111.

Hock , R. and B. Holmgren, 2005: A distributed energy balance model for complex topography and its application to Storglaciären, Sweden. *Journal of Glaciology*, 51(172), 25-36.

Holtzscherer, J.-J. & Bauer, A. 1954: Contributions á la Connaissance de l'Inlandsis du Groenland, *Publs. Assoc. Int. Hydrologiques Scient.* 39, 244–296. *(in French)*

Iken, A., Röthlisberger, H., Hutter, K. 1977: Deep drilling with a hot water jet. *Z. Gletscherk. Glazialgeol.* **12**(2), 143–156.

Iken, A. 1990. Tiefbohrungen im Jakobshavn Gletscher, Grönland. *Bull. Eidgen. Techn. Hochsch. Zürich* **228**, 8–10.

Joset, A. 1950: Sondages seismiques, *In: Exp. Polaires Francaises, Rapports Préliminaires* 15, 57–64. *(in French)*

Konzelmann, T., R. S. W. van de Wal, W. Greuell, R. Bintanja, E. A. C. Henneken, and A. Abe-Ouchi. 1994: Parameterisation of global and longwave incoming radiation for the Greenland Ice Sheet. *Global and Planetary Change*, 9, 143–164.

Munck, F. 1950: Gravimetrie, *In: Exp. Polaires Francaises, Rapports Préliminaires* 15, 57–64. *(in French)*

NIRAS Greenland. 2005: Energiplan 2020 – Hovedrapport, Greenland Homerule, 68 pp. *(in Danish)*

Nye, J. F. 1960: The response of glaciers and ice-sheets to seasonal and climatic changes. *Proc. Royal Soc.,* Ser. A, **256**, 1287, 559–584.

Nye, J. F. 1965a: The frequency response of glaciers. *J. Glaciol.* **5**(41), 567–588.

Nye, J. F. 1965b: A numerical method of inferring the budget history of a glacier from its advance and retreat. *J. Glaciol.* **5**(41), 589–607.

Oerter, H., Reeh, N., Letréguilly, A. & Thomsen, H. H. 1990: Oxygen-18 studies on the surface ice from the margin of the Greenland ice sheet at Pâkitsoq, West Greenland. *Annals Geophysicae, 1990 Special issue. European Geophysical Society. XV General Assembley, Copenhagen, April 1990*. Abstract, 153–154.

Ohmura, A. Steffen, K., Blatter, H., Greuell, W., Rotach, M., Konzelmann, T., Laternser, M., Ouchi, A. & Steiger, D. 1991: Energy and mass balance during the melt season at the equilibrium line altitude, Paakitsoq, Greenland ice sheet. Progress report No. 1, Department of Geography, ETH, Zürich, 108 pp.

Olesen, O. B. and Clausen, A. 1988: Test drilling with a hot water jet at the Inland Ice margin, Pâkitsup, central West Greenland. *Rapp. Grønlands geol. Unders.* **140**, 121–124.

Olesen, O. B. 1989: A Danish contribution to the family of hot-water glacier drills. *In* Rado, C. & Beaudoing, D. (eds.) Ice core drilling. *Proc. Third Internat. Workshop on Ice Drilling Technology, Grenoble, France, Oct. 1988,* 140–148.

Reeh, N. & Thomsen, H. H. 1986a: Stable isotope studies on the Greenland ice-sheet margin. *Rapp. Grønlands geol. Unders.* **130**, 108–114.

Reeh, N. & Thomsen H. H. 1986b: Model for simulering af ilt-isotop variationen i smeltevandsafstrømningen fra Indlandsisens rand ved Paakitsup Akuliarusersua, Jakobshavn. *Grønlands geol. Unders. Gletscher-hydrol. Meddr* **86/1**, 34 pp. *(in Danish)*

Reeh, N. Thomsen, H. H. & Clausen, H. B. 1987: The Greenland ice-sheet margin – a mine of ice for paleo-environmental studies. *Palaeogeography, Palaeoclimatology, Palaeoecology* **58**(1987), 229–234.

Reeh, N & Thomsen, H. H. 1988: Stable isotope studies of ice margins in Greenland. *18 Nordiske Geologiske Vintermøde, Jan. 1988, Copenhagen.* Abstract, 347–348.

Reeh, N. & Thomsen, H. H. 1989: Using stable isotopes for delineation of drainage basins on the Greenland ice-sheet margin for use in planning hydro-electric power supply, Ilulissat/Jakobshavn, West Greenland. *Annals of Glaciology* **13**. Abstracts, 296.

Reeh, N. & Thomsen, H. H. 1993: Using stable isotopes as natural tracers to delineate hydrological drainage basins on the Greenland ice-sheet margin. *Chemical Geology, Isotope Geoscience Section,* **109**, 281–291.

Reeh, N., Thomsen, H. H. & Bøggild, C. E. 1993: Palaeo-environmental studies on the Greenland ice sheet margin. *Rapp. Grønlands geol. Unders.* **159**, 66–68.

Reeh, N. & Thomsen, H. H. 1994a: Introduction to the Paakitsoq 1994 field programme. *In* Thomsen, H. H. & Reeh, N. (ed.) Field report on palaeo-environmental studies at the Greenland ice sheet margin, Paakitsoq, West Greenland, 1994. *Open File Ser. Grønlands geol. Unders.* **94/15**, 2–9.

Reeh, N. & Thomsen, H. H. 1994b: Ice sampling for ^{10}Be and δ^{18}O analysis on the ice margin at Paakitsoq, West Greenland. *In* Thomsen, H. H. & Reeh, N. (ed.) Field report on palaeo-environmental studies at the Greenland ice sheet margin, Paakitsoq, West Greenland, 1994. *Open File Ser. Grønlands geol. Unders.* **94/15**, 30–34.

Reeh, N., Thomsen, H. H., Oerter, H. & Bøggild, C. E. 1994c: Ice deformation and velocities along the Paakitsoq profile sampling line, West Greenland. *In* Thomsen, H. H. & Reeh, N. (ed.) Field report on palaeo-environmental studies at the Greenland ice sheet margin, Paakitsoq, West Greenland, 1994. *Open File Ser. Grønlands geol. Unders.* **94/15**, 20–29.

Reeh, N., Oerter, H. & Thomsen, H. H. 2002: Comparison between Greenland ice-margin and ice-core oxygen-18 records. *Annals of Glaciology* **35***,* 136–144*.*

Thomsen, H. H. 1983a: A glaciological field and mapping programme in connection with hydropower, West Greenland. *Rapp. Grønland geol. Unders.* **115**, 102–107.

Thomsen, H. H. 1983b: Glaciologiske undersøgelser ved Pâkitsup ilordlia 1982, Ilulissat/Jakobshavn. *Grønlands geol. Unders. Gletscher-hydrol. Meddr* **83/3**, 24 pp. *(in Danish)*

Thomsen, H. H. 1983c: Satellitdata – et redskab til studier af Indlandsisens randzone i forbindelse med vandkraftundersøgelser. *Grønland geol. Unders. Gletscher-hydrol. Meddr* **83/8**, 24 pp. *(in Danish)*

Thomsen, H. H. 1983d: Glaciological applications of Landsat images in connection with hydropower investigations in West Greenland. *Proceeding EARSeL/ESA Symposium on Remote Sensing Applications for Environmental Studies. European Space Agency* **188**, 133–136.

Thomsen, H. H. 1984a: Mass balance measurements at the margin of the Inland Ice near Jakobshavn, West Greenland. *Polarforschung* **54**, 1, 37–41.

Thomsen, H. H. 1984b: Glaciological reconnaissance, mass balance measurements and mapping programmes in connection with Greenland hydropower. *Rapp. Grønlands geol. Unders.* **120**, 95–99.

Thomsen, H. H. 1984c: Glaciologiske undersøgelser i Disko Bugt området 1983. *Grønlands geol Unders. Gletscher-hydrol. Meddr* **84/1**, 31 pp. *(in Danish)*

Thomsen, H. H. 1984d: Sammenfattende aktivitetsrapport for det glaciologiske delprogram for vandkraftprojektet bynære bassiner 1982–1984. *Grønlands geol. Unders. Gletscher-hydrol. Meddr* **84/4**, 12 pp. *(in Danish)*

Thomsen, H. H. 1985a: Glaciological field work and remote sensing in connection with hydropower investigations, West Greenland. *Rapp. Grønlands geol. Unders.* **125**, 95–99.

Thomsen, H. H. 1985b: Arbejdsnotet vedrørende afgrænsningen af dræningsarealer ved Paakitsup Akuliarusersua, Jakobshavn. *Grønlands geol. Unders. Arbejdsnotat*, 4 pp. (unpublished, *in Danish*)

Thomsen, H. H. 1985c: Present and future applications of remote sensing with examples from Greenland. *Western European Geological Surveys, Expert Meeting on Remote Sensing, Nov. 1985, Helsinki*. Abstract, 1–2. (unpublished)

Thomsen, H. H. & Braithwaite, R. J. 1985: Estimation of runoff for proposed hydropower stations near Jakobshavn and Christianshåb, West Greenland. *Rapp. Grønlands geol. Unders.* **125**, 100–103.

Thomsen, H. H. & Madsen, P. S. 1985: Radio ekko målinger af Indlandsisens randzone i Disko Bugt området 1984. *Grønlands geol. Unders. Gletscher-hydrol. Meddr* **85/1**, 20 pp. *(in Danish)*

Thomsen H. H. 1986a: Photogrammetric and satellite mapping of the margin of the Inland Ice, West Greenland. *Annals of Glaciology* **8**, 164–167.

Thomsen, H. H. 1986b: Is og Vandkraft: glaciologi i vandkraftprojektet bynære bassiner, 1982–1986. Glaciology and Hydropower: Glaciology for local hydropower projects, 1982–1986. *Grønlands geol. Unders. Gletscher-hydrol. Meddr* **86/2**, 73 pp. *(in Danish)*

Thomsen, H. H. & Reeh, N. 1986: Glaciological investigations at the margin of the Inland Ice north-east of Jakobshavn, West Greenland. *Rapp. Grønlands geol. Unders.* **130**, 102–108.

Thomsen, H. H., Thorning, L. & Braithwaite, R. J. 1986: Vurdering af de gletscher-hydrologiske forhold på Indlandsisen ved Paakitsup Akuliarusersua, Ilulissat/Jakobshavn. *Grønlands geol. Unders. Arbejdsnotat*, 42 pp. (unpublished, *in Danish*)

Thomsen, H. H. 1987a: Continued glaciological work north-east of Jakobshavn, West Greenland. *Rapp. Grønlands geol. Unders.* **135**, 84–87.

Thomsen, H. H. 1987b: Glaciologiske undersøgelser nordøst for Ilulissat. I: *Energiforsyning i arktiske områder. Selskabet for Arktisk Teknolog*i, 1–12. *(in Danish)*

Thomsen, H. H. & Braithwaite, R. J. 1987: Use of remote-sensing data in modelling run-off from the Greenland ice sheet. *Annals of Glaciology* **9**, 215–217.

Thomsen, H. H. 1988a: Mass balance, ice velocity and ice temperature at the Inland Ice margin north-east of Jakobshavn, central West Greenland. *Rapp. Grønlands geol. Unders.* **140**, 111–114.

Thomsen, H. H. 1988b: Mapping and modelling of glacier drainage in the Pâkitsoq basin, central West Greenland. *Rapp. Grønlands geol. Unders.* **140**, 114–117.

Thomsen, H. H. 1988c: Glaciological research for hydropower planning in Ilulissat/Jakobshavn, West Greenland. *Proceedings of the Seventh Northern Research Basins Symposium/Workshop, Ilulissat, Greenland, May 1988* on Applied Hydrology in the Development of Northern Basins, 47–56.

Thomsen, H. H., Thorning, L. & Braithwaite, R. J. 1988: Glacier-hydrological conditions on the Inland Ice north-east of Jakobshavn/Ilulissat, West Greenland. *Rapp. Grønlands geol. Unders.* **138**, map with text.

Thomsen, H. H., Olesen, O. B., Braithwaite, R. J. & Weidick, A. 1989a: Greenland ice-margin programme, a pilot study at Pâkitsoq, north-east of Jakobshavn, central West Greenland. *Rapp. Grønlands geol. Unders.* **145**, 50–53.

Thomsen, H. H., Thorning, L. & Olesen, O. B. 1989b: Applied glacier research for planning hydro-electric power, Jakobshavn/Ilulissat, West Greenland. *Annals of Glaciology* **13**, 257–261.

Thomsen, H. H. & Olesen, O. B. 1990: Continued glaciological investigations with respect to hydropower and ice-climate relationships, at Pâkitsoq, Jakobshavn, West Greenland. *Rapp. Grønlands geol. Unders.* **148**, 83–86.

Thomsen, H. H. & Olesen, O. B. 1991: Studies of hydraulics and hydrology on the Inland Ice. *Rapp. Grønlands geol. Unders.* **152**, 36–38.

Thomsen, H. H., Olesen, O. B., Braithwaite, R. J. & Bøggild, C. E. 1991: Ice drilling and mass balance at Pâkitsoq, Jakobshavn, central West Greenland. *Rapp. Grønlands geol. Unders.* **152**, 80–84.

Thomsen, H. H. & Reeh, N. 1994: Field report on palaeo-environmental studies at the Greenland ice sheet margin, Paakitsoq, West Greenland, 1994. *Open File Ser. Grønlands geol. Unders.* **94/15**, 44 pp.

Thomsen, H. H., Biscaye, P. E., Bourgeois, J. C., Bøggild, C. E., Oerter, H., Reeh, N. & Thorsteinsson, T. 1995: An international palaeo-environmental project at Paakitsoq, West Greenland. *Rapp. Grønlands geol. Unders.* **165**, 88–92.

Thomsen, H. H. & Reeh, N. 1996: Lettilgængelig overfladeis kan fortælle om klimaet. *Vand & Jord* **3**(4), 160–162. *(in Danish)*

Thomsen, T. & Thomsen, H. H. 1986: Hydrological data collection, interpretation and analysis in Greenland. *Proceedings of the Cocoa Beach Workshop, Florida, August 1984* on Hydrological Applications of Space Technology. *IAHS Publ.* **160**, 273–282.

Thorning, L. Thomsen, H. H. & Hansen, E. 1986: Geophysical investigations at the Inland Ice margin of the Pâkitsoq basin, central West Greenland. *Rapp. Grønlands geol. Unders.* **130**, 114–121.

Thorning, L. & Hansen, E. 1987: Electromagnetic reflection survey 1986 at the Inland Ice margin of the Pâkitsoq basin, central West Greenland. *Rapp. Grønlands geol. Unders.* **135**, 87–95.

Thorning, L. & Hansen, E. 1988a: Electromagnetic reflection survey 1987 in key areas of the Pâkitsoq basin at the margin of the Inland Ice, central West Greenland. *Rapp. Grønlands geol. Unders.* **140**, 117–118.

Thorning, L. & Hansen, E. 1988b: Construction and testing of a lightweight radar for ice-thickness determinations on the glaciers in the Pâkitsoq area, central West Greenland. *Rapp. Grønlands geol. Unders.* **140**, 118–121.

Thorning, L. & Hansen, E. 1989: Measurements of ice thickness on glaciers at Isortuarssûp tasia, southern West Greenland and Pâkitsoq, central West Greenland. *Rapp. Grønlands geol. Unders.* **145**, 59–63.

Shreve, R.L. 1972: Movement of water in glaciers. *J. Glaciol.*, **11**(62), p. 205-214.

Weidick, A. 1968: Observations on some Holocene glacier fluctuations in West Greenland. *Bull. Grønlands geol. Unders.* **73** (also *Meddr Grønland* **165** (6) 202 pp.).

Weidick, A., Knudsen, N. T., Bøggild, C. E. 1992: Glacier inventory of West Greenland, *Rapp. Grønlands geol. Unders.* **158**.

Weidick, A., Oerter, H., Reeh, N., Thomsen, H. H. & Thorning, L. 1990: The recession of the Inland Ice margin during the Holocene climatic optimum in the Jakobshavn Isfjord area of West Greenland. *Palaeogeography, Palaeoclimatology, Palaeoecology* **82**(3/4), 389–399.

Weidick, A. & Thomsen, H. H. 1983a: Lokalgletschere og Indlandsisens rand i forbindelse med udnyttelse af vandkraft i bynære bassiner. *Grønlands geol. Unders. Gletscher-hydrol. Meddr* **83/2** 129 pp. *(in Danish)*

Weidick, A. & Thomsen H. H. 1983b: Glaciology related to future hydropower for Greenland towns, *Rapp. Grønlands geol. Unders.* **115**, 100–101.

Weidick, A. & Thomsen, H. H. 1986: A decade of glacier investigations for utilisation of Greenland hydropower. *Rapp. Grønlands geol. Unders.* **128**, 157–169.