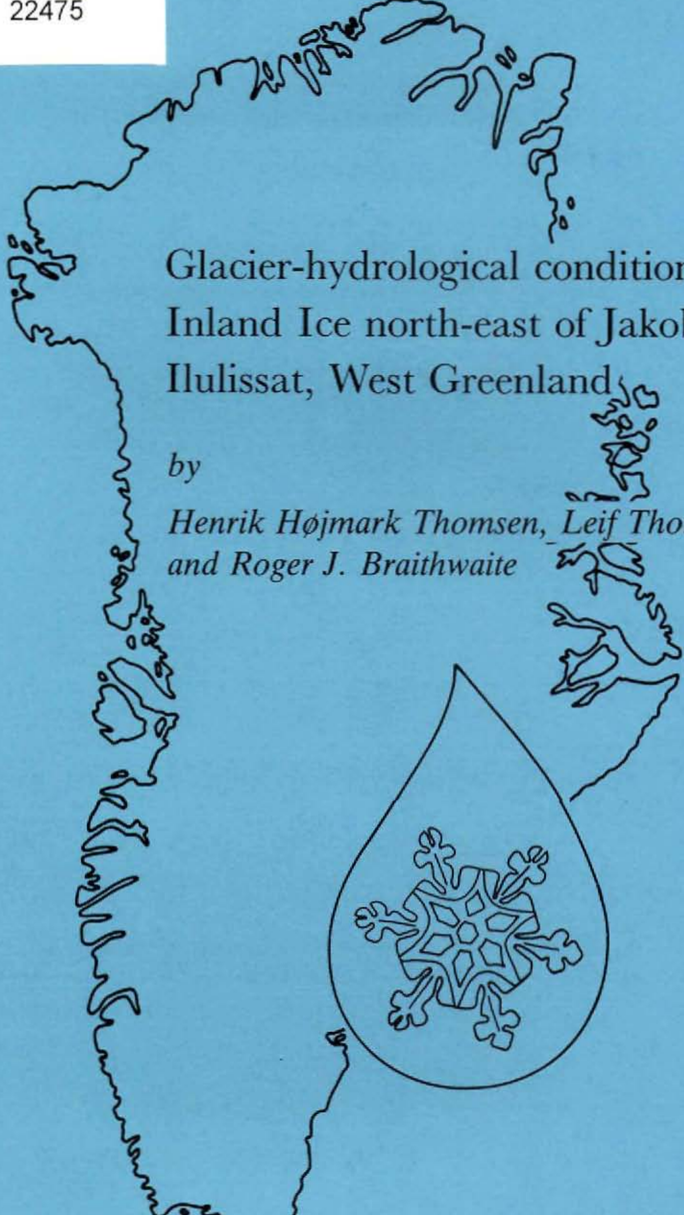

Glacier-hydrological conditions on the Inland Ice north-east of Jakobshavn/ Ilulissat, West Greenland

by

Henrik Højmark Thomsen, Leif Thorning and Roger J. Braithwaite

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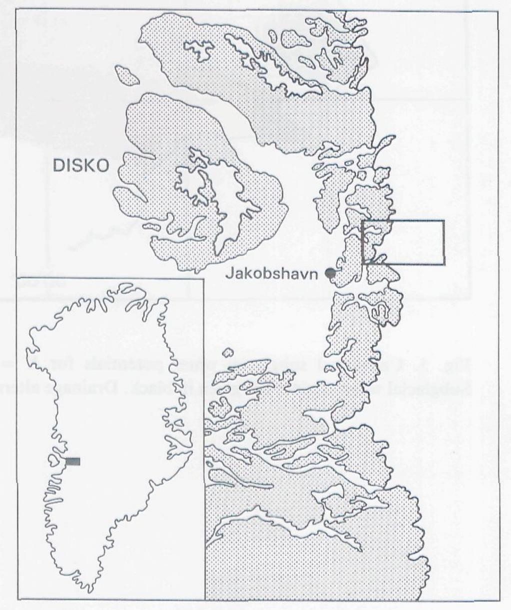
GRØNLANDS GEOLOGISKE UNDERSØGELSE

THE GEOLOGICAL SURVEY OF GREENLAND

Contours on exposed ice, 20 m interval — Contours on snow and firn, 20 m interval — Ice margin (certain, uncertain delineation) — Crevasses — Lineation — Debris covered ice — Stream — Supraglacial stream ending in moulin — Extent of empty lake — Trace of water drainage — Lake and fjord with icebergs — Contours on land, 50 m interval — Trim line zone — Moraine — Glaciofluvial deposits



MARGIN OF THE INLAND ICE NORTH-EAST OF JAKOBSHAVN

(PAAKITSUP AKULIARUSERSUA 1985)

GLACIER HYDROLOGICAL MAP

1:75 000

UTM zone 22 (NAD83 – NAD83) Lake heights in metres

0 5 10 15 km

THE GEOLOGICAL SURVEY OF GREENLAND

Compilation scale 1:75 000 on vertical photographs, 10 July 1985 by Markhurd Corporation on contract from University of Maine (Markhurd route JOV 6-16).

Aerial triangulation and plotting on Kern PG-2 stereo plotting instrument at Photogrammetric Laboratory of The Geological Survey of Greenland.

Absolute orientation based on control points supplied by Greenland Technical Organization and H. Brecher, Institute of Polar Studies, The Ohio State University, U.S.A.

Physiographic content and map compilation by H. Højmark Thomsen, 1986.

Contour plotting by O. Winding, 1986.

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GRØNLANDS GEOLOGISKE UNDERSØGELSE
THE GEOLOGICAL SURVEY OF GREENLAND

INTRODUCTION

The Greenland ice sheet, the Inland Ice, covers about 1.7 million km² which is about 79% of Greenlands total area. The Inland Ice plays an important hydrological role, global as well as local. About 7% of the world's fresh water is stored here and the largest rivers in Greenland originate from the ice sheet.

Since the mid 1970s there has been a great interest in developing hydropower in Greenland. Many of the proposed basins are connected to the Inland Ice. For example, a hydropower installation is planned at Paakitsup Akuliarusersua about 45 km north-east of Jakobshavn town, in the Disko Bugt area, West Greenland which will supply energy to Jakobshavn with 3500 inhabitants.

The basin has an ice-free area of 33.6 km² and includes a much larger adjoining sector of the Inland Ice, so that about 90% of the runoff from the basin is meltwater from the ice sheet. The water drains from the ice sheet through three lakes, 326 and 233 which are connected to the central lake 187. The lakes 233 and 187 are proposed as two separate reservoirs with tunnels leading to a power station in the fjord just west of lake 233.

Delineation of drainage basins within a continuous ice cover like the Inland Ice is an important topic within the hydropower investigations. Meltwater may drain supraglacially, often through large river systems, directly to the glacier margin. However, much of the water escapes down into the ice through crevasses and moulins from where it drains in a network of cracks and conduits in and beneath the ice. Meltwater draining within and beneath glaciers is difficult to study and has mainly been dealt with through theoretical studies (Shreve, 1972; Björnsson, 1982; Röthlisberger & Lang, 1987). It is generally agreed that the subglacial topography can seriously influence the direction of meltwater drainage by forcing the water to follow subglacial valley systems. Delineation of drainage basins, therefore, requires information about supraglacial and subglacial conditions.

ENGLACIAL AND SUBGLACIAL DRAINAGE

Studies of englacial and subglacial water drainage are concerned usually with drainage in temperate ice conditions. Meltwater drainage in and beneath glaciers can be compared with movement of groundwater through permeable cavernous limestone. Some water drains through tiny cracks and openings along the single ice crystals (Shreve, 1972) but most water drains in larger channels and conduits starting at moulins and crevasses on the surface. Moulins can reach several hundred metres down into the ice, directly to the bottom (Iken, 1972; Meier, 1973). At the bottom of the ice, water can drain in a thin water film or in subglacial channels, depending on the regional and local subglacial conditions (Röthlisberger & Lang, 1987).

No direct observations exist about englacial and subglacial drainage from the area at Paakitsup Akuliarusersua. Ice temperatures in the area are unknown but ice temperatures have been measured just 25 km north of Paakitsup Akuliarusersua at an elevation of about 750 m near the ice margin (Stauffer & Oeschger, 1979) which show mean ice temperatures decreasing from 0°C at the surface to about −6°C at a depth of 5 m, and increasing to 0°C again at a depth of about 40 m. Glacier dynamic modelling has been applied to the Inland Ice sector at Paakitsup Akuliarusersua by Reeh (1983) who suggests from calculation the existence of a zone 18 to 292 km from the ice margin where the ice is at the pressure melting point at the bottom, and another zone 0 to 18 km from the ice margin with full developed bottom sliding with high water pressure and possibilities for development of cavities. Observations from White Glacier on Axel Heiberg Island suggest that water escaping down into moulins is able to reach the bottom of the ice, even under conditions of cold ice (Iken, 1972).

From the above it is assumed that surface water at Paakitsup Akuliarusersua which escapes down into the ice through moulins quickly flows to the bottom of the ice, and there drains according to the subglacial condition.

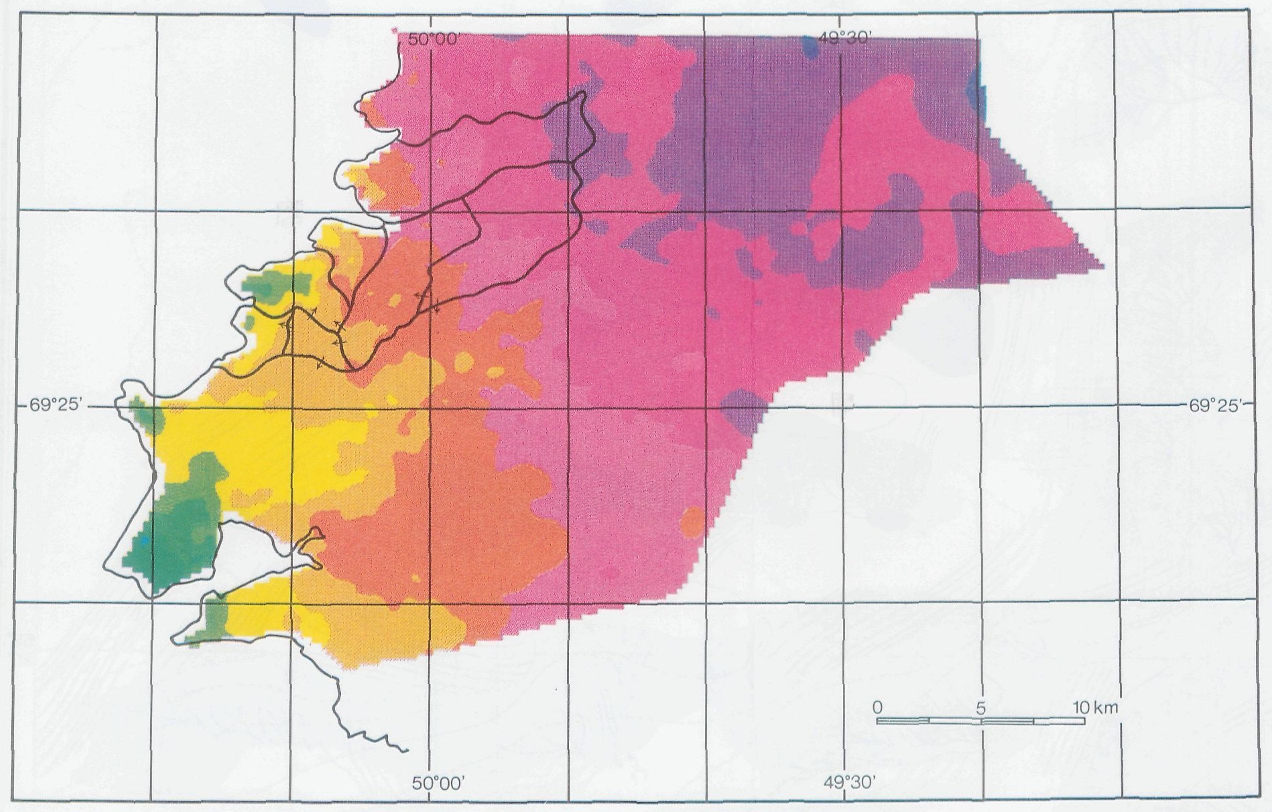

Fig. 4. Calculated subglacial water potential for $K = 0.7$. Units in $10^3 N/m^2$. Subglacial water divides are given in black. Drainage alternatives given with arrows. Colour scale as fig. 5.

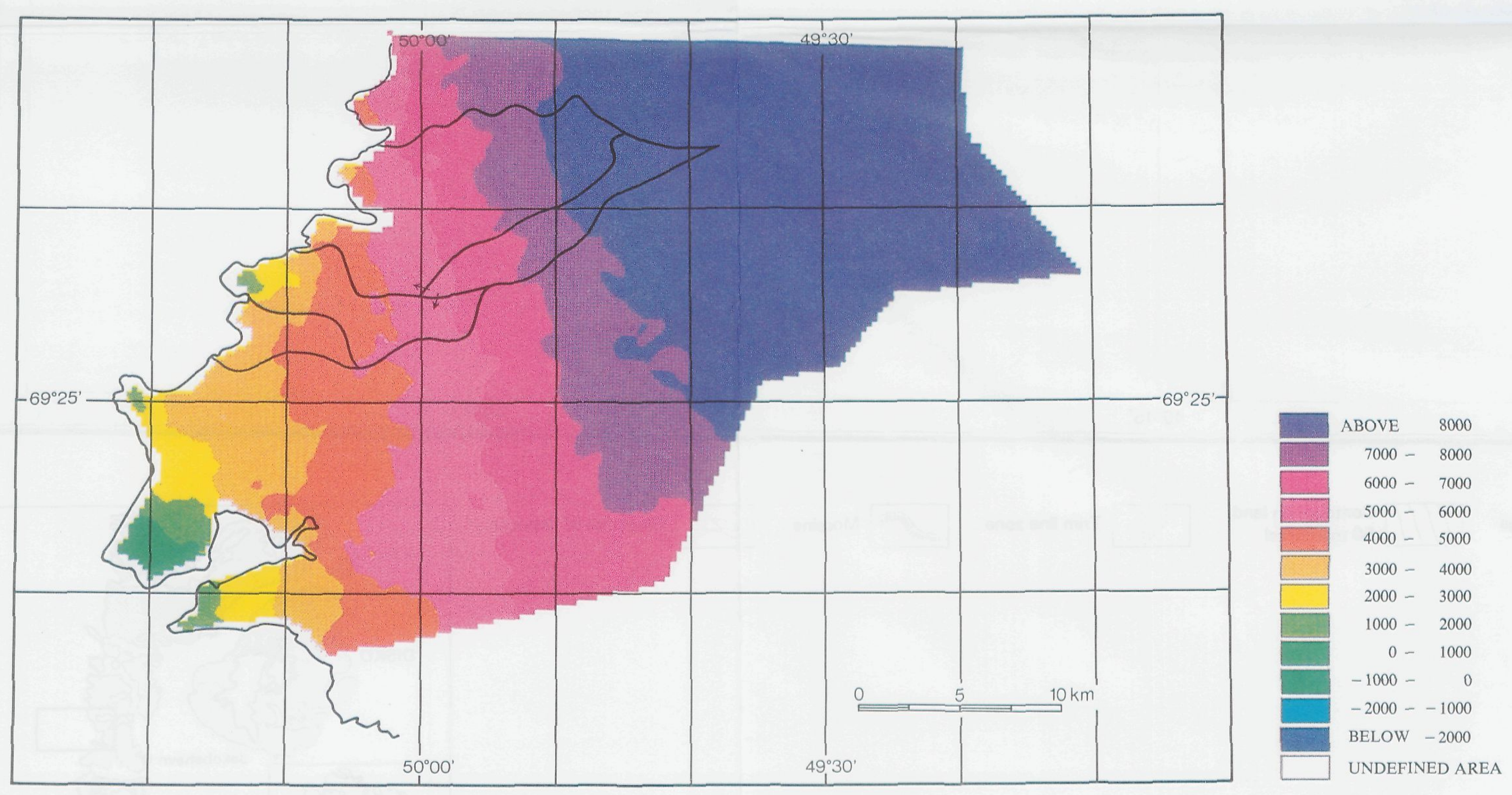

Fig. 5. Calculated subglacial water potentials for $K = 1.0$. Units in $10^3 N/m^2$. Subglacial water divides are given in black. Drainage alternatives given with arrows.

SURFACE TOPOGRAPHY AND DRAINAGE

A photogrammetric map in scale 1:75 000 was prepared covering the ice-free part of the basin and the adjoining sector of the Inland Ice (map sheet). The map, based on vertical aerial photographs in scale 1:150 000 from 10 July 1985, was plotted on a Kern PG-2 stereo-plotting instrument connected to a HP computer system. The map gives physiographic information and surface topography with contour intervals of 50 m in the ice-free area and 20 m on the ice. All possible details have been plotted for the glacier area and trim line zone. This includes features on the ice especially related to surface hydrology such as rivers, lakes and moulins as well as crevasses and lineaments influencing the drainage pattern. The map plotting is supported by observations in the field on foot or from helicopter.

Meltwater drains through innumerable small rivers, merging into large river systems which follow valleys in the strongly undulating ice surface. The drainage pattern is determined by both local and general surface topography and to a lesser degree by structural features such as healed crevasses and lineaments indicating the ice flow pattern. The large rivers occupy larger valleys on the ice where they drain in 1 – 3 m deep river beds cut down into the ice. The rivers drain over varying distances on the surface, but often escape down into the ice through moulins or crevasses before reaching the ice margin.

The moulins vary in size depending on age and the amount of water draining to them, but diameters of 1 – 2 m are normal for the area.

Many lakes exist on the ice surface, in many cases without any detectable surface outlet. They vary in size from only a few hundred metres in diameter up to about 1500 m and soundings in the lakes show mean water depths of 2 – 5 m. The lakes tap periodically. From visits to empty lakes it can be seen that moulins often exist in the bottom of the lakes and from visits to the same locations in spring it is reasonable to assume that these moulins are filled with snow during the winter, thus stopping the outlet. Lakes are then filled during summer until the moulins are reopened by a combination of pressure and melting. Similar observations were made in Spitsbergen (Liestøl *et al.*, 1980).

It is reasonable to assume that the described drainage pattern on the ice is semi-permanent which means that draining takes place in the same valley systems from year to year, because the valleys are features of local topography caused by ice movement over the hilly subglacial terrain. Comparisons with older maps, plotted from vertical aerial photographs from 1959 (Thomsen, 1986) and oblique aerial photographs from 1948 show a stable main drainage pattern with rivers draining in the same valley systems and moulins lying in approximately the same positions. This is despite a mean glacier thinning of 14 m up to an elevation of 500 m a.s.l. in the period 1959 to 1985.

On the basis of the glacier hydrological map, the ice surface was divided into a number of drainage cells, each draining to a moulin or moulin complex (fig. 1). In the highest elevation, the individual drainage cells have arbitrarily been cut at an elevation of 1100 m a.s.l. The mapped area consists of a total of 249 drainage cells. The meltwater drainage from each cell will depend on englacial and subglacial conditions below the point of escapement into the ice. Area sums of drainage cells have been used for calculation of surface runoff to individual subglacial sub-basins.

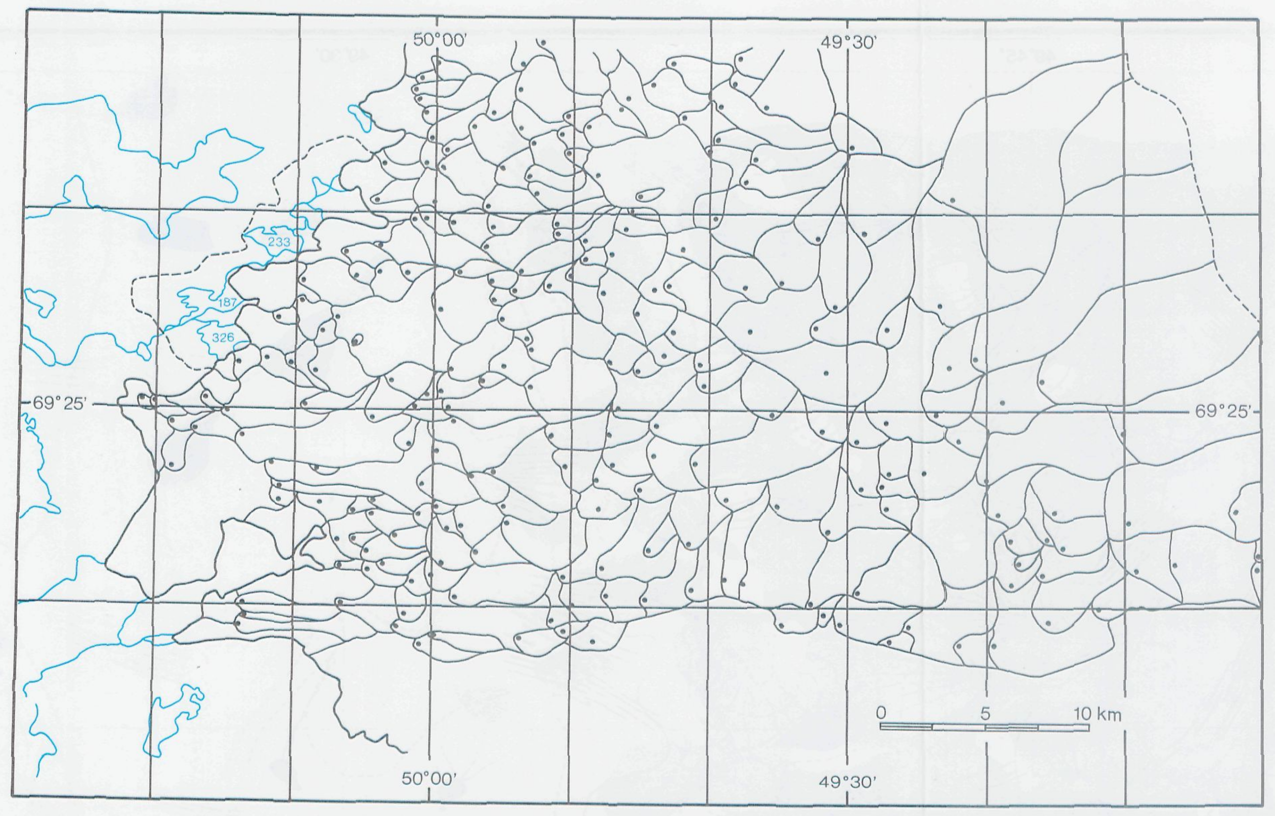

Fig. 1. Drainage cells, each draining to a moulin or moulin complex.

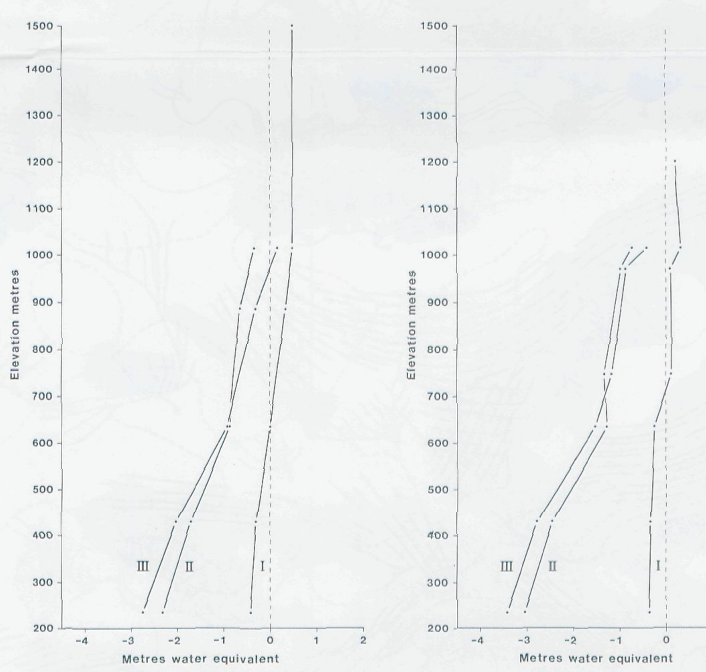

Fig. 6. Mass balance in relation to elevation.
a. I) Transient balance (15th August 1982 – 12th May 1983)
II) Transient balance (12th May 1983 – 11th August 1983)
III) Annual balance (15th August 1982 – 11th August 1983)
b. I) Transient balance (11th August 1983 – 15th May 1984)
II) Transient balance (15th May 1984 – 24th August 1984)
III) Annual balance (11th August 1983 – 24th Augsut 1984)

DELINEATION OF SUBGLACIAL DRAINAGE AREAS

A model by Björnsson (1982) describing subglacial drainage was used for calculating the drainage at the glacier bed. The model describes water drainage down the gradient of a potential which depends upon both surface and subglacial topography. The potential

$$POT = \varrho_w \cdot g \cdot Z_b + P_w \qquad (1)$$

is the sum of the gravitation potential and the basal water pressure P_w. ϱ_w is the density of water, g is the acceleration of gravity and Z_b the elevation of the glacier bed relative to a horizontal datum. The basal water pressure fluctuates according to the supply of water at the base and the resistance to flow in subglacial channels. As a first approximation of subglacial drainage on a regional scale, P_w can be expressed as

$$P_w = K \cdot P_i \qquad (2)$$

where P_i is the ice overburden pressure and K ($\leqslant 1$) a constant called the K factor. The ice overburden pressure is expressed as

$$P_i = \varrho_i \cdot g \cdot (Z_s - Z_b) \qquad (3)$$

where ϱ_i is the density of ice and Z_s the elevation of the ice surface relative to the same datum level as for the bed elevation Z_b. The potential can now be expressed as

$$POT = (\varrho_w - K \cdot \varrho_i) \cdot g \cdot Z_b + K \cdot \varrho_i \cdot g \cdot Z_s \qquad (4)$$

Water flowing in an isotropic basal layer will drain perpendicular to the potential lines. The model is a first-order approximation of subglacial drainage and does not describe details in water flow.

The model calculations require information about the K factor. For $K = 1$ the water pressure is equal to the ice overburden pressure. The basal water would stand with a piezometric surface at a level about 1/10 below the glacier surface if boreholes were drilled down with hydraulic connection to the bed. The actual water pressure may be somewhat lower because of the effects of irregularities in the bedrock and the strength of channel walls counteracting the ice overburden pressure, when water drains in channels.

The K factor will in practice vary in time and with location in a way which is difficult to describe and will as a minimum require extensive measurements of the basal water pressure. Subglacial potentials were therefore calculated for a number of K values varying from zero to one and plotted as potential maps some of which are shown in figs 4 & 5. Drainage along the subglacial topography corresponds to a K value of zero. The calculations are carried out using equation (4) paraphrased to a version that uses ice surface topography and ice thickness as input, thereby using the most original data, by a linear combination of the 100 × 100 m grids. Units for the plotted maps are in $10^3 N/m^2$.

The subglacial basin draining to the ice-free area under conditions of varying K factor was delineated from the potential maps. This includes delineation of sub-basins draining to the three separate lakes 326, 233 and 187 (figs 2, 4 & 5). The subglacial drainage area is slightly sensitive to changes in the K factor. The drainage area and the sub-basins are nearly constant for $K = 0$ to $K = 0.5$. For $K = 0.7$ there is a slight increase of drainage area in the north-eastern part of the basin and for $K = 1.0$ this tendency is further reinforced, but with a marked change in configuration of sub-basins.

SUBGLACIAL TOPOGRAPHY

Electromagnetic reflection (EMR) techniques were used for mapping the ice thickness over the Inland Ice sector. The instrument used is a 300 MHz radar with an antenna consisting of two dipole elements with a half-parabolic reflector especially adapted for mounting between the floats of a Bell 206 Jet Ranger helicopter.

The EMR survey was carried out in two periods in July 1985 and April 1986 (Thorning *et al.*, 1986; Thorning & Hansen, 1987). A total of 22 flying hours was used in each period for actual measurements. The operation in 1985 was originally planned as a series of experiments to evaluate the influence of natural conditions and to test technical configurations of equipment to obtain the best results. However, the results were so encouraging that an actual EMR survey was carried out. This was done without any navigation equipment using only visual navigation between identifiable structures on the ice surface and ice margin. The 1986 operation extended the survey further inland and used Del Norte line-of-sight navigation equipment with a resolution of one metre per second within sight of fixed stations established in the area. The EMR surveys were carried out at a low, constant elevation of about 10 m above the ice surface as the experiments in 1985 showed this to be of vital importance for obtaining results. Under these conditions it was necessary to accept a locally reduced accuracy in navigation in 1985 because no visual points could be identified and in 1986 because of loss of line-of-sight to the fixed stations.

EMR data were recorded on a video tape-recorder and navigation data on tape. Hard copies of EMR data were reproduced and reflections digitized. A package of computer programs was developed for compilation of EMR data including migration of the data and plotting in various stages of the processing. A map and model of the subglacial topography was produced (figs 2 & 3) by combining the topographical ice surface data with the compiled ice thickness data, both data sets transformed into a 100 x 100 m grid.

The hilly ice-free terrain can be seen to continue under the ice with elevations varying from about 300 m below sea level to about 650 m above sea level. A large valley system leads to the big outlet glacier reaching sea level south of the area and can be followed in NE – SW direction for about 6 km behind the ice margin. About 10 km from the ice margin, north-east of lake 233, the terrain is dominated by a high plateau with elevations of 400 – 650 m above sea level. Further inland from this plateau the terrain again drops to elevations below sea level in a large depression.

MASS BALANCE MEASUREMENTS

Stakes for measuring mass balance on the Inland Ice at Paakitsup Akuliarusersua were established in August 1982. The starting point of the stake network is situated on the glacier tongue ending in lake 187 and follows the ice flowline further inland. Initially the stakes were placed at intervals of 200 m altitude, but more stakes were placed later at intermediate elevations. The stakes are aluminum tubes with an outside diameter of 32 mm drilled into the ice with a lightweight motor drill.

The stakes are visited by helicopter every year in the beginning of May and again in late August to measure the mass changes during the winter and summer periods. The measured changes refer to changes in elevation of the glacier surface relative to the stakes. The surface changes are converted to water equivalent on the basis of density measurements in nearby snowpits and by assuming an ice density of 900 kg/m³. Generally the winter snow cover on the ice is very patchy and is confined mainly to drifts in gullies and crevasses up to an elevation of about 500 m, while it is more continuous further inland. As there is no sign that heavy melting occurs during the winter the observed distribution of winter snow is probably due to wind drifting.

At present mass balance has been measured for the years 1982/83 to 1985/86. The transient and annual balances for two selected years are shown in fig. 6. The mass balance data have been used as input for modelling the runoff from the areas.

RUNOFF SIMULATION

Drainage basin areas are converted into runoff volumes by simulations using the MB1 and RO1 models developed by Braithwaite (1984a). The principle is illustrated by the flow diagram in fig. 7 where the MB1 model calculates specific runoff from climatological data and the RO1 model calculates volumetric runoff from specific runoff and the assumed area distribution of the basin.

The MB1 model uses monthly data for air temperature and precipitation extrapolated from Jakobshavn to calculate water balance elements as a function of elevation. Melting of ice and snow is assumed proportional to the monthly sum of positive temperatures at each elevation. The precipitation total is divided into rainfall and snowfall for each elevation on the basis of temperature. New snow is added to the glacier surface which may be either bare ice or old snow, while meltwater and rainfall are absorbed into any snow cover by refreezing until its density reaches a specified critical density, after which runoff is allowed to occur. The time increment for the calculations is one month and the hydrological year is assumed to be from September to August. Running totals of the different water balance elements are stored from month to month and used for calculating annual mass balances at each elevation. The specific annual runoff is finally calculated as the sum of annual ablation and rainfall after allowing for the possible effects of refreezing.

The above model involves a number of parameters, e.g. the inland heating effect and lapse rate for extrapolating temperatures from Jakobshavn, the cooling effect as air moves from ice-free areas to ice-covered areas and the degree-day factor for calculating ablation from positive temperature sums. In the present case, initial values of these parameters were taken from detailed glacier-climate studies at Qamanârssûp sermia, e.g. by Braithwaite (1984b, 1985) and Braithwaite & Olesen (1985) or were found by trial and error. For example, the calculated equilibrium line in the model is reasonably realistic if strong refreezing is assumed but is too high if refreezing is wholly neglected. The model was also found to overestimate specific ablation initially but, by progressively reducing the amplitude of the assumed inland heating effect to only 20 percent of that found for Qamanârssûp sermia, a reasonable agreement was found between the model and observed annual balances. This is illustrated in fig. 8 for the two years 1982/83 and 1983/84.

The calculated specific runoff from the MB1 model is integrated over the assumed glacier area by the RO1 model to calculate the runoff volume. The calculation was repeated for various area configurations, e.g. different versions of the sub-basins for the three lakes 326, 233 and 187 corresponding to different K values and differing interpretations of drainage between sub-basins. A total of 25 different alternatives for the basin area were examined and results were compared with the observed runoff measured by GTO (1986) to find the more plausible alternatives.

The measured runoff for the whole area has a mean of 289 million m³ (GTO, 1986) for the six years 1980 – 1985. The calculated runoff volumes for the same period range from 337 million m³, for $K = 1.0$, to only 227 million m³ for the most pessimistic interpretation of the situation with $K = 0.0$, i.e. a range of +17 to −21 percent of mean observed runoff. Assuming model errors to be of the order of ±10% allows 14 of the 25 alternatives to be rejected as giving either too much or too little runoff. A further four alternatives can be rejected as contradicting indications, admittedly based upon only one year of measurements, that the lake 187 sub-basin should have slightly greater runoff than lake 233. The remaining seven alternatives, involving K values of 0.0 to 0.7, have mean runoffs of −1 to −4 percent of observed runoff.

From the above it seems most likely that the K value under present conditions lies between 0.0 and 0.7, in reasonable agreement with glacier hydraulic measurements made elsewhere, with best agreement between observations and runoff model for $K = 0.7$. The runoff simulation using this assumption for 1961-1985 is illustrated in fig. 9. From the graph it appears that there have been substantial variations in runoff over the past 25 years 1961 – 1985 but the mean runoff for the six years of measurements was only 3 percent less than the 25-year average, indicating that the measurement period is reasonably representative. However, even if climate remains constant, there is also a theoretical possibility of runoff change due to change in hydraulic conditions, e.g. by change in K value. If the present K value is 0.7 with a model mean runoff of 287 million m³ the worst that can happen in the future is the basin will change to a most pessimistic configuration with $K = 0$. In this case the model runoff will be reduced by 21 percent to 227 million m³.

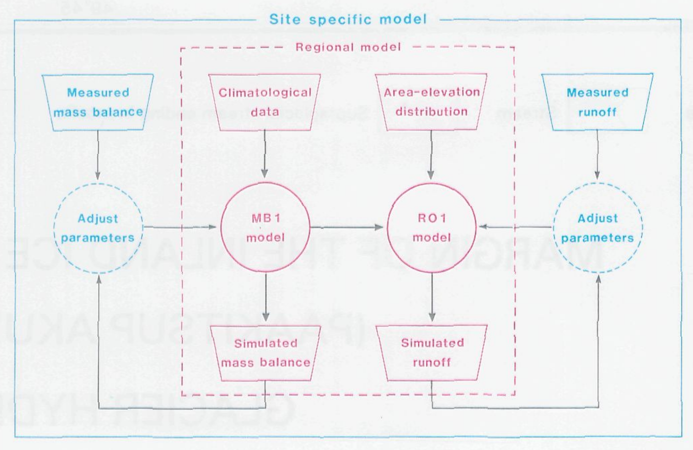

Fig. 7. Sketch of data and principle for simulation of runoff.

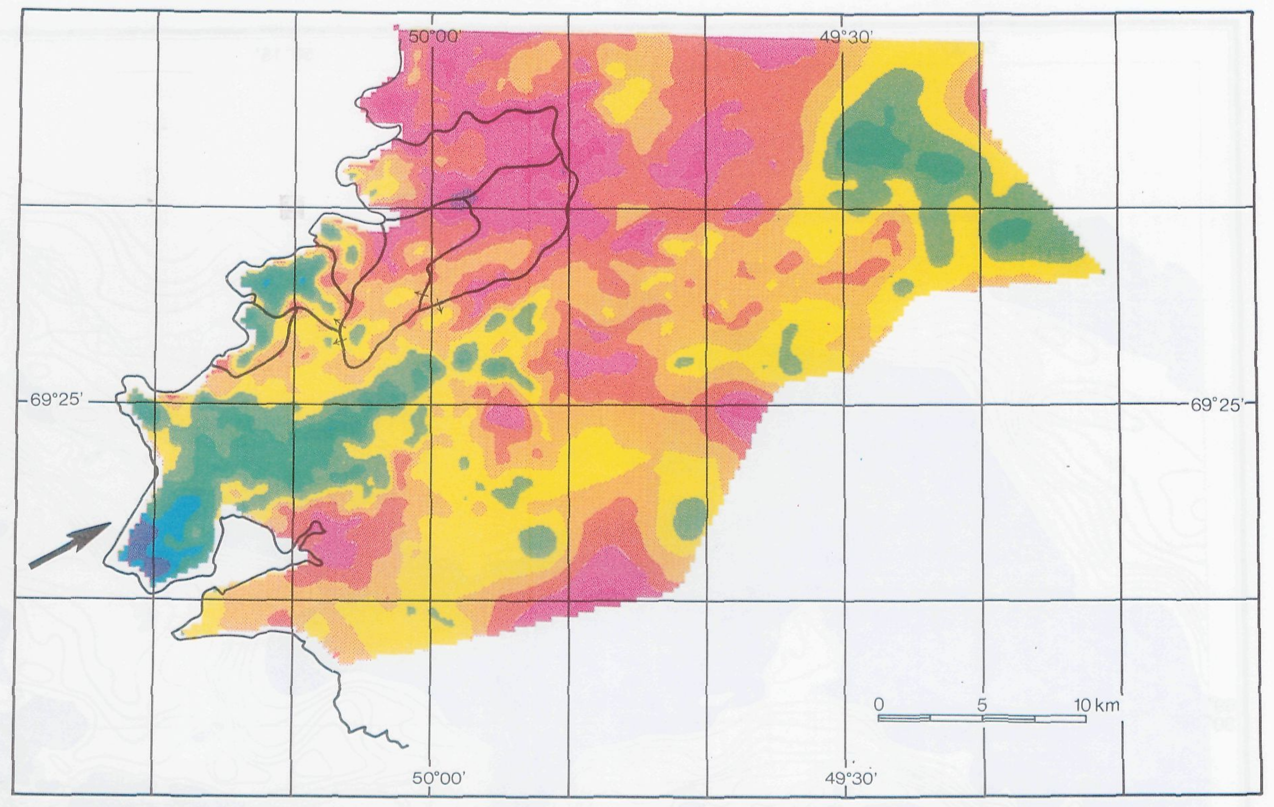

Fig. 2. Subglacial topography. Subglacial water divides are given in black. Drainage alternatives given with small arrows. Heavy arrow shows viewing direction of three-dimensional representation in fig. 3. Elevation in metres above sea level. Colour scale as in fig. 3.

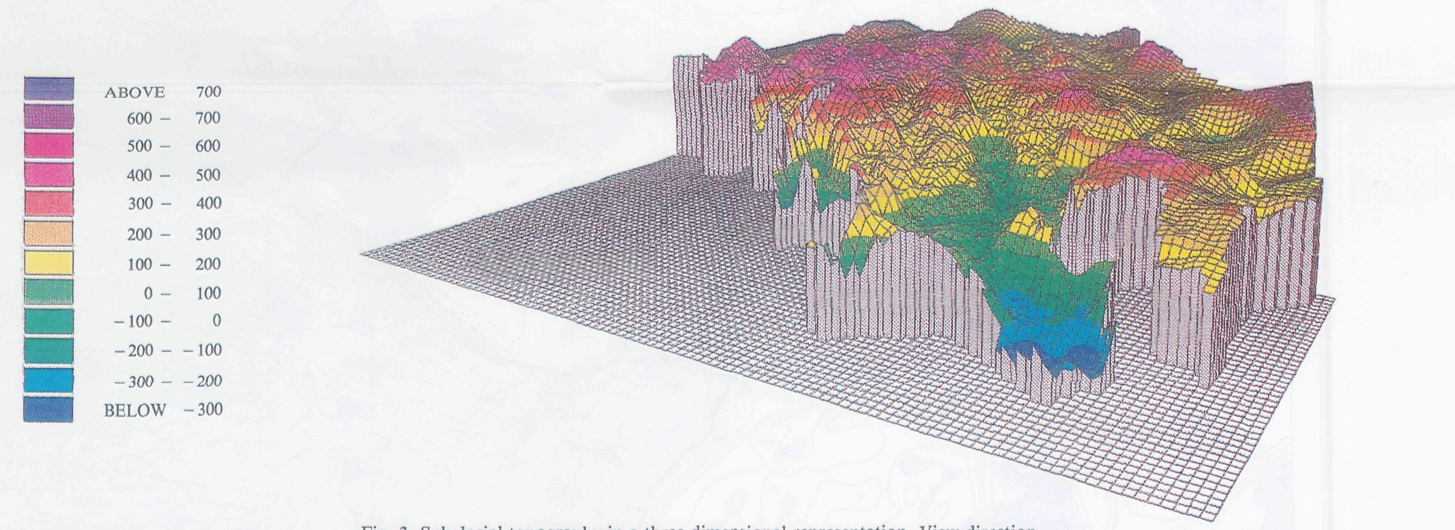

Fig. 3. Subglacial topography in a three-dimensional representation. View direction from south-west shown with heavy arrow in fig. 2.

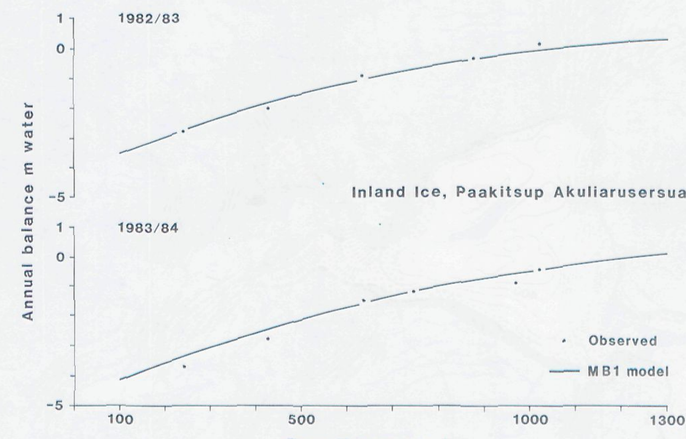

Fig. 8. Observed (dots) and calculated annual balance.

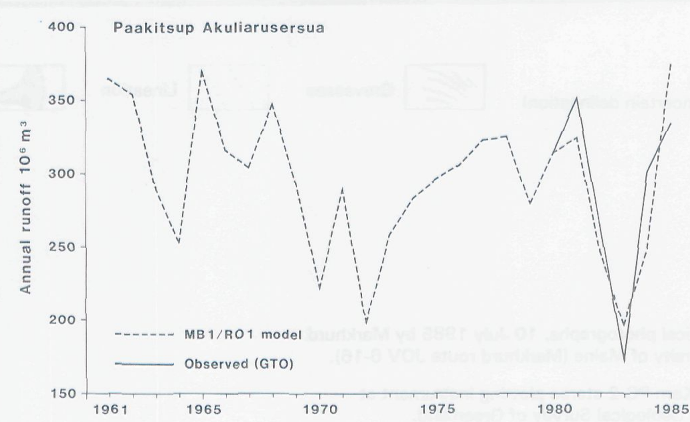

Fig. 9. Observed and calculated runoff.

CONCLUSION

The historical documentation through aerial photographs shows an overall stability of the supraglacial drainage pattern in the area over the last 37 years.

The information gained from mapping and model calculations of the subglacial conditions give a better insight and understanding of the overall drainage pattern when surface water escapes down into the ice.

Calculations of runoff are in good agreement with measured runoff which confirms the drainage basin and repeated runoff calculations show a limited sensitivity of the area to changes in hydraulic conditions expressed in terms of the K factor.

The experience gained through these investigations provide safer limits for planning hydropower in Paakitsup Akuliarusersua, and for a general understanding of glacier hydrology from a continuous ice cover like the Greenland ice sheet.

WORK IN PROGRESS

Work is now in progress to confirm assumptions and increase the understanding of englacial and subglacial drainage. This involves drilling to the glacier bottom with a hot water jet to investigate thermal and hydraulic conditions as well as further detailed EMR surveys and observation of ice thickness on the ice surface with monopulse radar.

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