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Glacier-dammed lake investigations in the Hullet lake area, South Greenland

Alastair G. Dawson

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Results are presented on the evolution of former glacier-dammed lakes and on former glacier oscillations in the Hullet lake area, South Greenland. In this area, the presence of 15 well-developed lake shorelines between 684 and 570 m a.s.l. indicate a complex sequence of Neoglacial lake level changes, several of which may have been associated with catastrophic lake drainage and the rapid emptying of lake waters through a c. 23 km subglacial tunnel beneath Sydgletscher and the Kiagtût sermiat glacier. The oldest lake had a volume of c. 950 × 10^6 m^3 and its sudden drainage resulted in local neotectonic crustal deformation. Lichen measurements of *Rhizocarpon geographicum* on glacial moraine debris suggest that a major Neoglacial expansion of glacier ice had taken place by c. 2350 years B. P. and was followed by widespread ice retreat and stagnation. There is no evidence to indicate an expansion of glacier ice in the Hullet area during the Little Ice Age. The progressive retreat of Sydgletscher was accompanied by the formation of a series of ice-dammed lakes each of which was drained subglacially. The drainage of the lakes is suggested as having been both a slow and a rapid (jökulhlaup) process. Evidence of slow lake drainage is illustrated by the last lake drainage event that took place during October 1981. During this period c. 60% of the volume of lake water drained slowly over a 14 day period with an average discharge of c. 200 m^3s^{-1}.

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The Hullet lake area is located at the margin of the South Greenland ice-sheet in Johan Dahl Land c. 30 km northeast of the settlement of Narssarssuaq (Fig. 1). Within this area most major valleys are occupied by large outlet glaciers that extend from the ice-sheet margin. The largest of these is the Kiagtût sermiat that flows from the ice-sheet to within six kilometres of Narssarssuaq. In its northern part, the valley of the Kiagtût sermiat glacier is connected to a north-south trending tributary valley. This valley is occupied by a glacier, Sydgletscher, that flows northwards from its source in the Kiagtût sermiat trunk glacier. Sydgletscher is therefore unusual since, unlike most glaciers, it flows *upvalley*. The lake Hullet occurs north of Sydgletscher and is impounded by the glacier terminus which forms the southern margin of the lake (Figs 1 and 2). Two outlet glaciers, Nordgletscher and Østgletscher, are located north and east of Hullet. Meltwater drained form these glaciers during spring and summer is discharged into the lake and causes it to rise. The lake also receives water from local streams, by precipitation and from the calving into the lake of icebergs at the snout of Sydgletscher.

Observations of changes in the level of Hullet have indicated that since 1960, the lake has been subject to periodic drainage (Weidick 1963, Brathay 1969, Clement 1982).

During each period of lake drainage, the lake is emptied beneath Sydgletscher and Kiagtût sermiat glacier along a c. 23 km subglacial tunnel. The last drainage of Hullet occurred during October 1981 and resulted in flooding of the lower Narssarssuaq valley and the discharge of large volumes of freshwater into the adjacent fjord. In the Hullet area, numerous glacial moraines and lake shorelines indicate that the size of Sydgletscher and the location of the ice-dammed lake Hullet have varied considerably since the Holocene climate optimum. In the following pages, evidence is presented for the sequence of former glacier oscillations and the development of ice-dammed lakes in this area.

Techniques and field methods

Prior to fieldwork, published maps of the Hullet area (Brathay 1969, 1980) were enlarged to 1:5000 scale. Owing to the changing level of Hullet, the mapped perimeter of the lake was drawn as the 474 m level measured by Brathay on 29th July, 1969. All ice-marginal landforms were initially mapped from aerial photographs supplied by the Brathay Exploration Trust, the Department of Geography, University of Aberdeen and the Geodetic Institute, Copenhagen. Thereafter, the

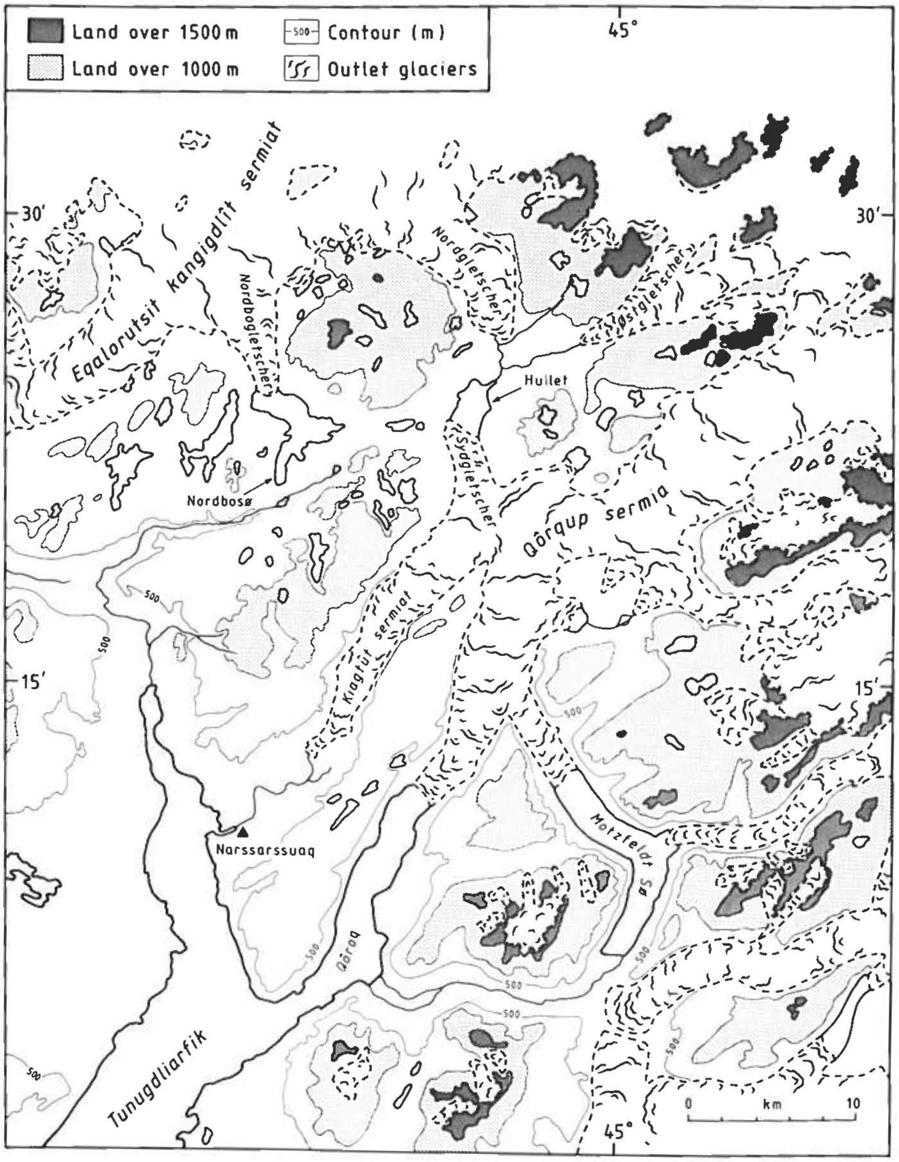

Fig. 1. Simplified topographic map of the Narssarssuaq–Hullet lake area, South Greenland, based on 1:250000 sheet 61 V.3 Narssarssuaq published by Geodætisk Institut, Copenhagen.

mapping of landforms was checked and amended in the field.

During fieldwork (August 3-23, 1982) the altitudes of all ice-dammed lake shorelines were determined by instrumental levelling using a Kern NAK2 level and a metric staff. All traverses were closed and where closing errors exceeded 0.05 m, the traverses were repeated. The datum level used in the research was selected as a cairn established by the 1969 Brathay Expedition on the main abandoned ice-dammed lake shoreline adjacent to Langesø (Fig. 4). Brathay (1980) re-surveyed the altitude of the cairn base by aneroid barometer and confirmed a value of 684 m a.s.l. Numerous temporary bench marks were established throughout the field area and all of these were closed with respect to each other and to the original datum. In order to calculate daily changes in the water level of Hullet a temporary bench mark was established adjacent to the lake. Thus, the lake surface altitudes were related to a local datum.

Drainage of Hullet – October 1981

In order to calculate the volume of the lake Hullet prior to its October, 1981 drainage, the following information was used. The only accurate map of Hullet was published by Brathay (1980), who showed that on 29th July, 1969, the lake had an area of 5.7 km² and an altitude of 474 m. We assume the same lake area for the

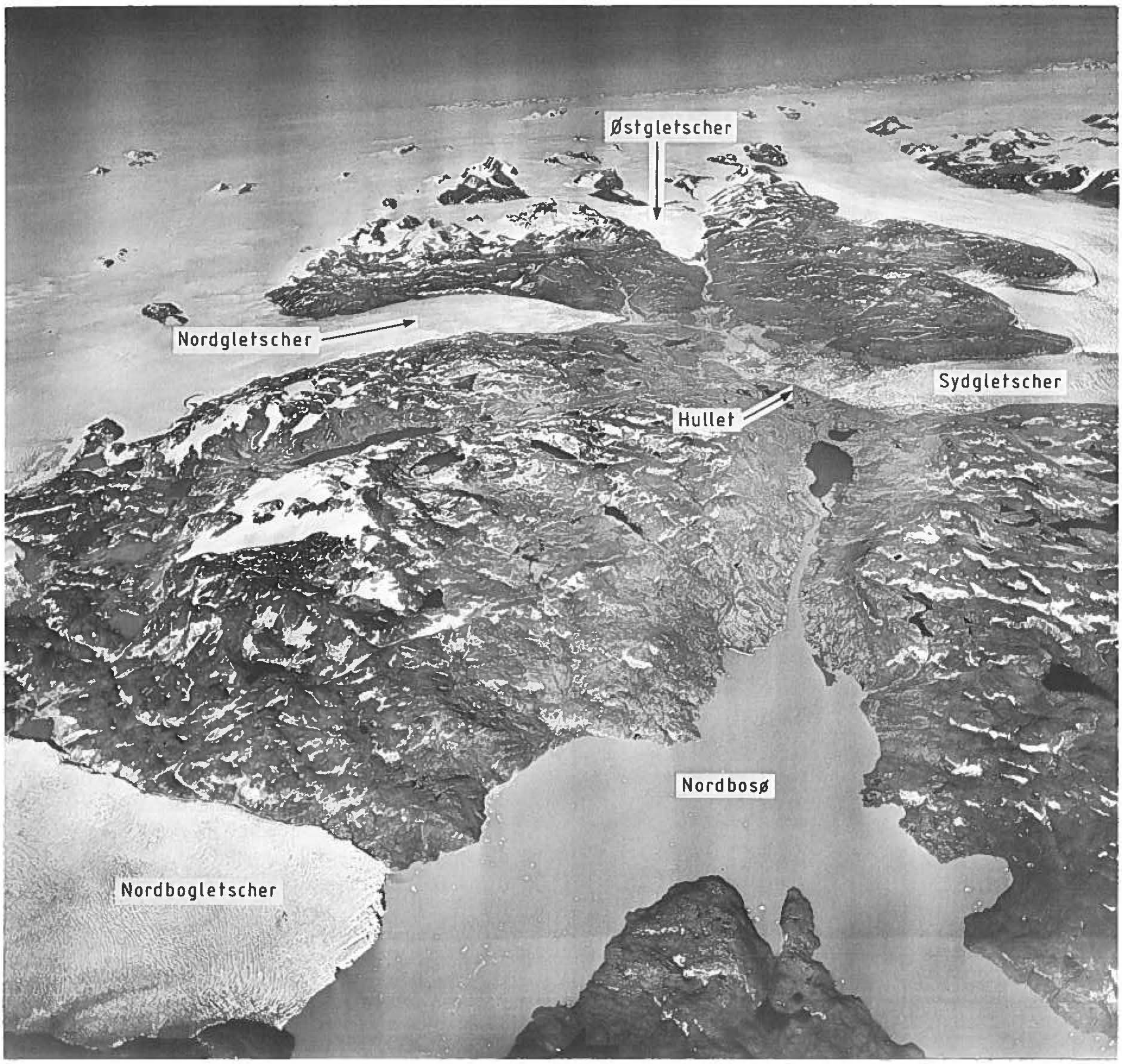

Fig. 2. Oblique aerial photograph of Johan Dahl land and the Hullet lake area, South Greenland. Geodætisk Institut, Copenhagen.

first 1982 survey of Hullet (4th August, 1982) when the lake had an altitude of 475.4 m. The maximum level of the lake immediately prior to the October, 1981 drainage is indicated by the upper limit of stranded icebergs located c. 40–45 m above the August 1982 lake levels. The crests of the highest group of stranded bergs occur at 520–521 m and are continued elsewhere around the lake by an abandoned shoreline below which are thick accummulations of lake sediments. Since the highest level of stranded bergs measured by Brathay during 1969 was 516 m, it would appear that the lake has regularly attained similar altitudes prior to its drainage during the last c. 20 years. These maximum altitudes also correspond with altimeter readings (515 m and 518 m) of maximum lake levels measured by Grønlands Geologiske Undersøgelse (Clement 1982, p. 60) on 22nd August, 1980 and 24th September, 1981.

Minimum lake level altitudes have been documented for 30th June, 1960 (Weidick 1963) and 3rd October, 1980 (Clement 1982). For example, during July, 1960, Weidick observed that the lake was almost entirely drained of water and that its level was c. 430 m. Hence, in order to calculate a minimum volume of the lake prior to drainage, it was assumed that:

(a) the 475 m lake has an area of 5.7 km^2
(b) the maximum lake level was 521 m
(c) the lake bed is flat at 440 m

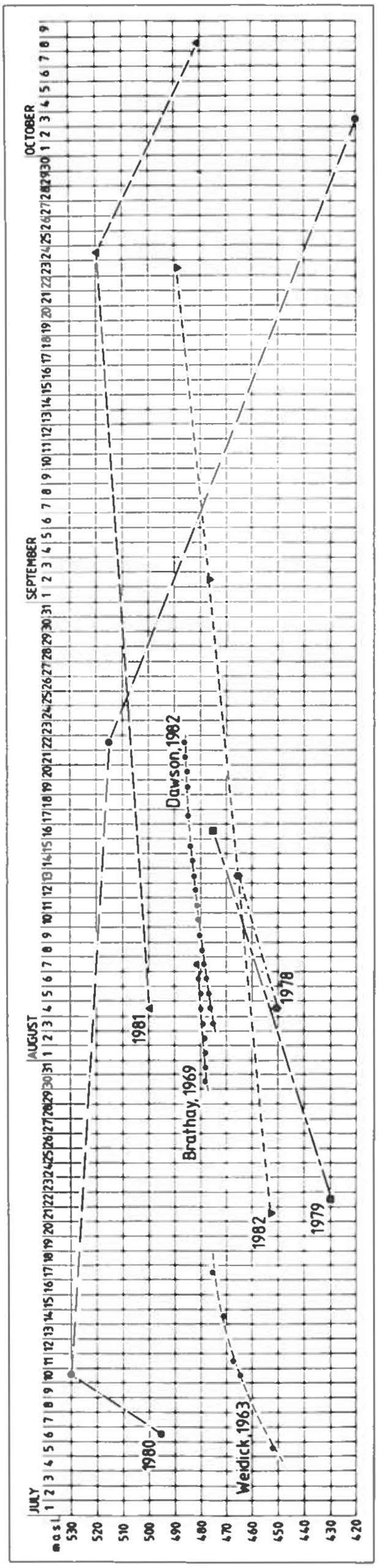

Fig. 3. Measured changes in the level of the glacier-dammed Hullet lake since 1960 based on Weidick (1963), Brathay (1969), Clement (1982) and this author.

(d) the average lake perimeter slope is 20°
(e) 5% of the lake volume is displaced by icebergs

The results indicate a lake volume of c. 397 × $10^6 m^3$. This value is suggested as a minimum lake volume prior to drainage since:

(1) the lake floor locally occurs below 440 m and
(2) the iceberg displacement volume may be too high

Observations on the changing levels of Hullet during June, 1960 (Weidick 1963) indicate that during this period, complete lake drainage occurred. Drainage of ice-dammed lakes may often occur as a catastrophic discharge of water during a short period of time. The catastrophic drainage, or jökulhlaup, of an ice-dammed lake is imperfectly understood. However, the principal conditions are:

i) that a subglacial tunnel is present beneath the ice so that the lake waters may escape. Under certain conditions, ice-dammed lakes may drain across the glacier surface or along its sides. However, in the case of Sydgletscher, lake drainage has always taken place beneath the ice (Weidick 1963).
ii) that the ice-dammed lake is of sufficient depth to float the barrier of ice that dams the lake. In this way the lake may undermine the glacier margin and eventually connect with a subglacial tunnel.
iii) that during lake drainage, thermal erosion of ice by meltwater in the subglacial tunnel results in the enlargement of the tunnel and hence enables the rapid discharge of the escaping meltwater.
iv) that lake drainage eventually leads to the collapse of the ice barrier which may cause blockage of the lake outlet. Diminished water flow within the tunnel (either as a result of blockage or after complete lake drainage) results in tunnel closure by ice pressures.

Although rates of lake level rise are known for several periods (Fig. 3), the rate at which Hullet is drained has only partly been documented. That the lake drained completely between 22nd August and 3rd October, 1980 is indicated by the reported fall in lake level from 515 m to 421 m during this period. However in 1981, between 24th September and 8th October, partial lake drainage took place from 518 m to 481 m (Clement 1982). The occurrence between 518 m and 521 m of the highest stranded icebergs indicates also that the lake did not exceed these altitudes before falling to 481 m. During this period of lake drainage approximately 235 × 10^6 m^3 of water was released from the lake and escaped subglacially beneath Sydgletscher. Hence the *average* rate of lake drainage during this period was c. 200 m^3s^{-1} and is equivalent to an *average* lowering of the lake level by 11 cm/hour. Evidence that the lake lowering from 518–521 m to 481 m was associated with a slow pattern of drainage rather than by a catastrophic outburst, is also indicated by a staircase of c. 14 shorelines at the northern end of Hullet that extend from 521 m to the present lake level (475.4 m on 4th August, 1982). All of the shorelines could only have been produced during the last (1981) lake drainage event. Their formation during a period of slowly (or pulsed) falling lake level is therefore indicated. Thus 14 separate shorelines were produced in fourteen days, during which period the lake was subject to a progressive and slow lowering. Indeed, the reconstructed lake volumes show that by 8th October, 1981, over 60% of the initial lake volume had been drained. By inference, the subglacial tunnel outlet was not opened sufficiently to permit a major jökulhlaup.

The above interpretation contrasts with many published accounts of other ice-dammed lakes whose observed patterns of drainage are catastrophic (see Clague & Mathews 1973, Blachut & Ballantyne 1976). Indeed Clague & Mathews suggested that a good relationship exists between the volume of water drained from an ice-dammed lake and the peak flood discharge of the water on its emergence at the glacier snout. The relationship proposed by them is based on measurements from ten ice-dammed lakes and is expressed as

$$Q\max = 75\ V\max^{0.67}$$

where Qmax = the peak flood discharge (m^3s^{-1})
Vmax = the total volume of water drained from the ice-dammed lake (in $m^3 \times 10^6$)

Application of this relationship to the 1981 Hullet drainage suggests a peak water discharge at the snout of Kiagtût sermiat glacier of c. 4130 m^3s^{-1}. Although no information is available for the rate of lake drainage after the lake had fallen to 481 m, when 40% of the original lake volume remained undrained, it would appear that the Clague & Mathews (1973) drainage model does not correspond with the inferred pattern of the 1981 drainage of Hullet.

Changes in the lake level of Hullet: August 4th–22nd, 1982

The 1982 Hullet Survey indicated that the lake altitude on the first day of measurement was 475.4 m. (4th August, 12.35 hours). Hence by assuming the basin shape and area parameters as outlined above, daily increases in lake level measured during August 1982 were calculated as equivalent volumes of water added to the lake. Table 1 shows the measured increases in the level of the lake Hullet for each 24-hour period from 4th–22nd August. Also shown are the estimated volumes of water entering the lake for each 24-hour period. These values have been adjusted for the changing shape and volume of the lake for each lake level rise. In this way the mean daily discharge of water into the lake was calculated. These values represent the sum of:

(a) the combined discharge of the Nordgletscher and Østgletscher meltwater rivers,
(b) the combined discharge of other minor rivers entering the lake,
(c) daily precipitation,
(d) any rise in lake level caused by water displacement accociated with iceberg calving from Sydgletscher.

It is assumed that during this period of measurement, there was no subglacial leakage of lake water beneath Sydgletscher. Indeed, any leakage that may have taken place during this period would increase the calculated daily discharge values.

Moreover, since it is considered that 5% of the lake volume is occupied by icebergs, the calculated discharge values may be too high (probably by 5–10%). The data exhibit several trends. Firstly there is a progressive decline in discharge throughout August. Since most discharge is derived from Nordgletscher and Østgletscher meltwater, the decline is considered to be related to diminishing ablation over the ice-sheet. Superimposed on this long term trend is a high daily variability in discharge values (for example 14–15 August and 20–21 August). This is partly due to the periodic occurrence of föhn winds that drastically increase ablation rates, and to periodic tapping and damming of small subglacial lakes beneath Nordgletscher.

The latter explanation, for example, is considered

Table 1. Surveyed lake-level rises of Hullet, August 4–22, 1982. The daily volumes of water added to the lake ($\times 10^6\ m^3$) and the daily average discharge into the lake (m^3s^{-1}) are also given.

Date (1982)	Cumulative lake level rise (m)	Volume increase ($\times 10^6 m^3$)	Average Q (m^3s^{-1})
August 4	1.20	6.799	77.9
August 5	2.00	4.543	52.2
August 6	3.03	5.888	69.3
August 7	3.70	3.839	44.4
August 8	4.37	3.860	44.7
August 9	4.92	3.166	36.6
August 10	5.64	4.169	45.0
August 11	6.22	3.359	41.9
August 12	7.06	4.895	48.9
August 13	7.50	2.563	34.6
August 14	8.24	4.331	43.9
August 15	8.58	1.994	26.5
August 16	–	–	20.0*
August 17	9.20	3.644	20.0*
August 18	–	–	37.2*
August 19	10.24	6.143	37.2*
August 20	10.44	1.173	13.2
August 21	10.96	3.084	37.1
August 22	11.48	3.091	34.5

* these values are calculated as averages.

responsible for the marked variations in average discharge between August 15th and 21st. On 15th and 19th August, large sections of the ice cliffs which form the snout of Nordgletscher were observed to collapse onto the proglacial zone. The ice blocks which collapsed along the western section of Nordgletscher resulted in the damming of the adjacent meltwater river and its diversion underneath the glacier. On both occasions, the ice falls were followed by major declines in the stage level of the Nordgletscher river and a simultaneous decrease in the rate of rise of the Hullet lake. Thus the process of river diversion and the decline in the stage level of Nordgletscher indicate clearly that the pattern of river diversion was accompanied by meltwater storage beneath Nordgletscher. On the 18th and 21st August, sufficient melting of the ice boulders had taken place to permit the restoration of the original drainage. Since the latter events were accompanied by major flooding of the Nordgletscher river along its original course, it is inferred that the source of the excess meltwater was produced by the tapping of the subglacial lakes which were filled during the preceding phases of river diversion. The maximum daily rate of lake level rise in Hullet was 1.2 m/day (4th August), while the total lake level rise during the period of observation was 11.5 m. These values are stressed since they clearly indicate the dynamic changes that take place in ice-dammed lake environments.

Former glacier oscillations in the Hullet area

The existence of a climatic deterioration after the mid-Holocene climatic optimum has been recognised in West Greenland (Kelly 1980) and was characterised by an increase in glacierisation. Within the Hullet area, numerous fossil lateral and terminal moraines testify to former glacier oscillations that took place after the climatic optimum. At present knowledge of these events is poor since the ages of most moraines have not been determined. The most comprehensive account of glacier oscillations in this area is by Weidick (1963) and is briefly summarised below (see also Fig. 4).

Weidick (1963) proposed that the main expansion of glacier ice accompanied the formation of the Narssarssuaq moraines which he subdivided as having been formed during stages I, IIa and IIb. At Narssarssuaq the presence of Norse buildings on outwash plains belonging to the stage II moraines (Weidick 1963, p. 51) indicates that these moraines are almost certainly older than those produced elsewhere during the Little Ice Age. Moreover, Kelly (1980) has argued on palynological grounds that the Narssarssuaq moraines were produced between 1000 and 2500 years B. P. Weidick, however, (1963, p. 31) proposed that the stage I ice advance was older than the Holocene climatic optimum and that this

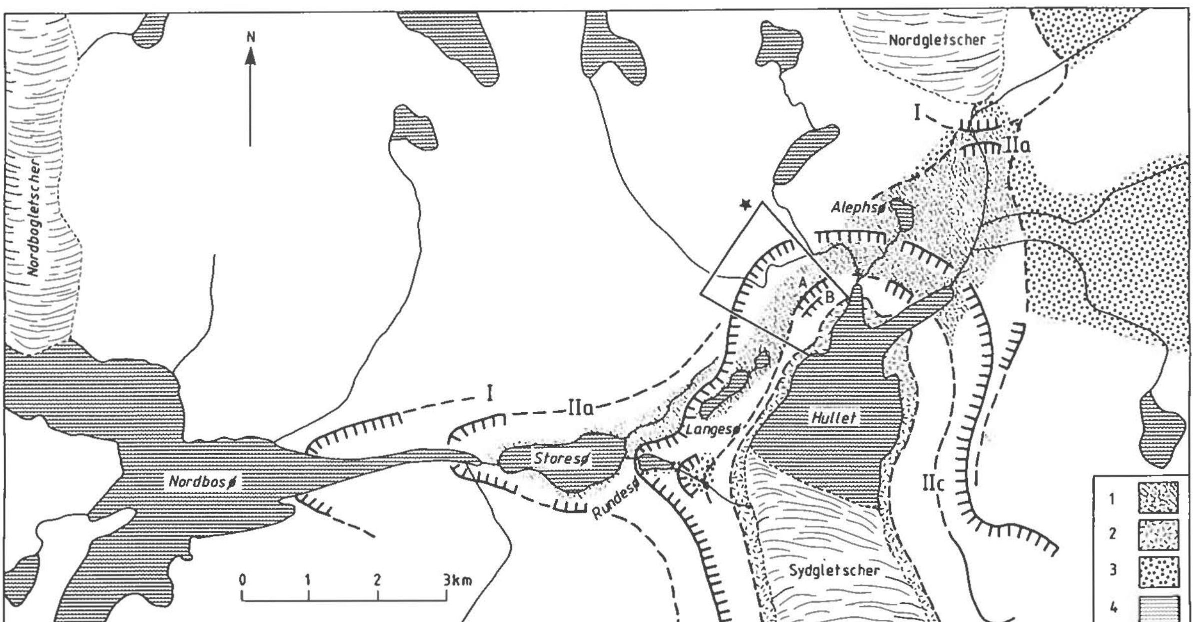

Fig. 4. Glacier limits in the Hullet area as presented by Weidick (1963, fig. 11). 1. Lake and river deposits. 2. Sydgletscher trim line zone. 3. Terraces (undifferentiated). 4. Lakes. The area shown by an asterisk represents the area from which measurements were undertaken on the lichen *Rhizocarpon geographicum*. A and B represent sections of the stage IIc West Hullet moraines (see text, Fig. 12 and Table 3).

was followed by an intermediate period during which the ice had a more limited extent than at present. Thereafter a major Neoglacial ice advance occurred during stage II.

Following stage II, there was a recession of the ice which was succeeded by a readvance of ice in historical times. Weidick suggested that evidence for the latter advance occurs at Sydgletscher where historical observations of ice dimensions and the presence of a trim line zone above the glacier margin indicate that the glacier retained its maximum extension until c. 1900. Thereafter significant ice retreat took place.

It should be stated that Weidick (1963) stressed that most of the SW Greenland glaciers responded differently to climatic change during the Neoglacial. Hence it is extremely difficult to correlate moraines between different glaciers. This is clearly illustrated by recent glacier oscillations in the Hullet area. For example, Nordgletscher has advanced 600 m since 1947 (Clement 1982) and similarly, c. 12 km farther west, Nordbøgletscher has advanced 665 m since 1942 (Clement 1982). In contrast Sydgletscher has remained essentially stationary since 1960 (cf. Weidick 1963, Brathay 1969, 1980) and may even have retreated.

In the following pages new evidence is presented on former glacier oscillations in the Hullet area. The information is presented within the context of stages I, IIa and IIb proposed by Weidick. The discussion also attempts to examine critically the field evidence on which Weidick's relative chronology is based.

Stage 1

Weidick (1963) suggested a distribution of stage I ice as shown on Fig. 4. He described a series of terminal and lateral moraine fragments at the eastern end of Nordbosø lake and proposed that they were produced by an outlet glacier that flowed westwards from Sydgletscher. The presence of the outlet glacier at Storesø and eastern Nordbosø during stage I requires a more advanced position of Sydgletscher. However, stage I moraines produced by Sydgletscher have not been identified, Weidick having suggested that they were eroded during the stage II advance of Sydgletscher ice. The position of Nordgletscher during stage I is uncertain. Weidick (1963, p. 29) described an exposure south of the 1963 Nordgletscher snout that he considered as indicative of a stage I ice advance (Table 2).

Weidick assigned units A and B to a glacial advance during stage II and unit D as part of a moraine produced during stage I. Consequently, unit E was proposed as having been deposited in an ice-dammed lake impounded between Sydgletscher and Nordgletscher prior to the stage I ice advance. This section has now been destroyed by the recent advance of Nordgletscher. However, numerous sections of glacial deposits are presently exposed opposite the snout of Nordgletscher. At numerous places, isolated lenses (up to 0.7 m thick)

Table 2. Lithostratigraphic units from the front of Nordgletcher described by Weidick (1963, p. 29).

Depth			Unit
0–0.5 m	Sandy till with large rounded boulders		A
0.5–1.5 m	Sandy till, sometimes passing into cross-bedded alluvial deposits with few rounded boulders	glacial beds	B
1.5–1.8 m	Red layer of large rounded boulders with a sandy matrix, cemented together by hematite	glacio-fluvial beds	C
1.8–2.1 m	Sandy-gravelly cross-bedded layers with sub-angular boulders		
2.1–2.7 m	Sandy till, exclusively with subangular boulders	glacial beds	D
2.7–6.5 m	Cross-bedded sand with rounded boulders, passing gradually downwards into almost varved silt and clay	glacio-fluvial beds	E
	Base		

of laminated silts and clays are both overlain and underlain by glacial till. In the Hullet area, such laminated deposits may accumulate in:

(a) ice-dammed lakes,
(b) in subglacial water bodies,
(c) in topographic depressions where fallen blocks of ice have collapsed from the glacier snout onto the adjacent ground surface (e.g. at the snout of Nordgletscher).

As a result, the presence of laminated sediments need not indicate the former presence of a *large* ice-dammed lake. Moreover, the presence of glacial till deposits should not be considered synonymous with an end moraine.

South of the present snout of Nordgletscher, the presence of well-developed terminal and lateral moraines indicates that Norgletscher reached a more advanced position during stage I or II (Fig. 4). The clearest feature is an arcuate end moraine located on the interfluve between the Nordgletscher and Østgletscher rivers. The end moraine is countinued west of Nordgletscher river by a large lateral moraine fragment that is continuous as far as the snout of Nordgletscher. The former movement of ice towards the position of the end moraine is also supported by the orientation of fresh glacial striae (160°) recorded on ice-moulded bedrock exposures in Nordgletscher river. The end moraine is succeeded upslope by shorelines of an ice-dammed lake produced during stage IIb. Thus formation of the end moraine during stage I or II requires that the snout of Nordgletscher terminated in an ice-dammed lake and that the moraine could only have been produced by the collapse of the glacier snout onto the floor of a drained ice-dammed lake.

Stage IIa

There is no morphological evidence to indicate whether the outlet glaciers receded between stages I and II (Weidick 1963). However, if the intervening period is represented by the Holocene climatic optimum, this would appear likely. The ensuing stage II ice advance was suggested by Weidick as representing a clear landscape unit. The earliest ice advance of this stage (IIa) is indicated by an end moraine and adjacent lateral moraine at the western end of Storesø (Fig. 4). In this area the end moraine is c. 30 m high and is succeeded westward by outwash terrace fragments that decline in altitude towards Nordbosø. Weidick (1963, p. 32) suggested that during stage IIa, Sydgletscher extended northwards to almost the present position of the Nordgletscher snout. He argued that Sydgletscher impounded an ice-dammed lake during this period and proposed that "kame terrace fragments" south of Østgletscher represented the margin of a 700 m lake that ultimately drained via Storesø into Nordbosø. It is suggested here that this interpretation requires modification. The "kame terrace fragments" described by Weidick on both sides of the Hullet valley occur at a uniform altitude of 684–685 m. The morphology and distribution of the features and their regionally uniform altitude indicate clearly that they are *shoreline* fragments of an abandoned ice-dammed lake. However, on the west side of the Hullet valley, the shoreline is eroded in glacial deposits produced during stage IIb (see page 12). Hence the 684 m lake was produced during or after stage IIb. There are no fossil shorelines above 684 m to indicate the existence of a similar ice-dammed lake during stage IIa. It is therefore likely that during stage IIa, Sydgletscher and Nordgletscher were contiguous ice masses.

Stage IIb

Weidick (1963, p. 32) considered that stage IIb represents the main Neoglacial stage in the Hullet area. During this period, a well-developed end moraine c. 40 m high was produced at the snout of Sydgletscher (Fig. 5). The end moraine extends almost continuously across the Hullet valley and is located at Alephsø c. 4.5 km north of the present snout of Sydgletscher (Fig. 4). The end moraine is continued on both sides of the Hullet valley by well-defined lateral moraines that can be traced intermittently over c. 22 km to the Narssarssuaq area. On the east side of the Hullet valley, lateral moraine fragments produced during this period extend as far as Kiagtût sermiat glacier. On the west side of the Hullet valley at Rundesø (Fig. 4) the lateral moraine passes into a terminal moraine, the western margin of which is located between Rundesø and Storesø lakes. That moraine deposition at Rundesø was complex is indicated in several areas by triple end moraine ridges. The continuity and size of the lateral and terminal moraines over a wide area leave little doubt that they were deposited during a glacial advance associated with

Fig. 5. Photograph of the Hullet area showing the lake, Sydgletscher and in the distance the Kiagtût sermiat glacier. The stage IIb end and lateral moraine complex of Sydgletscher is also a clear feature and is visible as far as the slopes above the lake. The trim line zone is also clear and is shown as a horizontal line on the slopes above the lake.

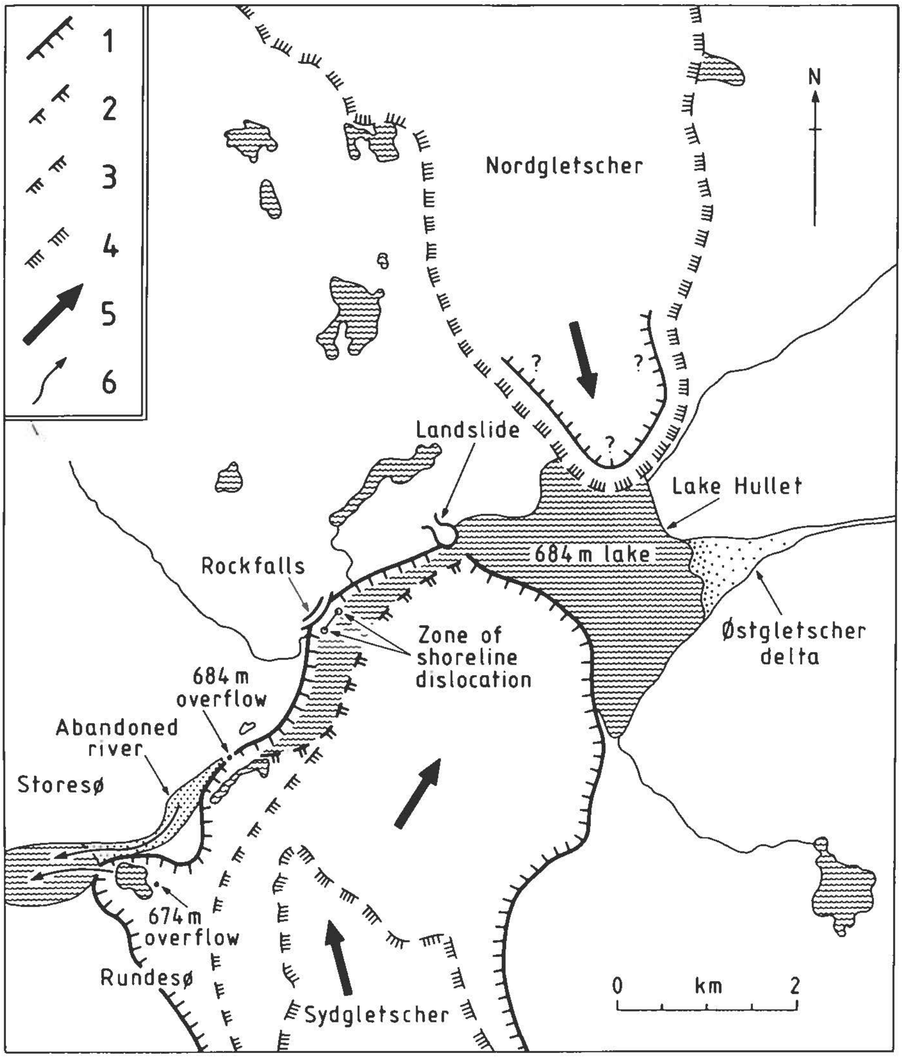

Fig. 6. Reconstruction of the Hullet area during the 684 m and 674 m lake phases. The locations of the 684 m and 674 m overflow channels are also shown. Owing to the recent advance of Nordgletscher, its position during the 684 m lake phase is not known. The locations of the landslide, the zone of shoreline dislocation and the major area of rockfall are also shown. 1. stage IIb ice-limit; 2. Location of stage IIb ice margin during maximum expansion of 684 m lake; 3. Location of stage IIb ice margin during 674 m lake phase; 4. Positions of Sydgletscher and Nordgletscher, 1982; 5. Ice-flow directions; 6. Overflow channel routes.

the expansion of the Kiagtût sermiat glacier, Sydgletscher and the Rundesø glacier lobe. Together the stage II moraines may be defined as the Narssarssuaq moraines (Weidick 1963).

During stage IIb a large ice-dammed lake was impounded between Sydgletscher, Østgletscher and Nordgletscher (Fig. 6). The lake margin during this period is defined by a well-developed shoreline at c. 684 m which is virtually continuous on both sides of the Hullet valley. The shoreline is locally up to 50 m wide and at several locations is backed by a rim of lake ice-pushed boulders. Discussion on the evolution of this lake (hereafter referred as the 684 m lake) is presented later. Southeast of the present Nordgletscher snout, the 684 m shoreline merges with a large fossil delta (Fig. 6), the inner edge of which occurs at the same altitude as the shoreline. The delta displays foreset and topset bedding and its altitude clearly indicates that the feature was deposited in the 684 m lake. The source of the sediment of which the delta is composed is therefore derived from meltwaters that issued from Østgletscher during the existence of the 684 m lake. Hence during stage IIb, the position of Østgletscher varied little from its present position.

Glacier oscillations after stage IIb

The former existence of an ice-dammed lake that was subject to numerous changes in lake level and which was contemporaneous with the formation of the Sydgletscher end moraine south of Alephsø is demonstrated by sets of abandoned lake shorelines eroded in the distal slopes of the end moraine. The contemporary presence of Sydgletscher ice during the existence of the lake is suggested by the absence of shorelines at similar altitudes inside the moraine. However, at lower altitudes, lake shorelines are eroded on both sides of the end moraine and extend considerable distances southward. The most conspicuous of these shorelines is at 577 m and extends as far southward as a second end moraine located between Hullet and the stage IIb Sydgletscher end moraine (Fig. 4). The end moraine and associated lateral moraine are best developed west of Hullet where they consist of a vegetated ridge c. 5 m high and c. 800 m long. South of the moraine, the 577 m shoreline is absent. Therefore it is suggested that this lake was produced following a glacial retreat of Sydgletscher from its stage IIb to IIc positions. Two brief glacial stillstands (IIb_1 and IIb_2) interrupted the general retreat of ice between stages IIb and IIc. Interpretation of these stillstands is based on evidence of former lake level changes. These are discussed on page 19.

South of the stage IIc moraine, the only other end moraine is reported as located on the floor of Hullet (Weidick 1963, pp. 36–37). Weidick suggested that this feature was produced around 1900 when Sydgletscher attained a more advanced position than at present. Weidick (1963, pp. 35–38) correlated the end moraine with a well-defined trim line zone that occurs parallel to and on the slopes above the present margin of Sydgletscher. Weidick therefore interpreted the trim line zone as indicative of the margin of Sydgletscher at c. 1900. The trim line zone occurs between 556 m and 571 m (Weidick (p. 39) quotes 600 m) around Hullet and is continuously developed at this level. However, the 556–571 m altitude zone also coincides with maximum levels of former ice-dammed lakes impounded by Sydgletscher during and prior to 1900. It is therefore argued (see p. 19) that although the trim line zone is a true ice-marginal landform elsewhere in this area, the feature also represents the upper zone of lake deposits around Hullet. It is also proposed that the position of the c. 1900 glacial trim line zone around Hullet cannot be determined but presumably is located at and below the level of the lake deposits. Hence it is clear that the investigation of various levels attained by former ice-dammed lakes in the Hullet area is critical to the reconstruction of former glacier oscillations in this area.

Former ice-dammed lakes

During the expansion of Sydgletscher during stage IIb a large ice-dammed lake was impounded by the glacier (Fig. 6). The altitude of the lake is represented by a well-developed shoreline at 684 m. The lake possessed an area of c. 10 km^2 and had an estimated volume of c. 947×10^6 m^3 of water (approximately 2.4 times the present volume of Hullet). The lake shoreline is locally up to 50 m wide and is virtually continuous around the margin of the former lake. The shoreline is generally eroded in unconsolidated Quaternary sediments but locally (e.g. between Nordgletscher and Alephsø) the feature is eroded in granite bedrock. On the eastern side of Nordgletscher, the shoreline extends as far as the present ice-margin. Farther east, the shoreline merges with the Østgletscher delta that resulted from the progradation of fluvioglacial sediments into the 684 m lake. Notably, the shoreline terminates at the stage IIb end moraine northeast of Hullet. In the centre of the Hullet valley, the end moraine is everywhere located below 684 m. Hence the absence of the 684 m shoreline inside the stage IIb ice limit on the east side of the Hullet valley indicates the contemporary presence of active Sydgletscher ice during stage IIb over the eastern and central parts of the Hullet valley (Fig. 6).

On the western side of the Hullet valley, the 684 m shoreline is present both inside and outside of the stage IIb ice-limit (Fig. 6). For example, southwest of Alephsø the 684 m lake shoreline is a continuous feature and is eroded in the proximal and distal slopes of the end moraine. Thereafter, the feature is continued southward for 3 km and is eroded on the inside of the stage IIb lateral moraine. The shoreline extends as far south as Langesø where it merges with an abandoned lake overflow channel (Fig. 6).

The shoreline is best preserved between Langesø and Alephsø lakes (Fig. 4) where it is backed by lake ice-push boulder ramparts (Fig. 7). The shoreline surfaces are generally flat and are usually bare of vegetation. Notably the shoreline fragments are free of local topographic depressions and undulations. At three locations, however, exceptions to this pattern occur. Between Langesø and Alephsø (Fig. 4), the continuity of the shoreline is interrupted by three large kettle hole depressions. The features are c. 40 m deep and are 85 m, 100 m and 145 m wide respectively. The kettle holes possess open eastern flanks and are continued downslope by widespread boulder accumulations. The bases of the kettle holes occur at c. 640 m and, significantly, their inner slopes, despite being partially colonised by vegetation, are not notched by any shorelines (i.e. between 640 m and 684 m).

The overflow channel near Langesø of the 684 m lake (Fig. 6) breaches a series of lateral moraine fragments produced during stage IIb. The channel surface occurs at 684.6 m and passes into an abandoned braided river that functioned as the overflow route for the 684 m lake. The abandoned river is ca. 2 km long and up to 500 m wide and consists of a series of anastomosing channels separated by channel bars consisting of large boulders. The channel bars are locally pitted by circular and oval depressions generally 1–2 m wide and 1 m deep. The most likely origin of these features is that they resulted from the deposition and melting of river ice blocks during periods of flood. This interpretation is supported by observations of recent river ice deposition in the proglacial river at the snout of Nordgletscher during periods of high river stage. During these periods, calved ice blocks are eroded by fluvial processes into c. 1 m^3 ice boulders and deposited on river bars. Thereafter the boulders were observed to melt *in situ* producing topographic depressions within the channel bar deposits.

Fig. 7. Fragment of 684 m shoreline near Langesø. The inner edge of the shoreline is defined by a boulder rampart formed by lake ice-push processes. Note that several of the boulders exhibit pronounced "lichen lines" that separate the rock surfaces where lichen growth only commenced after lake drainage from those where growth commenced before the shoreline was produced.

At the southern end of the abandoned river, fossil fluvial deposits merge into a raised shoreline that surrounds the edge of Storesø. The raised shoreline here occurs at 679 m and is located 6 m above the present level of Storesø. The raised shoreline occurs at a uniform altitude around Storesø and is therefore demonstrably *not* a kame terrace (cf. Weidick 1963). At the western margin of Storesø the shoreline is succeeded westward by fossil fluvial deposits that decline in altitude towards Nordbosø. Thus during the existence of the 684 m lake, Nordbosø was continuously supplied with overflow waters from this lake. At the western end of Rundesø, raised shoreline fragments at similar altitudes are eroded in and deposited upon the proximal slopes of the stage IIb Rundesø end moraine. Hence, during the existence of the raised (679 m) Storesø lake, sufficient ice stagnation and retreat of the Rundesø glacier lobe had taken place to enable the Storesø lake to extend inside the Rundesø end moraine.

However, although stage IIb ice had retreated from the Rundesø end moraine, ice remained active east of Rundesø and at the western edge of Sydgletscher. This is demonstrated by the presence of an overflow channel at 674.2 m located immediately east of Rundesø. Hence, the existence of the 684 m overflow route demonstrates that the 674 m Rundesø overflow channel was not used until the entire Rundesø glacier lobe had melted. The significance of this inference is fundamental since it demonstrates that during the existence of the 684 m Hullet lake, ice had melted from the western margin of Sydgletscher from Alephsø to Langesø but not as far as Rundesø. Thus the 684 m overflow channel was utilised. However, the absence of the 684 m shoreline inside the stage IIb moraines on the east side of the Hullet valley shows that while the western flank of Sydgletscher was in stagnation, the eastern edge of the glacier was active.

Reconstruction of the dimensions of the 684 m lake indicate that the lake had an estimated volume of c. 950 $\times 10^6$ m^3. The abandonment of the shoreline was presumably caused by lake drainage that could only have taken place either under or over Sydgletscher and Kiagtût sermiat glacier. Since present-day drainage of Hullet takes place by drainage through a subglacial tunnel beneath both glaciers, it is suggested that the drainage of the 684 m lake took place in this manner.

From the aforementioned, it could be argued that the 684 m shoreline is a composite feature caused by the repeated emptying of the lake and its refilling on several occasions to the 684 m level. This process is unlikely to have taken place for the following reasons. The refilling of the lake to the same level requires exceptional conditions. It requires that Sydgletscher had neither advanced or retreated during the refilling phases and that its thickness remained constant. That widespread stagnation of Sydgletscher was taking place during stage IIb is indicated by the erosion of the 684 m shoreline on the inside of the stage IIb lateral moraine between Alephsø and Langesø. Furthermore, the locations of the 674 m and 684 m overflow channels indicate the widespread glacier retreat that had taken place between the 674 m and 684 m lake phases and which had probably commenced during the existence of the 684 m lake.

The shoreline of the 684 m lake varies between 5 m and 50 m in width and is generally bare of vegetation. The feature is usually eroded in unconsolidated sediments but locally is eroded in granite bedrock (Fig. 8). The shoreline platform fragments in granite are attri-

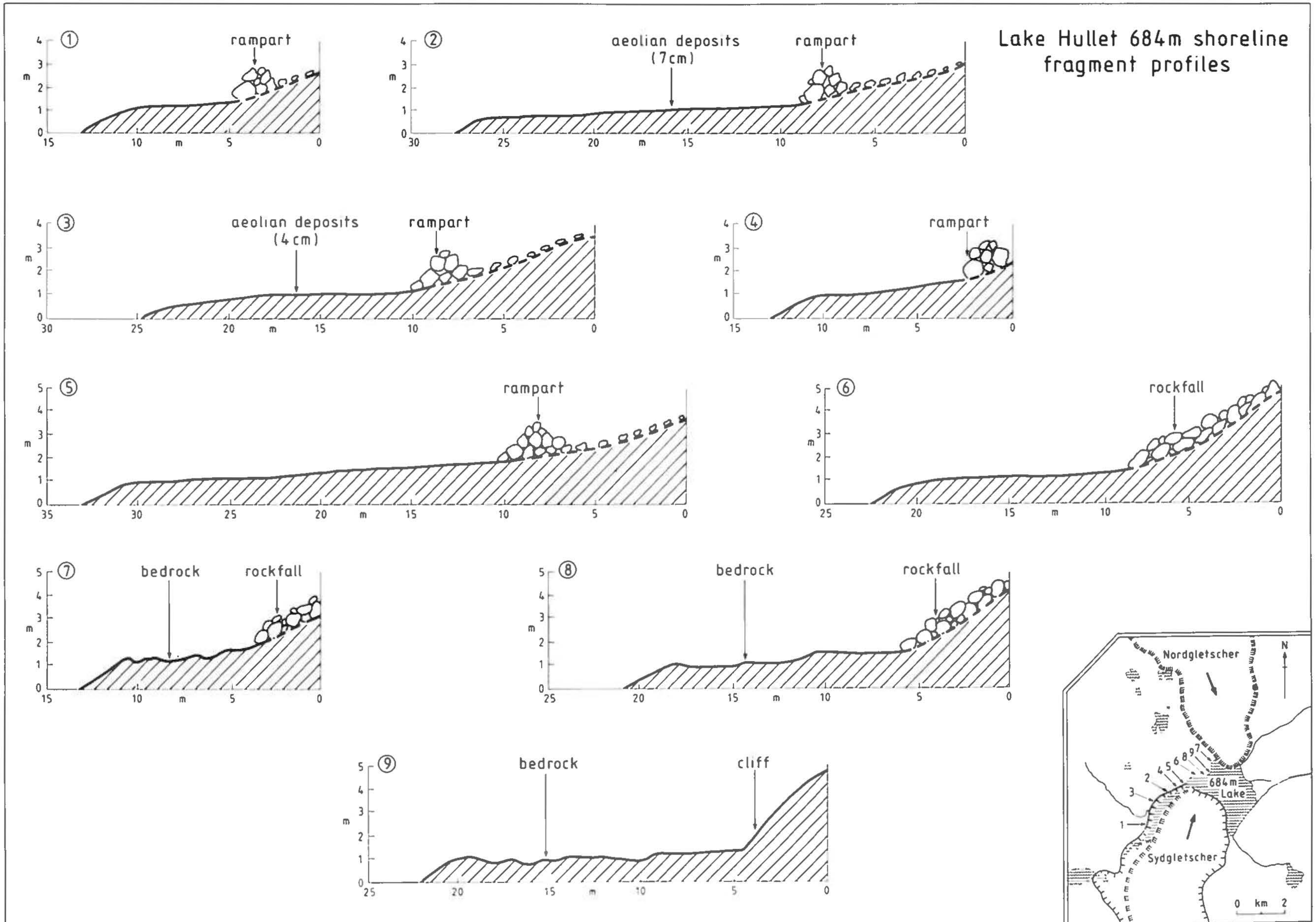

Fig. 8. Surveyed cross-profiles of 684 m shoreline fragments. The locations of the measurement sites are shown on the inset.

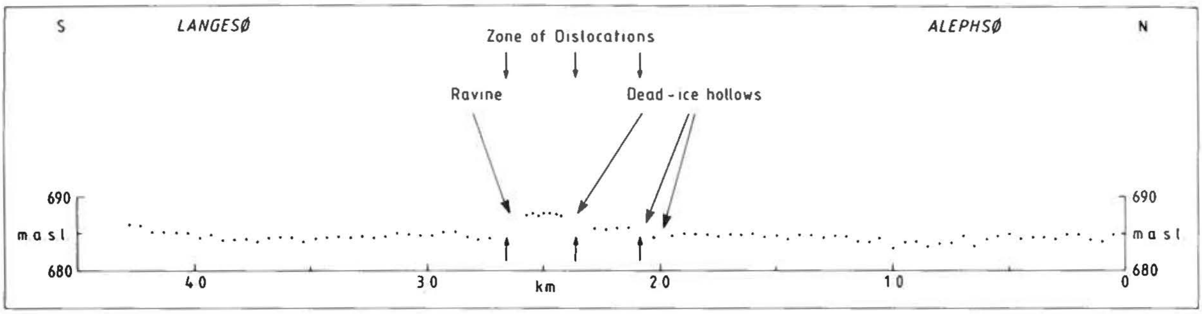

Fig. 9. Height-distance diagram of 684 m shoreline between Langesø and Alephsø showing the principal sections of dislocated shoreline. The locations of the kettle holes (dead ice hollows) and the incised alluvial channel (ravine) are also shown.

buted to frost-shattering on littoral cliffs and the removal of debris by lake ice during spring break-up and are similar to features described from other former ice-dammed lakes (e.g. Sissons 1978). The constructional role of lake ice is also demonstrated by the presence of boulder ramparts that locally define the inner edge of the shoreline (Figs 7 and 8). The ramparts most probably owe their origin to lake ice-push and ice expansion processes that may also be responsible for the extensively planated nature of the shoreline surface (Fig. 8) (cf. Worsley 1975).

The shoreline of the 684 m lake was surveyed at 50 m intervals along its length (Fig. 9). The altitudes clearly show, with two exceptions, the uniformity of altitude of the feature over a wide area. The exceptions to this trend occur c. 1 km north of Langesø where two blocks of the shoreline are dislocated by c. 1 m and 3 m above the general level of the feature. The two blocks of uplifted shoreline are respectively 200 m and 250 m in length (Figs 9 & 10). The dislocated shoreline fragments are both clear features c. 15–25 m wide and backed by lake ice-push boulder ramparts. Moreover, the uplifted blocks of shoreline exhibit only minor variations in altitude along their lengths (less than 0.3 m) (Fig. 10) and thus indicate that the process of dislocation involved the block uplift of two entire sections of shoreline 200 m and 250 m in length. The dislocated shoreline fragments are separated from each other and from the remainder of the 684 m shoreline by two kettle holes (Fig. 10, A and B) and by an incised river channel eroded in Quaternary sediments and underlain by a dolerite dyke. Hence the locations of the dislocations occur in the areas occupied by the kettle holes. However, the *precise* mapping of the dislocation zones is not possible owing to occurrence of unconsolidated surface sediments in both kettle holes.

From the above, it may be suggested that the zones of dislocation are associated with the presence of the kettle holes. However a third kettle hole (Fig. 10, C), c. 100 m wide, is also present farther north and interrupts the continuity of the shoreline in this area (Figs 9 & 10). In contrast to the displacements described above, the 684 m shoreline occurs at exactly the same altitude on both sides of this kettle hole and therefore indicates that the shoreline dislocations are unlikely to be related to kettle hole formation.

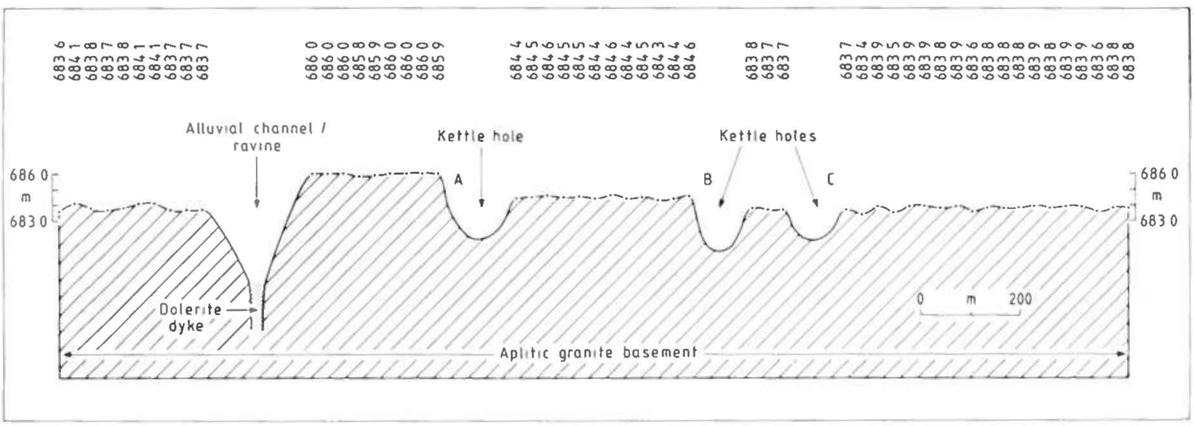

Fig. 10. Detailed height-distance diagram of area of shoreline dislocation showing individual surveyed altitudes. The widths of the kettle holes and the alluvial channel are drawn to scale. The depths of the latter features are represented diagrammatically and are not to scale.

Evidence is available of the timing of the shoreline dislocation. At several locations in the Hullet area, a poorly developed lake shoreline c. 1 m wide occurs c. 10 m below the 684 m shoreline. This shoreline is everywhere eroded in gravel and boulder accumulations that form the frontal slopes of the 684 m terrace. Owing to the narrow width of the feature and its development among large boulders, the measured inner edge altitudes of the shoreline fragments enable the reconstruction of the shoreline altitude to ± 0.5 m. The measured range of shoreline fragment altitudes is 673–675 m and includes fragments located in the dislocated area of the 684 m shoreline. The regional horizontality of the 674 m shoreline fragments therefore suggests that the dislocation of the 684 m shoreline took place *prior* to the formation of the 674 m shoreline.

Several explanations can be invoked to account for the differential block uplift of sections of the 684 m shoreline and most encounter major difficulties. For example, the view that shoreline dislocation took place through the collision of icebergs and shoreline sediments is not supported by the undisturbed nature of the uplifted shoreline blocks and their considerable dimensions. Furthermore, deformation of the shoreline fragments as a result of waterlogging of shoreline sediments is likely to induce local slumping rather than differential block uplift. Instead, it is suggested here that the differential block uplift of the shoreline fragments resulted from neotectonic activity of the earth's crust caused by the catastrophic drainage of the 684 m lake. This remarkable explanation is supported by two arguments. Firstly, block uplift associated with the drainage of the 684 m lake is in accordance with the inferred timing of the dislocations since these could only have taken place after the formation of the 684 m shoreline yet before the production of the 674 m shoreline. Secondly there are numerous published accounts of crustal faulting and earthquake activity induced by a) reservoir loading and unloading (Carder 1945, 1970, Gough & Gough 1962), b) postglacial glacio-isostatic uplift (e.g. Lundqvist & Lagerbäck 1976, Lagerlund 1977, Mörner 1981), and by c) the catastrophic drainage of ice-dammed lakes (Sissons & Cornish 1982). The latter authors describe faulted shoreline blocks and landslide deposits that resulted from the catastrophic drainage of a former ice-dammed lake in Glen Roy, Scotland. Moreover, they demonstrated that crustal faulting took place during a period of glacier retreat. Carder (1970) and Sissons & Cornish (1982) have also noted that earthquake activity is particularly common where a lake or reservoir is deeper than 100 m owing to the high strains imposed on the underlying crust (the depth of Hullet during the 684 m lake period was c. 120 m).

It is therefore suggested that drainage of the 684 m lake triggered crustal activity that resulted in neotectonic faulting and the block uplift of separate shoreline fragments. That block uplift of this magnitude was triggered by drainage of the lake from 684 m to 674 m is unlikely. For example, a lowering of the lake level by c. 10 m implies the drainage through subglacial tunnels of c. 100×10^6 m^3 of water. It is suggested instead that the differential uplift of the shoreline blocks was associated with the complete drainage of the 684 m lake and the catastrophic drainage of c. 950×10^6 m^3 of water as a jökulhlaup. Therefore the 674 m shoreline represents the level attained by the lake after it had refilled (Fig. 11). An additional factor that may have contributed to the differential uplift of the shoreline fragments is the

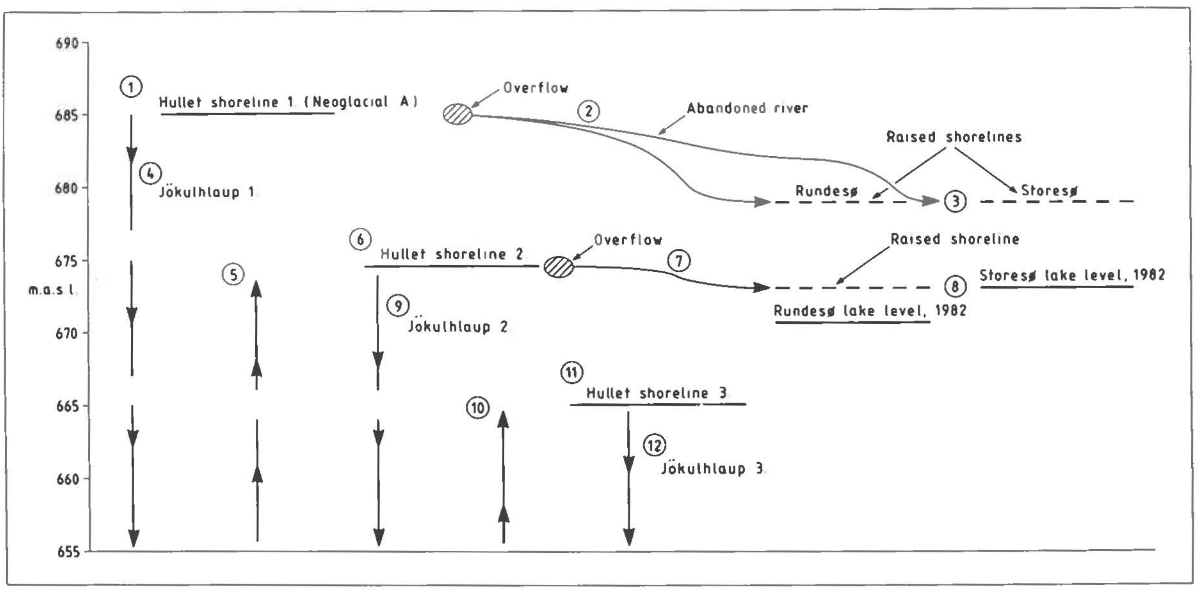

Fig. 11. Reconstructed sequence of drainage events associated with the 684 m, 674 m and later lakes.

widespread retreat of Sydgletscher that followed the formation of the 684 m shoreline. During this period the crust was also recovering from glacier unloading (cf. Sissons & Cornish 1982, p. 285) and is likely to have been subject to considerable strain. In the area of shoreline dislocations, the basement granite rock is generally mantled by glacial sediments and rockfall debris and thus the search for small-scale fault scarps has to date proved unsuccessful. This should be a major objective for future research in this area. Additional evidence that earthquake activity may have occurred in this area is suggested by the presence of a large landslide formed of aplitic granite boulders c. 1 km north of the shoreline dislocations on the slopes above Alephsø. The landslide is 150 m wide and extends c. 400 m upslope to a well-defined scarp. The foot of the landslip mantles the 684 m shoreline and thus demonstrates the occurrence of landslip activity after the formation of the 684 m shoreline. Similarly, extensive rockfall accumulations of aplitic granite debris mantle the bedrock cliffs located adjacent to the dislocated shoreline fragments and locally rest upon the stage IIb lateral moraine of Sydgletscher in this area. Since there is no evidence of modern large-scale rockfall and landslide activity in this area it is possible that these major mass movement events were also associated with earth movements following the jökulhlaup of the 684 m lake.

The coincidence in altitude of the 674 m overflow channel east of Rundesø and its associated shoreline indicates that a significant ice retreat of Sydgletscher followed the drainage of the 684 m lake. In order for the 674 m overflow channel to operate, deglaciation of the Rundesø glacier lobe must have taken place. Thus deglaciation of the western flank of Sydgletscher as far south as Rundesø is indicated. The overflow waters that entered Rundesø lake are locally represented by small terrace fragments eroded in the higher terraces (679 m) around Rundesø and Storesø. The water that entered Rundesø during this period raised the level of the level of the lake by 3.4 m and overflowed into Storesø lake and thereafter into Nordbosø lake (Fig. 11).

The 684 m and 674 m lakes are the only two fossil ice-dammed lakes in the Hullet area that possessed overflow channels. However, whereas the 684 m lake level is represented by a well-developed shoreline and the large Østgletscher lacustrine delta, the 674 m lake level is represented by only a poorly-defined shoreline. It is therefore suggested that the 684 m lake maintained its maximum level for a considerable period. In contrast, the 674 m lake was a short-lived feature. The most likely explanation for the brief existence of the latter lake is that once it attained maximum level, the subglacial tunnel beneath the adjacent glacier, eroded during the 684 m lake jökulhlaup, was more easily reopened to provide an escape route for the waters of the 674 m lake.

At altitudes below 674 m, shoreline fragments at different levels indicate the former existence of additional

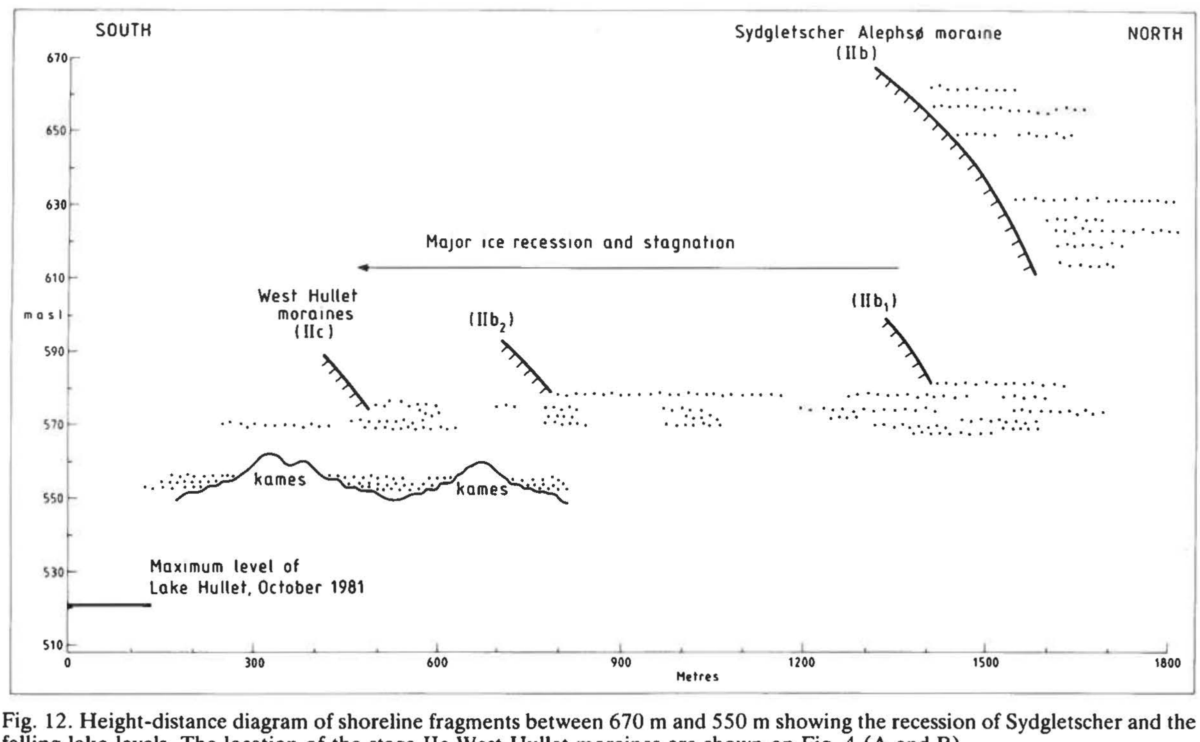

Fig. 12. Height-distance diagram of shoreline fragments between 670 m and 550 m showing the recession of Sydgletscher and the falling lake levels. The location of the stage IIc West Hullet moraines are shown on Fig. 4 (A and B).

Fig. 13. Staircase of abandoned lake shorelines north of Hullet. Each shoreline is c. 1 m in width.

ice-dammed lakes in the Hullet area (Figs 12 & 13). Many of these lakes can be related to different positions of the Sydgletscher ice barrier. The changing distribution of the lakes show clearly a progressive retreat of Sydgletscher that was associated with the formation of lakes at successively lower levels and by inference lower volumes.

Following the drainage of the 674 m lake a series of seven lower lakes were produced between Sydgletscher and Nordgletscher (Fig. 12). Each of the former lake levels is indicated by shorelines eroded on the distal slopes of the Alephsø end moraine and occur repectively at 665, 659, 652, 634.5, 626, 622 and 617 m. The distribution and altitude of these shorelines indicate that a) the formation of the Alephsø moraine had taken place prior to the production of the 665 m shoreline, b) the Hullet river area that presently breaches the Alephsø moraine remained covered by ice during the formation of all of the above shorelines, c) drainage of lake waters could only have taken place through the gap in the end moraine. Unlike the 684 m and 674 m lakes where jökulhlaup activity has been proposed, there is no evidence to indicate whether the seven shorelines were produced in association with one lake that decreased in volume or whether the shorelines indicate the maximum levels of separate lakes. The former proposal, however, requires a series of drainage events that are separated by periods of stable lake level. A consequence of this interpretation is the repeated opening and closing of the drainage outlet beneath Sydgletscher. In the areas where the shorelines are present, all measured features are regionally horizontal.

Lake drainage processes

Many authors have suggested that a prerequisite for the drainage of ice-dammed lakes is the flotation of the glacier terminus in the lake (Thorarinsson 1939, Glen 1954, Weidick 1963, Clague & Mathews 1973, Blachut & Ballantyne 1976). In this way, subglacial drainage is initiated once the water depth at the ice dam reaches nine tenths of the glacier snout thickness and the ice barrier begins to float. Under such conditions the water at the ice barrier is capable of enlarging ice cavities through the inducement of horizontal stresses (Glen 1954) and is also capable of enlarging ice tunnels by thermal erosion (Liestol 1956). However, a problem associated with the barrier flotation hypothesis is that the subglacial water outlet may close once the lake depth is less than nine tenths of the ice barrier thickness (Glen 1954). In contrast, the progressive enlargement of the subglacial tunnel by thermal erosion by escaping meltwaters may permit lake discharge to continue or possibly increase (Liestol 1956). Together, these processes have had a profound effect on the drainage of the ice-dammed lakes of the Hullet area. That the patterns of lake drainage have been highly variable is described below.

The 684 m lake was the first ice-dammed lake impounded by Sydgletscher. That the 684 m lake level was maintained for a considerable period of time is demonstrated by the well-developed shoreline and the large volume of deltaic sediments deposited in the lake by meltwaters that issued from Østgletscher. This interpretation is also supported by the fact that the formation of the 684 m lake was accompanied by a major advance of Sydgletscher to the position of the stage IIb end moraine. As a result it is likely that a considerable period of time would have elapsed before a 23 km subglacial tunnel could be produced between the snout of Sydgletscher and Narssarssuaq in order to permit lake drainage. That drainage of the 684 m lake through the subglacial tunnel was rapid is suggested by the dislocated shoreline fragments of this lake.

In contrast, it is suggested that the 674 m lake existed only for a short period of time. This lake, although possessing an overflow channel, resulted in the formation of a poorly developed shoreline c. 1 m wide. It is suggested that the narrowness of the shoreline resulted from the relatively quick re-opening of the subglacial tunnel that was produced during the preceding jökulhlaup.

The subglacial drainage of Hullet that took place between 24th September and 8th October 1981 differs markedly from those described above (see pages 4–7). During this period lake drainage took place relatively slowly. The reasons for this are unclear although the confining ice pressures around the subglacial tunnel(s) are likely to have played an important role. The most important inferences, however, are: –

a) *that former periods of Hullet drainage have been associated with both rapid (jökulhlaup) and slow discharges and*
b) *that the Hullet ice-dammed lakes have existed for both long and short periods of time before being subject to drainage.*

The lower shorelines and the recession of Sydgletscher

All shorelines below 600 m are located inside and south of the stage IIb Alephsø end moraine and were produced in association with the retreat and stagnation of Sydgletscher. Consequently the highest of these shorelines are confined to areas immediately south of the Alephsø end moraine whereas the lower shorelines extend father south. Most of the shorelines have abrupt southern limits. It is suggested that the southern limit of each shoreline coincides with former recessional positions of Sydgletscher. However only one shoreline is related to distinct lateral and terminal moraines. The sequence of shoreline formation is described below.

The highest of the above shorelines is at 583.5 m and is eroded along the inner margins of the Alephsø moraine. The feature extends c. 100 m south of the moraine and indicates that a minor recession of the Sydgletscher ice-margin accompanied the formation of this lake. Hence, during this period, the Sydgletscher end moraine was already a fossil feature despite the close proximity of the glacier snout.

Approximately 2.5 m below the 583.5 m shoreline an extremely well-developed shoreline c. 3 m wide extends c. 700 m farther south at an altitude of 581 m. Hence during the period between the existence of the 583.5 m and 581 m lakes a considerable recession of Sydgletscher took place. Similarly, the shoreline immediately below the 581 m feature which is at 577 m extends to within 400 m of the modern lake Hullet. However, rapid ice retreat and stagnation of Sydgletscher is best demonstrated by shorelines at 574.5 m, 572 m and 570.5 m. These shorelines extend from the Alephsø moraine to the slopes surrounding Hullet. However, in the intervening areas the shorelines are locally eroded in kames produced during the stagnation of Sydgletscher. Hence, the retreat of Sydgletscher from Alephsø to Hullet was associated with progressive ice stagnation and the formation of six separate shorelines. It is not possible to determine whether the shorelines represent the drainage of six separate lakes or the slow leakage of one lake. However, it is clear that following the initial retreat of Sydgletscher from the stage IIb moraine, the glacier was subject to rapid retreat and downwasting.

Numerous lake shorelines occur below the 570.5 m feature, the most conspicuous of these are at 557 m, 556 m, 525 m and 512 m. These shorelines occur adjacent to Hullet. Most of the features are interrupted by numerous crater depressions and abandoned channels produced by the decay and collapse of icebergs. The 570.5 m, 557 m and 556 m shorelines occur within the trim line zone suggested by Weidick (1963) as representative of the margin of Sydgletscher at around 1900. The trim line zone is represented on the ground surface as a transition zone between an upper area of lichen-encrusted slope debris and a lower zone that is draped by relatively recent lake sediments exhibiting only a sparse lichen cover. The trim line zone is located above and parallel to the present margin of Sydgletscher and rises gradually in altitude southwards from c. 560 m to the present firn line (c. 1600 m). Hence on the slopes that surround Hullet the coincidence in the altitudes of the shorelines and the trim line zone, in addition to the widespread occurrence of iceberg depressions, renders it impossible to correlate the shorelines with former positions of the Sydgletscher ice margin. The exception is Weidick's (1963) description of an end moraine on the floor of Hullet and the correlation of this moraine with the trim line zone. If this interpretation is correct, it would appear that the shorelines at 570.5 m, 557 m and 556 m were produced around 1900.

Lichenometry

As mentioned previously, the age of the stage IIb Narssarssuaq moraines has not been determined, although Kelly (1980) has suggested that they were produced between 1000 and 2500 years B.P. (see page 8).

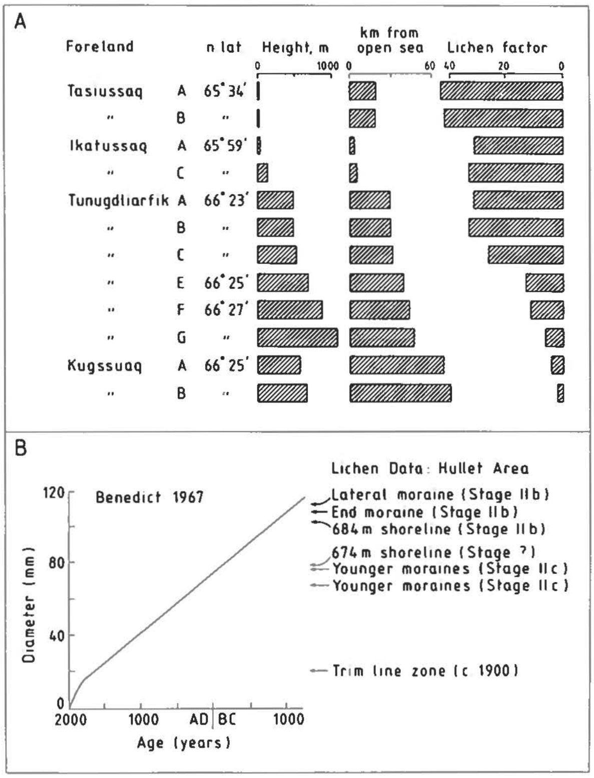

Fig. 14.
A. The lichen factor (maximum diameter of century-old thalli of *Rhizocarpon* in mm) in relation to continentality of glacier forelands in West Greenland (after Beschel 1961, fig. 1).

B. Growth curve for *Rhizocarpon geographicum* from the Colorado Front Range (after Benedict 1967, fig. 9) and the maximum lichen thalli diameters of landforms in the Hullet area. It should be noted that no correlation between the Colorado and Hullet data is implied (see text).

In order to clarify this problem, the maximum diameters of 50 *Rhizocarpon geographicum* lichens over 10 mm thallus diameter were measured from each of 105 separate locations on 7 major landforms in the Hullet area (Figs 4 & 14, Table 3). Thus, lichen diameters were measured at 15 locations of similar aspect on each landform. Comparison of the average maximum lichen diameters (Fig. 14, Table 3) permits the establishment of *relative* ages for the respective landforms and supports the relative chronology of glacial and glacio-lacustrine events already proposed. Inspection of the data suggests strongly that the stage IIb moraines and the 684 m shoreline are of the same general age and that the 674 m shoreline is a considerably younger feature. Furthermore,the trim line zone is a relatively recent feature.

The average maximum lichen diameters can be compared with those measured from W Greenland (Beschel 1961, Ten Brink 1973) and from the Colorado Front Range (Benedict 1967). However, direct correlation of lichen diameters with these areas is not possible since lichen growth rates are inversely proportional to the continentality of the area under investigation and are also sensitive to local microclimates (Beschel 1961) (Fig. 14). Although no direct correlation of the Hullet lichen data with those of Beschel, Benedict and Ten Brink is warranted, it is instructive to apply the *maximum* growth rates identified by these authors as a basis from which to derive *minimum* ages for the principal landforms in the Hullet area.

For example Benedict (1967, p. 830) suggested that *Rhizocarpon geographicum* in the Colorado Front Range grows at a rate of 14 mm/100 years during an initial 100-year "great period" before slowing to a rate of growth of 3.3 mm/100 years. Similarly Miller & Andrews (1972, p. 1135) concluded that the initial lichen growth rate on Baffin Island is 15 mm/100 years and then is 2.7 mm/100 years. In West Greenland, Ten Brink (1973, p. 329) suggested that the initial "great period" growth rate is 17 mm/100 years and is followed by a long term growth rate of 2 mm/100 years. Comparison of the Hullet lichen data with those of Beschel (1967, fig. 1, p. 1048) is particularly instructive. Beschel demonstrated that the maximum diameter of 100 year-old thalli of *Rhizocarpon geographicum* (mm) in West Greenland is related to the continentality of the area as measured by km from the open sea and by altitude (Fig. 14). The Hullet trim line zone occurs between 556 m and 571 m, is c. 100 km from the open sea and was produced c. 1900. The maximum lichen diameter of the trim line zone is 20 mm, a value considerably greater than that predicted by Beschel (Fig. 14) but similar to that of Ten Brink (1973, fig. 4, p. 326). As a result a trim line zone maximum "great period" growth rate of c. 20 mm/100 year is suggested for the first 100 years of lichen growth in the Hullet area. Thus a subsequent maximum growth rate of 4 mm/100 years (higher than the maximum growth rate of 3.3 mm/100 years meas-

Table 3. Maximum diameters (in mm) of *Rhizocarpon geographicum* measured for the principal landforms in the Hullet area. The location of the lichen measurement area is shown on Fig. 4 and includes the positions of the younger moraines A and B. The calculated *minimum* ages for the features are also shown (see text) and are based on an initial 100 year "great period" growth rate of 20 mm/100 yr and a subsequent growth rate of 4 mm/100 yr.

Landform	Maximum thalli diameters (mm)	Inferred *minimum* ages (years B.P.)
lateral moraine (stage IIb)	111	2375
terminal moraine (stage IIb)	109	2325
684 m shoreline (stage IIb)	102	2150
674 m shoreline (stage ?)	78	1550
younger moraine A (stage IIc)	76	1500
younger moraine B (stage IIc)	67	1275
trim line zone (c. 1900 A.D.)	20	c. 100

ured by Benedict (1967, p. 830)), if applicable to the Hullet area, suggests a *minimum* age for the stage IIb moraine of c. 2350 years B.P. (maximum lichen diameter of 110 mm) (Table 3). Similarly the 684 m shoreline is a slightly younger feature (c. 2150 years B.P.) than the stage IIb moraines, yet is clearly older than the 674 m shoreline. The data also suggests that the stage IIc moraines were produced prior to the Little Ice Age and by inference it would appear that Sydgletscher was not characterised by marked advance during this period.

Conclusion

During the Neoglacial, the changing position of the Sydgletscher ice-margin resulted in the formation of numerous glacier-dammed lakes in the Hullet area. The earliest ice-dammed lake had an altitude of 684 m and accompanied the formation of the Narssarssuaq moraines during stage IIb (Weidick, 1963). The 684 m lake had an area of c. 10 km^2 and a volume of c. 950×10^6 m^3 of water. During the existence of the lake, a well-defined shoreline up to 50 m wide was eroded in bedrock and in Quaternary sediments. The 684 m lake level was also associated with the formation of a large lacustrine delta that resulted from the deposition in the lake of fluvioglacial sediments that issued from Østgletscher. The 684 m lake possessed an overflow channel near Langesø. The overflow waters resulted in the formation of a 2 km long river that drained into Storesø. During the period of overflow, the level of Storesø was raised by 5.9 m. The raised Storesø lake also extended inside the stage IIb end moraine at Rundesø, while farther west the water overflowed into Nordbosø. Eventually the 684 m lake drained catastrophically through a subglacial tunnel beneath Sydgletscher and Kiagtût sermiat and thus caused the extensive deposition of sediments at Narssarssuaq. The rapid drainage of the 684 m lake and the retreat of Sydgletscher are considered responsible for the vertical dislocations of sections of the 684 m shoreline caused by straining of the earth's crust.

The drainage of the 684 m lake was followed by the refilling of the lake Hullet to 674 m. Deglaciation of the Rundesø glacier lobe and of the western margins of Sydgletscher enabled the 674 m lake to overflow into Rundesø and thereafter into Storesø and Nordbosø lakes. The 674 m shoreline is a narrow feature (c. 1 m wide). It is therefore inferred that the 674 m lake existed for only a short period of time before lake drainage occurred as a result of the reopening of the subglacial tunnel.

The drainage of the 674 m lake was followed by the formation of a series of seven lakes each of which is represented by an abandoned shoreline. These lakes occurred at 665, 660, 652, 634.5, 626, 622 and 617 m. All of these shorelines were produced after the formation of the stage IIb end moraine yet during a period when the Sydgletscher ice margin was located immediately behind the end moraine. Drainage of the 617 m lake was succeeded by an initially slow and later rapid retreat and stagnation of Sydgletscher. The initial recession of Sydgletscher to a position c. 100 m south of the end moraine was accompanied by the formation of a shoreline at 583 m. Thereafter shorelines were produced at 581, 577, 574.5, 572 and 570.5 m as Sydgletscher was subject to widespread stagnation and retreat. It is suggested that by c. 1900 the maximum levels of Hullet prior to drainage had fallen to between 570 and 555 m. During this period, the snout of Sydgletscher was located farther north than present and is believed to have produced an end moraine that is presently located on the floor of the lake.

The last drainage of Hullet prior to this work took place during October 1981. During a fourteen day period the lake level fell from its maximum level of 518–521 m to 481 m. This lowering of lake level was accompanied by the formation of c. 14 well-defined shorelines and was equivalent to the drainage of c. 60% of the original lake volume of 397×10^6 m^3. The formation of the shorelines during a period of falling lake level suggests that catastrophic lake drainage did not occur during October 1981. Together, the evidence of glacier-dammed lakes in the Hullet area suggests that a) former periods of Hullet drainage have been associated with both rapid (jökulhlaup) and slow discharges, and b) the Hullet glacier-dammed lakes have existed for both long and short periods of time before being subject to drainage.

Although it has not proved possible to determine the age of the events described above by ^{14}C dating of organic material, the lichen data provide valuable information on the relative ages of the major phases of lake drainage and glacier expansion. The suggested lichen growth rates in the Hullet are *maximum* growth rates based on published information from other arctic and alpine environments. Thus they provide *minimum* ages for the moraines and the abandonment of the 684 m and 674 m lake shorelines. As a result it is suggested that the stage IIb expansion of Sydgletscher had occurred by c. 2350 years B.P. and that the 684 m lake was drained approximately 200 years later. The 674 m shoreline and the stage IIc moraines appear to represent younger features although it is surprising that the 674 m shoreline appears to be c. 500 years younger than the 684 m feature. However, there can be little doubt that the stage IIb and IIc moraines were produced considerably earlier than the Little Ice Age and that no evidence is present to indicate a period of glacier expansion in the Hullet area during the latter period.

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1982

8. P. R. Dawes and J. W. Kerr (eds):
"Nares Strait and the drift of Greenland: a conflict in plate tectonics". 392 pp.

Exploration in the Nares Strait region in the late 19th and early 20th centuries was connected with the seaway's position as a principal route of geographic discovery. Few of the early expeditions were directed towards obtaining data for the young science of geology. At the turn of the century, with the passing of the main era of geographical discovery, including the race for the North Pole and the establishment of Greenland's insularity, geological understanding of the region advanced rapidly and geologists more or less became 'standard' members of expeditions to this part of the Arctic.

Systematic geological studies in the Nares Strait region began when Lauge Koch mapped the Greenland side of the Strait in the period 1916–23; such investigations on the Canadian side by the Geological Survey of Canada took place in the 1950s and later. Early private expeditions and later work by university and petroleum and mineral enterprises have also contributed many geological data, as have several 'military operations' centred on Thule Air Base in Greenland. Regional geological mapping in Greenland was renewed by the Geological Survey of Greenland in the 1970s and continues today.

Cooperative Danish–Canadian projects, initiated by "Operation Grant Land" in 1965–66 in northern Nares Strait, have aimed at coordinating field studies in order to better assess correlation of stratigraphy and structure across the Strait.

Greenland and Nares Strait held important positions in the early ideas about the horizontal mass movements of the continents, and both Frank B. Taylor and Alfred Wegener featured the narrow linear channel between Ellesmere Island and Greenland in their respective theories of continental drift. These two creative theorists built on the immense global knowledge assembled by Eduard Suess, who published one of the earliest appraisals of the region. Early geological maps of the Nares Strait region by Bailey Willis and Lauge Koch were used by Wegener in support of his theory of continental drift.

1982

9. C. K. Brooks and T. F. D. Nielsen:
"The Phanerozoic development of the Kangerdlugssuaq area, East Greenland". 30 pp.

This paper presents an up-to-date description of the state of knowledge on the post-Precambrian geology of the Kangerdlugssuaq area, which is a key area for the early stages of continental break-up in the North Atlantic. The area is analogous to present-day Iceland but differs from Iceland in that continental crust is present and the erosional level is deeper.

The area was affected by the Caledonian orogeny as revealed by the Batbjerg intrusion, which contains screens of Palaeozoic limestones and unique potassic rocks which relate it to the Assynt Province of Scotland.

Basin formation in the early Cretaceous heralded a period of sedimentation and volcanism which formed deposits several kilometres in thickness. The basalts are believed to have been extruded just prior to anomaly 24 (i.e. 55 – 53 m.y. ago) which reaches the coast just north of this area. The basalts are overwhelmingly tholeiites of "plume" type and include picrites. They may be derived from two different mantle sources.

Layered gabbroic intrusions which are penecontemporaneous with the basalts are widespread in the area and a number of ultramafic plugs also occur. Syenites, both under- and oversaturated, are the most voluminous rock types of the area at the present erosional level and are somewhat later. The syenites show abundant signs of contamination with the country rocks.

The Gardiner complex is the eroded core of a nephelinitic volcano and contains melilite rocks and carbonatites. Related nephelinitic lavas are found in inland areas.

In the area many dike swarms are recognized which vary from tholeiitic to strongly alkaline suites emplaced between ca. 55 and 35 m.y. ago and which give good evidence of the magmatic and chronological development of the area.

Tertiary tectonism includes three main elements: the well known coastal flexure, a major dome centred on Kangerdlugssuaq and regional plateau uplift.

1983

10. B. Fredskild:
"The Holocene vegetational development of the Godthåbsfjord area, West Greenland". 28 pp.

Holocene pollen and macrofossil diagrams from four low arctic lakes at Godthåbsfjord are presented. Each core has been divided into radiocarbon-dated palaeovegetation zones, based on the remnants of terrestric plants. The PV zones are physiognomically similar, but differences as to the composition and frequency of species can be seen between the two lakes in the interior and the two lakes from the outer coast area. The vegetation which invaded the deglaciated soil was open but rich in species, and 64 species or genera have been determined from the pioneer stage (*c.* 9400–8000 B.P.). Open soil plants were dominating, but dwarf-shrubs entered the vegetation, with species from snow-patches and snow-covered heaths dominating in the beginning. By *c.* 8000 B.P. *Salix glauca* and *S. herbacea* immigrated, and gradually the pioneer plants and chionophilous dwarf-shrubs were decimated. This *Salix-Cyperaceae* stage lasted until *c.* 6300 B.P., when *Betula nana* spread all over the area within a few centuries. A *Betula nana-Juniperus* stage lasted until *c.* 3500 B.P. In the subcontinental interior this was followed by an *Alnus crispa-Betula nana* stage, which in turn was replaced by a *Betula nana-Ericales* stage around 1800 B.P. *Alnus* has never been able to grow at the maritime outer coast, where *Betula, Cyperaceae, Empetrum* and other *Ericales* dominated after *c.* 3500 B.P. Later on, *Empetrum, Cyperaceae* and snowbed plants gradually spread at the expence of *Betula nana*.

After the deglaciation the temperature increased, reaching today's values between 8000 and 7500 B.P. At which time during the coming millennia the temperature curve peaked is not known, but it may have been fairly late, presumably during the *Betula nana-Juniperus* stage. Major climatic changes are registered in the interior at 3900–3600 and 1800 B.P., and at the outer coast at *c.* 3600 and 2500–2000 B.P.

From around 8000 B.P. the development of the lakes is fairly independent of the physical conditions of the surroundings, being dependent mainly on the trophic stages of the lakes. These pass through a succession: highly productive, eutrophic – less productive, mesotrophic – very poor, oligotrophic. As well as in the flora and fauna, these stages are reflected in the sediment, which at the beginning was a clay gyttja followed by a jelly-like gyttja and, finally, by a loose, watery gyttja consisting mainly of precipitated humus. Chemical analyses of one of the cores confirm the oligotrophication.

The pollen influx in the pioneer stage is less than 100 grains per cm² per year, increasing during the Hypsithermal to *c.* 300 in three of the lakes and *c.* 1000 in the richest one, but since then the influx decreases somewhat upwards.

A survey of the immigration or first appearance of some species palynologically important to South and West Greenland shows big time lags in the spreading of some species, e.g. *Thalictrum* and *Angelica,* whereas others, like *Empetrum* and *Juniperus,* have a more effective dispersal capacity.

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