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OBSERVATIONS ON SOME HOLOCENE GLACIER FLUCTUATIONS IN WEST GREENLAND

BY

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WITH 58 FIGURES IN THE TEXT, 5 TABLES
AND 3 PLATES

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prodekan

CONTENTS

Abstract

Historical and contemporary data record a major fluctuation of the position of the Inland Ice and local glaciation ice margins in the area. Regardless of the glacier type these frontal fluctuations are mostly in phase, with glacial readvances occurring around 1650(?), 1750(?), 1850(?), 1890 and 1920 A.D. Correlation with meteorological data suggests the operation of a delay of a few to twenty years before glacier response to climatic fluctuation. Whilst the individual readvances generally are recognisable throughout the area their magnitude shows a regional variation. Thus, near the coast and in South Greenland the readvances before 1850 produced the historical maximum extent of glaciers, whilst in the northernmost part of the area, Nûgssuaq peninsula and Umanak district, the advance of 1920 in part was responsible for the maximum extent. The historical frontal fluctuation corresponded with a fluctuation of the glaciation limits of 100–200 m.

As a whole the deposits of the historical glacier advance form a zone marking a single stage in the extent of the glaciers. Zones of Inland Ice margin deposits of a similar magnitude of prehistoric age, have been widely recognized in the area. Three zones have been distinguished; an inner zone, an outer zone and a nunatak zone. The inner zone possibly includes several stages, but the main features date from subboreal or early subatlantic times. The outer zone comprises two stages formed at 7,500–8,500 and 9,000–9,500 B.P., whilst the nunatak zone (comprising several stages) was formed before or around 10,000 B.P.

Prehistoric ice margin stages of local glaciers have been less extensively investigated. In general, they indicate only late and slight development of local glaciers due, it is believed, to the glaciation limit at the time of the retreat of the continental glaciation being already too high for their widespread development. An exception from this general trend is in the Julianehåb district where the more rapid disappearance of the continental glaciers may have favoured the better development of local glaciers.

INTRODUCTION

The aim of this paper is to discuss the nature of the Holocene fluctuations in the glaciation of West Greenland between approximately 60° and 70° N and to compare them with the historical fluctuations.

The term Holocene is used here as a collective term for Late-Glacial and Postglacial. This definition is in accordance with the proposal of GAMS (1961, p. 188) for the INQUA Congress in Poland, according to which Holocene covers the last c. 14,000 years and is separated into Late-Glacial and Postglacial with the division at c. 10,000 years B.P. at the beginning of the recession of the Scandinavian ice sheet from Salpausselkä II stage.

The name historical time corresponds to the period between c. 1600 A.D. and the present. Recent data on the glacier and climatic fluctuations of this period have been given by VEBÆK *et al.* (1962) for Greenland, and by AHLMANN (1953) and MATTHES (1942) for other parts of the world.

Because of the uniformity of the Greenland place names along the coast and the great length of the area treated, the term "district" is often used in the text. The division of the area into districts is based on the Atlas to the volumes 60 and 61 of "Meddelelser om Grønland" (see fig. 1). Exceptions are the districts of Jakobshavn, Godhavn and Ritenbenk of the Atlas which are here divided into the island of Disko, the Disko Bugt area surrounding it and the Nûgssuaq peninsula.

The place names in the text are mostly taken from the Geodetic Institute's map sheets. Exceptions to these, such as those used in the early literature, are given in inverted commas. The names of existing glacier lobes present special problems as many have been given a variety of names by visitors and investigators of different nationalities, whilst others are unnamed. All glaciers are therefore referred to by their catalogue number in the archive of the Geological Survey of Greenland (Grønlands Geologiske Undersøgelse). These numbered localities are shown in plate 1. Where two names for a locality mentioned exist, both the current authorised name and its alternatives are given in the locality index at the end (index of place names, pp. 193–202).

1.1. Former investigations in West Greenland

The earliest investigations of the glacier lobes and their deposits were made by FABRICIUS around 1770, near Frederikshåb, and by GIESECKE between 1806 and 1813 at many points along the coast of West Greenland (FABRICIUS 1788, GIESECKE 1910). Their observations, and those of many later investigators, were made at isolated localities and were generally of limited scope since they were rarely the purpose of the visit. Subsequent investigators whose observations are used in the compilation of historical glacier fluctuations are referred to in table 2 (pp. 162–187) and to a certain extent also in the regional descriptions in part 4.3 (pp. 35–60).

Among the more extensive investigations mention must be made of RINK's survey in Disko Bugt and Umanak districts in the 1850's and the work of FRODA, HAMMER, JENSEN, KORNERUP, STEENSTRUP and SYLOW between the 1870's and 1890's, in connection with the first regular topographical mapping of West Greenland. The important foreign contributions were those of A. NORDENSKIÖLD in the Egedesminde district and DRYGALSKI and BARTON in the Umanak district. The results of these investigations made in the last half of the 19th century have been summarised by AMDRUP *et al.* (1921) and BØGGILD (1928).

In this century general mapping of glacial deposits has been carried out by O. NORDENSKIÖLD (1910, 1914), KRUEGER (1928), JAHN (1938) and BELKNAP (1941), and the main morphological features of the area have been summarised by BIRKET-SMITH (1928), and NOE-NYGAARD and ROSENKRANTZ (1950). In addition, marine deposits and associated glacial features in the northern half of the area have been investigated by HARDER *et al.* (1949) and LAURSEN (1950).

Currently Quaternary geological investigations are being carried out by Ohio State University in Sukkertoppen (leader: R. GOLDTHWAIT) and by the Geological Survey of Greenland in the Frederikshåb district (M. KELLY) and the Holsteinsborg district (A. WEIDICK).

1.2. Present investigations

In 1955, 1956 and 1957, the author was engaged in a survey of the historical fluctuations of the glaciers of West Greenland. With the realisation of the essentially uniform behaviour of the glaciers over a very wide geographical area within historical time, it became desirable to determine if this same uniformity could be discerned from the deposits left by prehistoric fluctuations. Field work in the Julianehåb and Sukkertoppen districts (1958 and 1960), of Sukkertoppen to Holsteinsborg district and in Disko Bugt (1961 and 1963), and around Godthåbsfjord and in the Holsteinsborg district (1965) was therefore devoted mostly to the mapping of prehistoric marine and glacial deposits. Some of the material from the Julianehåb district has already been published (WEIDICK 1963b) and the results will only be referred to briefly for comparison with the conditions in the northern districts.

Since it has not been possible to cover all of this area of c. 100,000 km² field investigations have been concentrated in geographically suitably spaced "key sectors" where the ice margin deposits were relatively abundant and which were easily accessible from the outer coast. In all cases the surveyed sectors are west-east traverses along the fjords. The areas between have been mapped from the aerial photographs of the Geodetic Institute, Copenhagen, combined with information compiled from earlier investigations.

In the chapters to follow the localities are dealt with in order from south to north.

2. GEOLOGICAL OUTLINE OF WEST GREENLAND

Physiographically the treated area is a fjord landscape in the sense of HOLMES (1954, pp. 224–226) with great similarities to western Norway (HOLTEDAHL 1960, pp. 518–521). The Greenlandic fjords are incised into a high mountain landscape the highest parts of which forms plateaus indicating one or several erosional levels (peneplanes). The highest situated stretches (NORDENSKIÖLD: Rumpfgebiete, 1914, p. 427) reach c. 2000 m a.s.l. and are concentrated in four parts of the area: the eastern and southern parts of Julianehåb district, the southernmost part of Godthåb district, a stretch between Søndre Isortoq and Nordre Isortoq in Sukkertoppen and Holsteinsborg districts and another stretch from Disko island over the outer, western part of Nûgssuaq to the northeastern part of Umanak district. The intervening stretches are undulating Hügelland (NORDENSKIÖLD 1914, pp. 427, 429–434) which reaches altitudes of between 500 and 1500 m a.s.l.

With the exception of the highlands around Sukkertoppen and Disko–Nûgssuaq, the landscape becomes gradually lower westwards towards the Davis Strait. A strandflat, such as in Norway, has only in part developed along the West Greenland coast ranging in width between 10 and 30 km (BIRKET-SMITH 1928, p. 466).

Offshore there are extensive banks with surfaces between 50 and 100 m below the sea level. They are separated from each other by troughs up to 900 m deep (RIIS-CARSTENSEN 1948, pp. 58–59, Charts of the Danish Hydrographic Office 1100, 1200, 1300, 1400, 1500 and 1600). These troughs are continuations of some of the fjords to the margin of the continental shelf. However, most fjords end at the banks.

The bedrock of the area is Precambrian in age with Cretaceous and Tertiary formations on it west and north of Disko Bugt. Recent observations on Lower Palaeozoic deposits at Søndre Isortoq, Sukkertoppen district (POULSEN 1966, p. 26), are too restricted in extent to influence a general topographic-geological description of the area.

Fig. 1. Index map showing district boundaries and the areas covered by detailed maps as text figures (indicated by frames). Map base: map of the Danish Meteorological Institute.

2.1. Precambrian

The Precambrian of West Greenland has been divided into three main units consisting of younger fold belts on either side of a central older area. The major plutonic activity in the three areas (respectively from north to south) occurred at approximately 1700–1800 $\times 10^6$ years, 2500–2700 $\times 10^6$ years and 1500–1800 $\times 10^6$ years (LARSEN 1966). The greater part of the rocks formed during the plutonic conditions accompanying the development of these fold belts are gneisses, crystalline schists and occasional granites (BERTHELSEN and NOE-NYGAARD 1965).

The older central fold belt of West Greenland was cut by faults and intruded by several series of post-orogenic basic dykes (BERTHELSEN and BRIDGEWATER 1960, JENSEN 1962) which predate the younger fold belts to the north and south.

In southwest Greenland the plutonic rocks were overlain by sandstones and lavas and were extensively faulted and intruded by members of the Gardar alkali province in the period 1000–1300 $\times 10^6$ years (BRIDGWATER 1965).

2.2. Younger pre-Quaternary deposits

Cretaceous and Tertiary sediments and basalt occur on Disko island, the western part of Nûgssuaq peninsula and the southwestern Svartenhuk peninsula. A general description of these areas has been given by ROSENKRANTZ (1951, pp. 155–158), KOCH (1959) and BIRKELUND (1965, pp. 11–20). The sediments are of both marine and continental facies formed during the late Mesozoic and beginning of the Tertiary as a former "coastal zone adjacent to a high potential source area of erosional debris", and the source area of these sediments may be found in areas now covered by the Inland Ice (KOCH 1964, p. 537). Tectonic activity in the Mesozoic and Tertiary resulted in faulting, which may have been important in determining the present outline of the coast (WEGMANN 1939, p. 40), and in multiple uplift of peneplaned surfaces to form plateaus up to an altitude of c. 2000 m a.s.l. During the subsequent periods of glaciation these plateaus have been important areas of firn accumulation whilst after the deglaciation they have acted as climatological and topographical barriers to the extension of the Inland Ice (cf. WEGMANN 1939, p. 42, CAILLEUX 1952, FRISTRUP 1966).

In the Umanak district the peneplain surfaces cut both the Precambrian gneiss and Tertiary basalt and peneplanation can be dated presumably to the late Tertiary. This is also the age of peneplains in South Greenland where they cut dykes of Cretaceous or Tertiary age. The age of these young dykes is given by BERTHELSEN (1961, fig. 2) and LARSEN (1966, p. 60).

2.3. Quaternary

Nothing is known about the early history of the Inland Ice. Indications of gradually falling temperature are provided by the late Mesozoic and Tertiary floras and faunas of North America (CHARLESWORTH 1957, p. 698, fig. 123, DORF 1955, fig. 3) and Europe (WOLDSTEDT 1954, p. 9, fig. 1, LLIBOUTRY 1965, p. 897, fig. 21. 14) suggest that it was first formed at the end of Tertiary.

The occurrence of glacial striae and erratic boulders in all parts of the country indicates a virtually complete cover by ice during at least one glaciation. There is evidence that the most recent major glaciation also formed a complete cover. This is based, for example, on the evidence of the recent isostatic uplift of all the present area of land shown by the widespread occurrence of young, raised marine deposits and strand lines. It is also suggested by the continued existence of south-north trending, large negative gravity anomalies over the area, with the exception of the outermost skerries (NØRGAARD 1948, KEJLSØE 1958, SAXOV 1958a, SVEJGAARD 1959). Another line of evidence is the presence in the outer coastal areas of fresh glacial striae and polished rock surfaces below marine deposits younger than the last glaciation.

The extent of the Inland Ice during interglacial times is unknown. However, the occurrence of derived concretions of sediments of presumed interglacial age at the ice margin at Frederikshåbs Isblink and Ujaragssuit påvat, Godthåb district (BRYAN 1954) shows that the Inland Ice in at least one interglacial period had an extent less than at present at these localities.

2.3.1. Glacial deposits

Deposits are commonly relatively thin, though in places they can form ridges 50–100 m high.

A large proportion of the existing Quaternary sediments have been through several glacial depositional and erosional cycles and consequently the commonest glacial deposits are moraine gravels with numerous rounded, and often large boulders. Boulder fields are often found on the high plateaus and in the northern or alpine parts of West Greenland.

2.3.2. Holocene uplift of West Greenland

Raised marine deposits are common over much of the area, especially in the northern and eastern parts. The oldest sandy-silty members, the *Portlandia* clay of HARDER *et al.* (1949) and LAURSEN (1950) which are often over 50 m thick extend around and to the south of Disko Bugt. However further south in the westernmost central parts of the area, shell banks with or without sandy deposits are the dominant facies, though according to their fauna, which includes *Pecten islandicus* and *Mytilus edulis*, they are younger than the *Portlandia* clay.

Laursen (1950) has put forward a succession for the marine faunas in these deposits in the area of West Greenland between Sukkertoppen and Umanak districts. The dating of his zones was based exclusively on long distance correlation with the faunal sequences in the marine deposits of Norway and Iceland. From this chronology, he has suggested dates for some of the former sea levels, but because of the uncertainties inherent in this approach a check of these dates by radiometric methods would be useful.

The uplift of a single area has been investigated by Iversen (1953) who determined by pollen analysis the date of the isolation from the sea of three lakes in the interior of Godthåbsfjord, situated at 8, 50 and 100 m a.s.l. respectively. A radiometric date of 2390 ± 120 B.C. for the 8 m level of this locality has recently been published by Fredskild (1967, p. 50).

Recent archeological investigations in Disko Bugt show that the sea level can not have been more than 2 m above the present one since the time of the Sarqaq culture, between 3570 ± 150 B.P. (K-144, Tauber 1960a) and 2740 ± 100 B.P. (K-516, Tauber 1960b).

Important information on the isostatic uplift has been obtained from other parts of Greenland. From East Greenland numerous radiocarbon dates have been obtained for marine deposits at Mesters Vig by Stuiver and Washburn (1962) and Lasca (1968). A smaller number have been obtained from samples collected by Davies and Knuth (Knuth personal communication) in Peary Land, North Greenland. The latter indicates that the uplift of the land in Peary Land was somewhat delayed compared with the Mesters Vig area though the uplift curves have nearly the same trend. Davies (1961, p. 101) also stated that there has been differential uplift in parts of northeast Greenland, where the former beaches incline towards the outer coast at 5–1.5 cm/km.

Data on the uplift of West Greenland have been obtained from the radiocarbon age determinations of H. Tauber on material collected by M. Kelly and the author. These dates are listed in table 5 (p. 191) and are compiled graphically in fig. 2, together with other published dates from this area.

The altitudes of the samples collected by the author in West Greenland were determined by repeated measurements by hand level. Even working on slopes up to c. 100 m a.s.l. this method has proved to give more exact values than altimeter readings, where additional triangulation with a theodolite has provided a control.

The trend of isostatic uplift of Greenland as a whole is of a type well known from other former ice loaded regions. This is shown in a

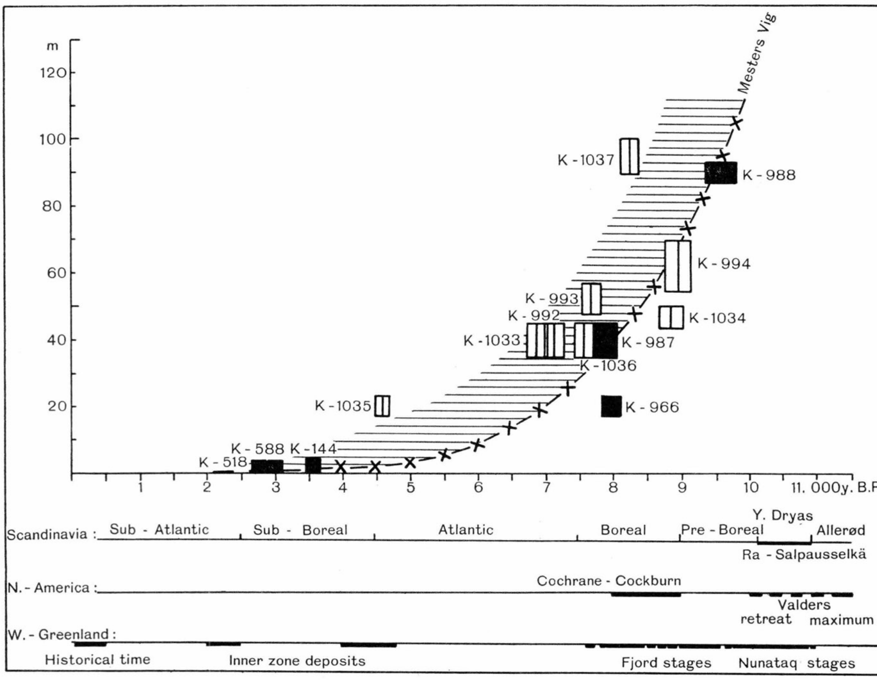


Fig. 2. Uplift of middle West Greenland. The data plotted are not corrected for the eustatic change of sea level. The vertically divided rectangles represent data collected by the author whilst the black rectangles are data derived from other sources. Horizontal scales under the abscissae give the chronostratigraphic divisions for Scandinavia, after LUNDQUIST (1965, p. 156), and for North America, after FALCONER *et al.* (1965b, p. 150), and WAYNE & ZUMBERGE (1965, p. 76). The thick parts of the line represent periods of glacial advances.

generalised manner by the hatched zone of fig. 2 within which lie the curves from the various parts of Greenland with the curve for North Greenland near to the left side of the zone and the curve for Mesters Vig near to the right. However, without the correction for an apparent age of 550 years for modern shells applied by STUIVER and WASHBURN to their dates, the Mesters Vig curve would be to the right of the zone.

In spite of the wide geographical spread of the data collected from West Greenland, they are nearly all close to the hatched area of fig. 2. Assuming that the uplift curves are an expression of the relaxation of the earth's crust after the removal of an ice load, curves towards the left of the zone should represent the uplift of the most recently deglaciated areas, *i.e.* the northernmost and interior parts of the coastal stretch. Conversely curves to the right should be from early deglaciated areas, such as the outer coast (*e.g.* Godthåb K-966) and the Julianehåb district.

There is a wide variation amongst the values quoted in the literature for the upper marine limit between the Godthåb district and Disko

Bugt (BØGGILD 1928, LAURSEN 1950). However, several of these have proved on closer inspection to have been based on the misinterpretation of ice marginal terraces, *e.g.* one at 270 m a.s.l. in the inner part of Søndre Strømfjord and one at 175–185 m a.s.l. in Disko Bugt (LAURSEN 1950). The remaining outstandingly high marine level reported from the area is that at 175 m a.s.l. at Gieseckes Sø (LAURSEN 1950, p. 126). Its description suggests that it may also be based on the evidence of terraces of dubious origin.

The only two criteria found to be reliable evidence of former sea levels in the area are undisturbed shell-bearing beds of reasonably great extent or well marked series of beach ridges. If these criteria only are accepted, it seems that the highest well developed strandlines lie some tens of metres above 100 m a.s.l. in West Greenland but at slightly lower altitudes in Julianehåb district in the south.

2.3.3. Holocene climatic development

From the faunal evidence, LAURSEN (1950) postulated the existence of two cooler phases separated by a warmer during the deposition of the *Portlandia* clay. His correlation of this with the European Older *Dryas*-Allerød-Younger *Dryas* should be considered perhaps as tentative until verified by radiocarbon age determination. Another possible piece of information about the early climate is from IVERSEN's demonstration of the early existence of *Atriplex* pollen in lake sediments in inner Godthåbsfjord, which he suggests indicates that temperatures were similar to todays at a time thought to be early Boreal (IVERSEN 1953).

Little is known about the subsequent period except that a warm climatic optimum was followed by cooler conditions during the last 1500 years (LARSEN and MELDGAARD 1958). With reference to humidity, both IVERSEN (1934), for Godthåbsfjord, and LARSEN & MELDGAARD, for Disko Bugt, suggested that the climate has become more continental during the last 600–750 years. This however appears to be at variance with the information about the accumulation on the northern part of the Inland Ice given by LANGWAY (1962, p. 116). From his preliminary investigations of a 411 m long core from Site II, c. 360 km east of Thule, and 2200 m a.s.l., it seems that the accumulation there has been constant, at a little over 30 cm water annually, from 907 A.D. to 1878/1880 A.D. After 1880 precipitation appears to have increased, reaching a maximum of c. 50 cm water annually around 1920 since when it has been decreasing to 35–40 cm around 1950. LANGWAY however, could not exclude the possibility of errors due to the squeezing out of the lower firn layers.

The instrumentally recorded climatic fluctuations of the last century are treated in detail by LYSGAARD (1949). He stated that at Jakobshavn,

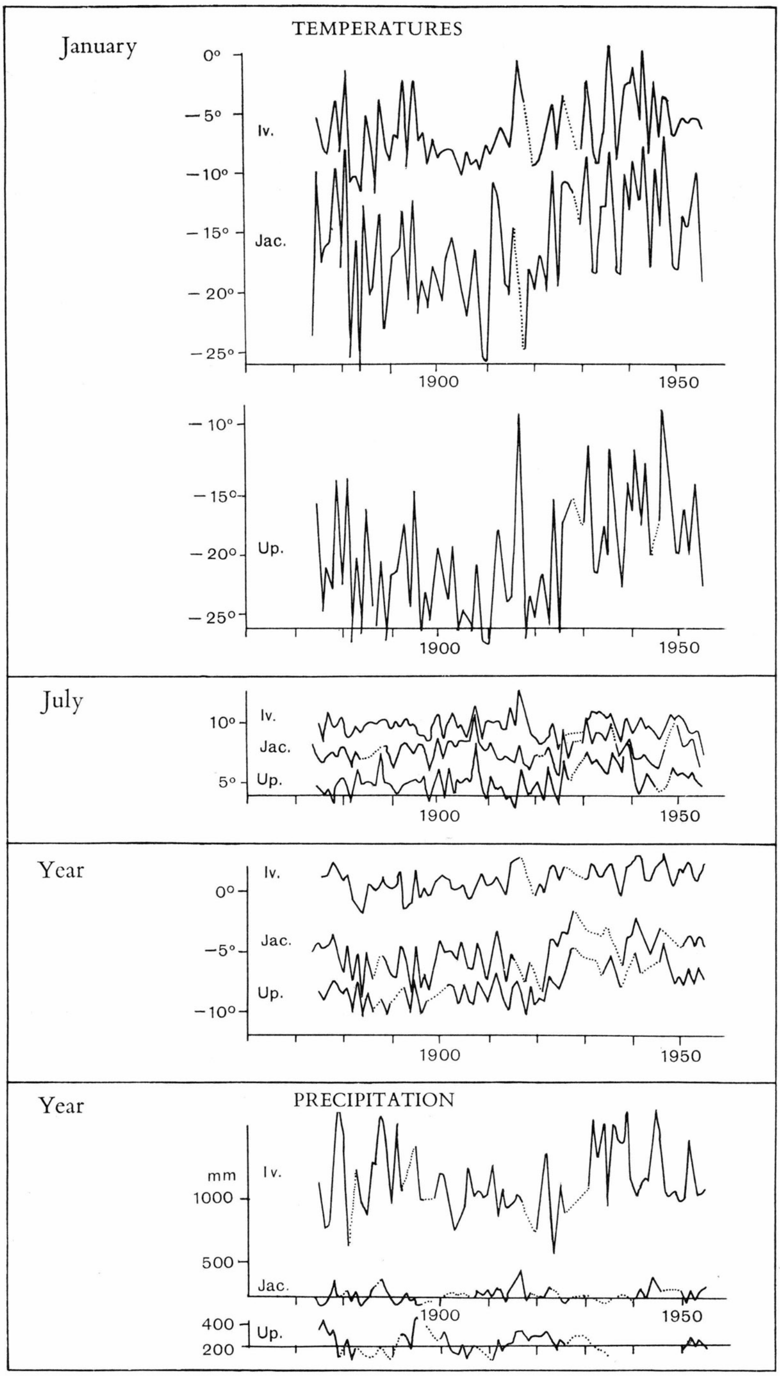


Fig. 3. Mean January, July and annual temperatures and annual precipitation at Ivigtut (Iv.), Jakobshavn (Jac.) and Upernavik (Up.). Data compiled from the Annual Reports of the Meteorological Institute, Copenhagen.

the mean January temperature rose 4.4° between the two periods 1872/1911 and 1911/1940, the July mean rose 0.4° between 1865/1904 and 1911/1940 whilst the annual mean rose 1.9° between 1872/1911 and 1911/1940.

Other studies of the recent temperature changes have been made by PUTNINS (1962) and VIBE (1967). On the basis of observations compiled from 13 stations PUTNINS stated that "drop or rise of temperature is out of phase at stations not very far apart", though for the years investigated, the January, July and annual means "generally show a similar course". VIBE (1967) has investigated the temperature fluctuations at Godthåb recorded during approximately 150 years of observations. According to him cold periods occurred in 1807/21, 1860/1866, 1880/1890 and to a lesser extent between 1913 and 1916.

Measurements compiled from the Yearbooks of the Meteorological Institute for the stations Ivigtut, Jakobshavn and Upernavik are shown in Fig. 3. Though the minor fluctuations are out of phase, the curves have similar trends. The smallest fluctuations occur in the summer temperatures and the greatest ones in the winter temperatures, which means that though the summer temperature has not risen much throughout the last century, the length of the summer, and therefore of the glaciers ablation season, has increased in the same period. From these curves and from VIBE's curve for Godthåb it may be concluded that the period 1807 to 1821 was cold throughout the whole area.

The climatic fluctuations in West Greenland have also been dealt with by BENSON (1962, pp. 58–59). Besides giving the general increase in the annual mean temperatures throughout this century he has plotted the mean annual temperature against latitudes along the west coast of Greenland (station values for Ivigtut, Godthåb, Jakobshavn, Upernavik, Thule and Alert) and he demonstrates that, in the stretch from Ivigtut to Upernavik, the annual mean temperature decreases northwards by 0.8° per 1° of latitude.

Fluctuations in precipitation in coastal North Greenland and on the Inland Ice seem to a certain extent to be in phase. The studies of DIAMOND (1956, p. 2), GERDEL (1961, p. 94) and LANGWAY (1962, p. 116) indicate an accumulation or precipitation maximum around 1920 at the Inland Ice stations of Eismitte and Site II (Eismitte situated on 71°11′ N, 39°56′ W, at an elevation of about 3000 m), and Upernavik. However, at Ivigtut, this period is one of minimum precipitation. Up to now, no clear trends in the fluctuations of precipitation can be seen, nor does the material allow any zonal shift of precipitation maxima or minima to be traced.

3. PRESENT GLACIATION OF WEST GREENLAND

The numerous existing glaciers and the glaciation of the eastern margin of the area by the Inland Ice are the natural starting point for the investigation of the past glacier fluctuations in West Greenland. In connection with this it is of importance to determine the climatic and topographic factors which limit the extent of the present glaciation and to deduce how far these can be inferred from the deposits of the present glaciation and hence how far they can be used as a key to the interpretation of the older ice margin deposits.

A general expression of the conditions necessary for glacier formation is provided by the snow or glaciation limits. Sporadic mention of these limits in West Greenland has been made since an early date (Crantz 1770, I, § 11, p. 41; Rink 1857, I, pp. 68–71), but they have never been thoroughly mapped, nor has their relationship to the mass balance of the glaciers been studied. Only to a very slight extent has the author had any possibility of making measurements relating to the niveometric balance or to the mass balance of glaciers and the following discussion incorporates both earlier observations in the area and results from the Alps, Scandinavia and North America.

3.1. Snow and glaciation limits

Many different definitions of these "limits" exist but in the following the terminology mainly follows that of Ahlmann (1948, pp. 41–42) or Meier (1962, pp. 256 & 259).

The temporary snow line is the lowermost limit of snow cover at a given time of the year on unglaciated areas.
The transient snow line is the corresponding line on glaciers.
The climatic snow line is the uppermost limit of the temporary snow line.
The firn line is the uppermost limit of the transient snow line.
The equilibrium line is the line separating areas of accumulation and ablation on a glacier.
The glaciation limit is the lowermost level for the formation of glaciers, as determined by the "summit method".

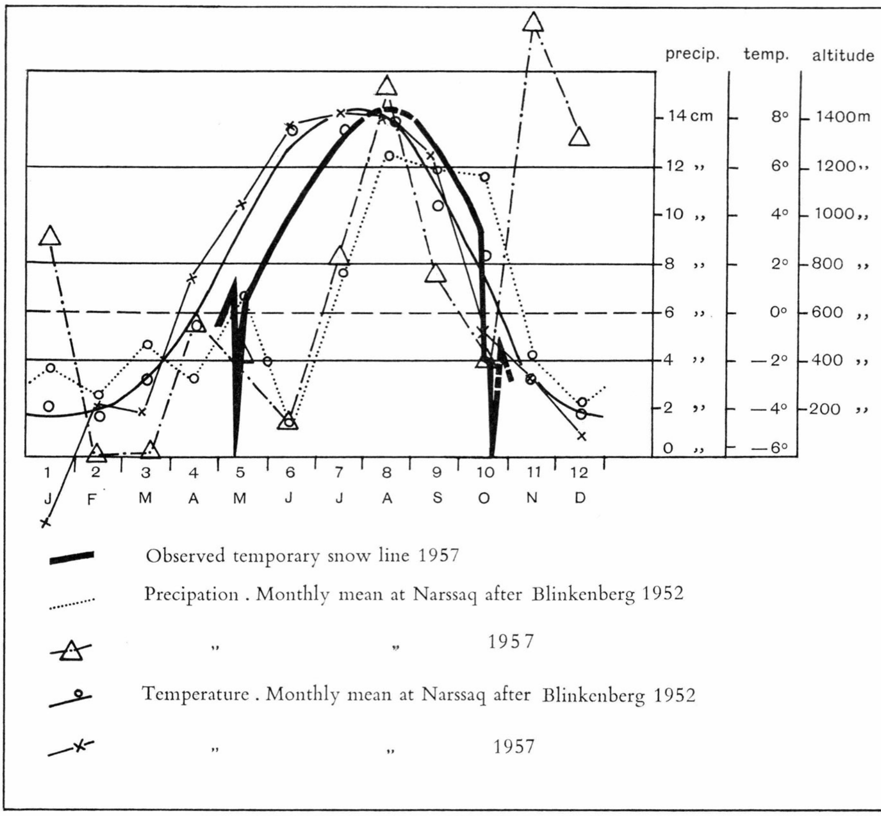


Fig. 4. Altitudinal variation of the temporary snow line in the Ilímaussaq area, Julianehåb district in the summer of 1957.

Also given are: 1) Monthly means of precipitation and temperature for 1957 and 2) Monthly means of precipitation and temperature for 1944–1948 (BLINKENBERG 1952). The meteorological observations were made at "Narssaq point" (Nûgarssuk) and provided by the Meteorological Institute, Copenhagen.

An example of the course that the temporary snow line can take is shown in fig. 4. The movement of the line has been followed from May 5th to October 20th, 1957, on the slopes surrounding the small glacier on Ilímaussaq mountain near Narssaq (loc. 7 in plate 2). The sharp notches on the curve in May and October correspond to major snow falls. It is seen from the climatic data included in the figure that in 1957 the temperature and precipitation did not differ essentially from the means of 1944–1948. Furthermore, it was clear that:

Fig. 5. Altitudinal variation of transient snow line, northwestern part of Søndre Isortoq, Sukkertoppen district, compiled from aerial photographs. Numbers on the maps are the observed altitudes of the transient snow line in hectometres. 5a. Geodetic Institute's route 507 D–N (17.6.1948). 5b. Geodetic Institute's route 505 D–N II (17.7.1948), 507 C–N and 507 C–S (18.7.1948). 5c. Geodetic Institute's route 505

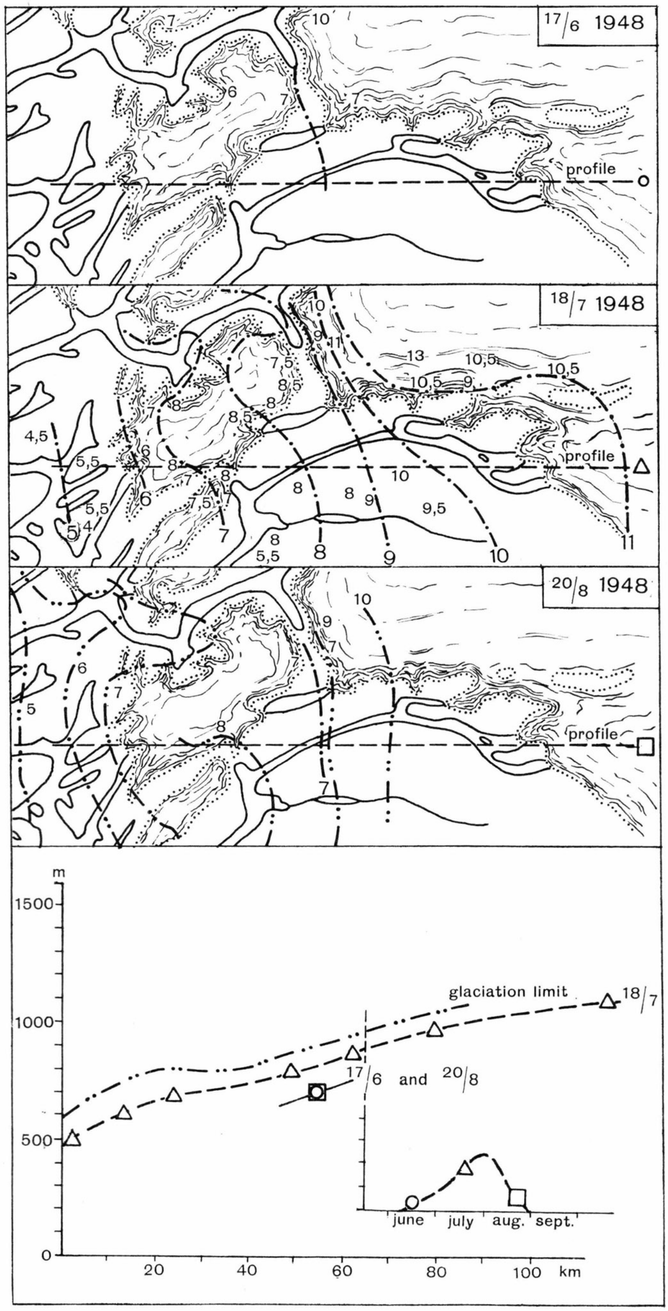


Fig. 5a

Fig. 5b

Fig. 5c

Fig. 5d

Fig. 5
(continued)

D–Ø (20.8.1948); –·–·– transient snow line 20.8.1948 and –··–··– glaciation limit. 5d. Profile of the transient snow line across the investigated area compared with the glaciation limit, determined by the summit method. The line of the profile is shown in figs. a, b, and c. The observations are projected into the profile line from the north and south without any allowance for the Massenerhebungseffect (Klebelsberg 1948, p. 30) over the ice caps around Evighedsfjord.

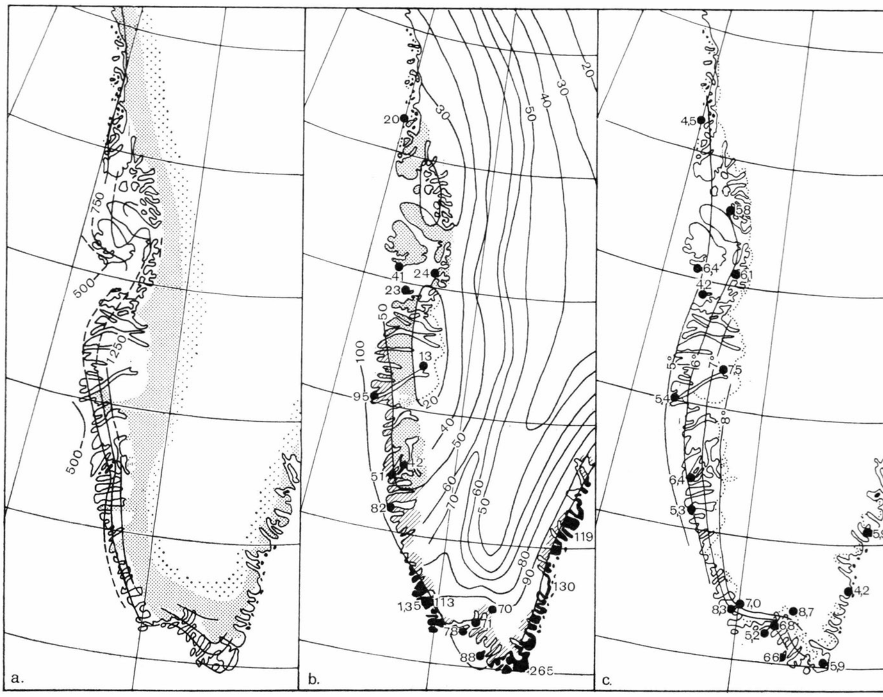


Fig. 6a. The glaciation limits in West Greenland. On the Inland Ice the firn line is the limit between the densely dotted (the ablation facies) and the thinly dotted (the soaked facies) zones. These zones are after Benson (1961, fig. 5, p. 27). 6b. Precipitation after Blinkenberg (1952) and accumulation on the Inland Ice after Benson (1961, fig. 3, p. 22). 6c. Mean temperature of the summer months. June–September, after Blinkenberg (1952).

1) the transient snow line on the glacier ("Narssaq glacier") was situated 100–200 m below the temporary snow line and 2) that both the climatic snow line and the firn line lie above the highest parts of Ilímaussaq and "Narssaq glacier" respectively.

In both Norway (Liestøl and Østrem 1962, p. 324) and the Alps (Klebelsberg 1948, I, p. 31) the firn line is stated to be about 100 meters lower than the climatic snow line.

Information about the variation of the transient snow line can also be obtained where there exist a suitable series of aerial photographs, for example for the Søndre Isortoq area in the southernmost part of the Sukkertoppen district. In spite of the more limited nature of the observations the same conclusions can be drawn as in the Julianehåb district. The evidence from Søndre Isortoq-Evighedsfjord area is illustrated in fig. 5. No climatological data are available for this area, the nearest meteorological stations being situated about 150 km away (Qôrnoq, Godthåb and Søndre Strømfjord).

The diagrams also show the similarity of the trends of the glaciation limit and the climatic snow line across the area. Because of the large error involved in their determination they can be considered as almost identical.

The height of the glaciation limit in West Greenland is shown in fig. 6. The values given are derived from the 1:250,000 map sheets of the Geodetic Institute, which are based on aerial photographs from the 1940's and 1950's with the exception of the Godhavn sheet, which is based on terrestrial measurements in the 1930's. Study of the aerial photographs and field work has shown that snow patches have occasionally been drawn as glaciers on the maps and minor glaciers have been left off. However, these few exceptions hardly effect the determination of the glaciation limit. However, if the many ice caps on high plateaus in West Greenland which recent aerial photographs reveal to be completely without perennial snow or firn, are used in the determination of the glaciation limit its value will tend to be too low, and less than the height of the climatic snow line. In contrast, the presence of steep, snow-bare peaks above the climatic snow line will result in the determined value of the glaciation limit being too high, a possibility mentioned by SVENSSON (1959, p. 11).

In spite of all the errors involved in the determination of the different limits, the limits can be considered as all lying in a zone 100–200 m wide, the altitude of which increases from the coast towards the Inland Ice. The eastwards extrapolation of the glaciation limit strikes the surface of the Inland Ice at altitudes between 1400 and 1700 m a.s.l. (see fig. 42), which is in general agreement with the heights of the firn line determined by LOEWE (1936, pp. 323–324), BAUER (1955b, p. 460) and BENSON (1961, p. 27).

No measurement of the altitude of the equilibrium line on a local glacier has been made in West Greenland. Whilst this line generally is lower than the firn line because of the refreezing of meltwater, MEIER (1962, p. 256) believes that on temperate glaciers they are nearly identical, though on a polar glacier there may be an important segment of the glacier between them. However, even in an area as far north as the ice margin at Thule, NOBLES (1960, p. 3, fig. 3 and p. 24, fig. 18) puts the height interval between the two zones at only 60 metres and SCHYTT (1955, pp. 47, 52–57) seems to put the value at a nearby locality at c. 100 m. Since most of the glaciers in West Greenland are described as being partly or fully temperate anyway (SUGDEN and MOTT 1937, ETIENNE 1940, FRISTRUP 1961 and BULL 1963), the equilibrium line and the firn line can only be separated by detailed studies.

3.2. Mass balance of the glaciers

From the text above, it is clear that the exact determination of the equilibrium line requires more detailed information about accumulation and ablation than is now available. However, even a rough estimate of the accumulation and the ablation give an impression of the activity index (MEIER 1962, p. 259, also the glaciation energy of SHUMSKII 1950, p. 6). This is defined as the vertical gradient of ablation plus the vertical gradient of accumulation at the equilibrium line of the glacier (see fig. 7).

Only a few ablation measurements have been published from the local glaciers and the Inland Ice margin in West Greenland. In the Julianehåb district, four ablation measurements on the margin of the Inland Ice from c. 600 m a.s.l. have been published by WEIDICK (1963b, pp. 40–41). Though these measurements extended over only one month, they indicate that the annual ablation here may be of the order of 1.3–2.2 m water. At Kangerdluarssuk, c. 60 km further west, ablation measurements of 6 stakes made by the author in the period 31st of May to 13th of June 1957, gave the following results:

4.5 cm water/24 h at 130 m a.s.l. 5.5 cm water/24 h at 170 m a.s.l.	Equals total annual ablation of 7–9 m(?) water at 100–200 m a.s.l
3.6 cm water/24 h at 300 m a.s.l. 3.0 cm water/24 h at 320 m a.s.l. 3.9 cm water/24 h at 340 m a.s.l. 4.1 cm water/24 h at 350 m a.s.l.	Equals total annual ablation of 6–8.5 m(?) water at the altitudinal interval 300–350 m a.s.l.

From the Holsteinsborg district, BRECHER and KRYGER (1963, p. 13) give the annual ablation on the eastern part of Sukkertoppen Iskappe as nearly $1^1/_2$ m water per year at c. 1200 m a.s.l.

Further north, ablation measurements made at low altitudes on Upernivik Ø, Umanak district (KUHLMANN 1959), on local glaciers on the north side of Nûgssuaq peninsula and on the margin of the Inland Ice in Umanak district (LOEWE and WEGENER 1933, LOEWE 1934) shows that ablation in the area is fairly uniform. Several measurements of the ablation on the Inland Ice margin have been made also in Disko Bugt (DE QUERVAIN and MERCANTON 1925, p. 243; LOEWE 1934, p. 363; BAUER *in* AMBACH 1963, p. 195, fig. 76). LOEWE has also pointed out that the decrease of ablation with altitude on the Umanak district and in Disko Bugt are identical. LOEWE's and BAUER's data are plotted in fig. 7.

Accumulation measurements are relatively numerous for the Inland Ice and a recent compilation of them is given by BENSON (1961, fig. 3), here reproduced in part in fig. 6. However, for the local glaciations there are only those made on Sukkertoppen Iskappe (ETIENNE

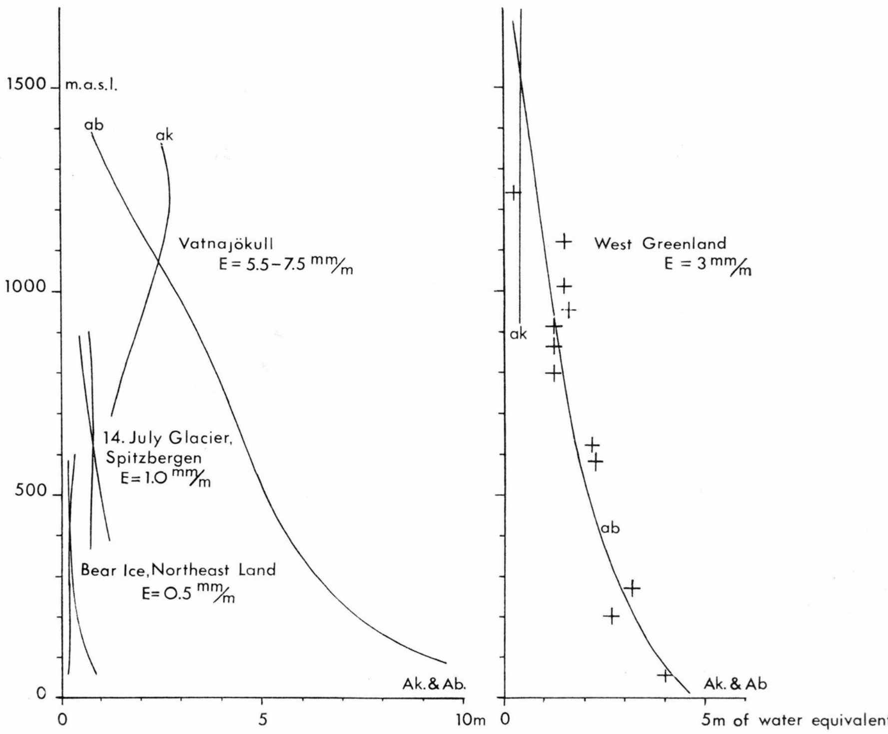


Fig. 7. Ablation (ab) and accumulation (ak) in Disko Bugt and Umanak district, West Greenland, as a function of altitude. Curves from Spitzbergen and Iceland (after SHUMSKII 1950, fig. 3, p. 6) shown for comparison. Abscissae; accumulation and ablation in water equivalent (Ak & Ab). Ordinates; metres above sea level. E; activity index.

1940, HOLLAND 1961, BULL 1963). Hence, in fig. 6, precipitation figures derived from the coastal stations have to be used. These however, can be taken as a guide to the accumulation. The accumulation measured on Sukkertoppen Iskappe, at least down to 200 m above the firn line, agreed well with the values calculated from the coastal precipitation.

The data on ablation and accumulation mentioned above, derived from the Inland Ice margin in the Disko Bugt–Umanak area is compiled in fig. 7. This curve, however, can only be regarded as a first approximation to the "net budget curves", given by MEIER (1962, p. 258) for South Cascade Glacier, Washington State, and by LIESTØL and ØSTREM (1962, p. 328) for Norwegian glaciers.

The curve in fig. 7 indicate an activity index of c. 3 mm/m, for Disko Bugt and Umanak district, a magnitude which is characteristic also of the glaciers in Spitzbergen. Julianehåb district possibly has a greater activity index than the Disko Bugt–Umanak district, and the lower portions of the glacier lobes there may have, in places, as high an ablation as the Icelandic glaciers. One result to be expected of a higher activity index in the Julianehåb district would be a faster response of glaciers there than further north to fluctuations in climate.

The maps in fig. 6 imply that the height of the glaciation limit is dependent on both summer temperature and annual precipitation. The observed related values are listed in table 1a, p. 161 and their relationships were kindly treated by regression analysis first by K. PODER, and later (see table 1b) by E. SVEINSDOTTIR who also determined the correlation coefficients. Based on the data of table 1b from West Greenland, the simplest relationship between glaciation limit and climatic conditions seems here to be near to:

$$-99 + 176\,s - 0.9\,p \backsimeq g$$

where s is the summer temperature at sea level in degrees centigrade, p the annual precipitation in cm and g the appropriate altitude of the glaciation limit in metres.

The simplified equation above and the simple correlation coefficents of table 1b ((s, g) being o.764, (p, g) −0.072) show that the summer temperature s is the principle factor for the position of the glaciation limit (g). As a check SVEINSDOTTIR calculated correlation coefficents of other combinations of the climatic data, *e.g.* mean winter temperature versus winter precipitation/summer precipitation/annual precipitation, but none of these combinations gave as satisfactory results as those, listed in table 1b.

No account has been taken of the following factors; possible anomalies in the local climate of the stations, variations in the length of time weather observations have been made and the variations of the thermal lapse rate. With reference to the last factor, LOEWE (1964, p. 7) has demonstrated in the Søndre Strømfjord area that "big lapse rates are typical for mountain regions at a time when the temperature at the upper station is kept low because part of the incoming heat is used for the melting of ice and snow". This means that a straightforward correlation between observed temperatures at sea level and the glaciation limit does not exist. A determination of the relative influence of the two factors on the glaciation limit from the relationship given here must be considered as a first approximation only.

3.3. Fluctuations of the glaciation limit

Aerial photographs taken in the month of August in 1948 and 1949 show several minor ice caps to be without firn. From the form of the ice caps at the time of their maximum extent the firn line in historical time can be estimated to have been 100–200 m lower than at present. Information about the past position of the firn line can be deduced some-times from the literature. For example STEENSTRUP (1901, pp. 290–291) states that a small firn field, which has since disappeared, still existed on Qeqertaq island, Disko Fjord, in 1898. As the island is a basalt plateau c. 600 m a.s.l. and the present glaciation limit is interpolated from nearby localities to be c. 700 m a.s.l. it must have been c. 100 m lower in the nineteenth century. A similar result may be given by Ubekendt Ejland, Umanak district (STEENSTRUP 1883b, p. 226). It is also clear that the climatic snow line in Upernavik district cannot have been more than 100–200 m lower than now, otherwise the island of Akuliaruseq would have been capped by ice, which was not the case in c. 1850 according to a description by RINK (1857, I, p. 35).

Apparently the change in altitude of the glaciation limit within the historical period has been 100–200 m over the major part of the area. This would correspond to a change in the mean summer temperature of 1°, which is the recorded value. As explained above, fluctuations in precipitation may locally have contributed to the movement of the glaciation limit. The deduced fluctuation of the glaciation limit is of similar magnitude to that expected from the observed climatic temperature fluctuations.

For the determination of glaciation limits of glaciations older than the last part of the historical period, other criteria than those used for the present glaciation limit have to be used. Their evaluation depends on the exactness to which it is possible to determine the outline of the former glacier from the remnants of its moraines. If the form of the firn basin and the size of the glacier lobe can be estimated, the approximate position of the firn line can be found by comparison with an existing glacier of similar shape and size. Under such ideal conditions the error in the determination of the height of the supposed glaciation limit is estimated with an error less than 200 m.

The above method is valid for local glaciations only and it is not possible to determine former glaciation limits for the Inland Ice or even for larger local ice caps. However, it is believed that marginal interlobate and nunatak moraines are not formed above the glaciation limit. This is a consequence of the flow line theory of glacier movement by FINSTERWALDER (1897).

4. GLACIER FLUCTUATIONS IN HISTORICAL TIME

Fluctuations of glaciers can be considered as fluctuations of their frontal margins, their areas or their volumes. In the following section, however, most attention will be paid to frontal fluctuations for the following reasons:

1) Too few maps of sufficient detail (i.e. a scale of at least 1:10,000 and a contour interval of 10–25 m) exist to enable the computation of fluctuations in more than one dimension.
2) By far the greatest amount of information concerns frontal fluctuations.
3) Frontal and area fluctuations are the most easily computed in the case of prehistoric fluctuations.

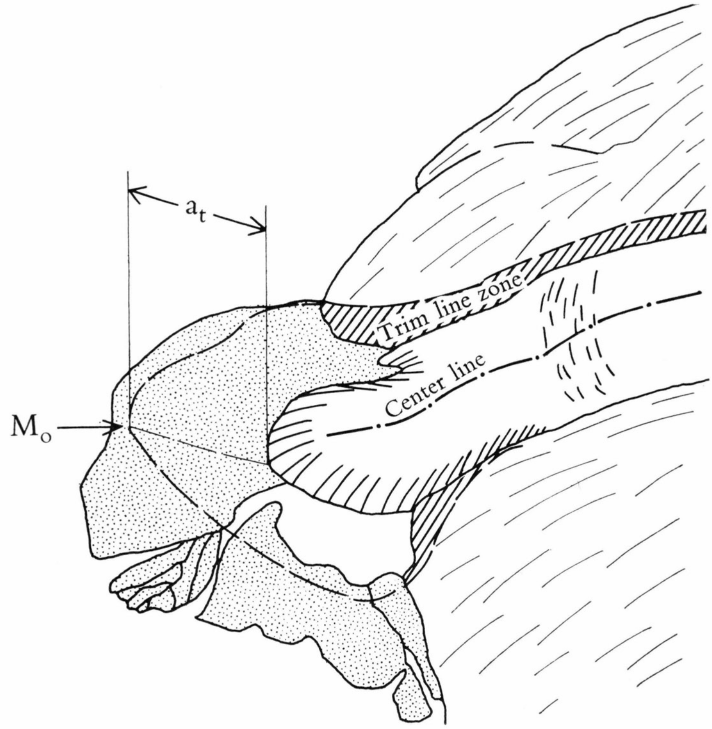
4.1. Frontal fluctuations

The exact nature of the relationship between oscillations in climate, glacier mass balance and the frontal fluctuations of glaciers is a subject for widely differing opinions. AHLMANN (1953, p. 14) accepted that some general connection exists, in particular between the summer temperature and its influence on the length of the ablation period, whilst LLIBOUTRY (1965, p. 836) stresses the influence on small glaciers of both summer temperature and winter precipitation. However, MELLOR (1964, p. 104) states that: "the popular notion that all glaciers expand if the climate gets colder, and recede if the climate gets warmer, is completely unjustified as a generalization" and a glance at the tables of the current positions of the glacier fronts in the Alps and in Scandinavia (J. Glaciol, Zeitschr. Gletscherkunde and Jøkull) shows also that individual glaciers within the same period and in the same area can behave differently. Such difference may be due to the influence on the mass balance of the glaciers by other meteorological elements besides the temperature, as well as by the subglacial topography and/or the dynamics of the glacier. A certain delay in a glacier's response to climatic change is generally accepted. AHLMANN (1953, p. 8) considered that a small glacier would react faster than a large one because of the smaller area of firn and stated furthermore (*ibid.* p. 9) that the "variations in the positions of the termini of outlet glaciers from inland-ice masses are related to the supply of ice as it was determined by climatic conditions of a long time before". A theoretical treatment of "kinematic waves" in glaciers by NYE (1960) shows that secular alterations are transferred through a glacier with a speed 2 to 5 times greater than the movement of the ice in the glacier lobe, and that the time before frontal response for a common valley glacier is considered to be between 3 and 30 years, whilst for typical "ice sheets" it is thousands of years. Although KAMB (1964, p. 361) says that the discovery of the delay raises serious questions about the interpretation of glacier behaviour, he admits that the effect of the kinematic waves on glacier margins may be weakened by smoothing or interference. It is commonly agreed that there is a response delay of some years or decades on glaciers, but the extreme consequences of the theory must be treated with caution. For instance, the rapid response of the Scandinavian ice sheet to the climatic fluctuation of the younger *Dryas* time and of the Inland Ice to the fluctuations of historical time throw doubt on the existence of long delays in large ice sheets.

In order to study the problem of fluctuations of the glacier lobes in West Greenland and their connection with climate, data concerning the position of the glacier front in the historical period have been compiled for as many glaciers as possible, in total about 500 glacier lobes. The uncertainty in the number is due to several glaciers having joined or divided during the period. For most of these glaciers the interval for which observations are available is only the last 20–40 years, during which time they have been in constant retreat. The remainder, for which there are observations covering this century at least, are listed in plate 2. These total 135 glacier lobes, from Kap Farvel in the south to Upernavik district in the north. The position of the glacier lobe localities investigated are given in plate 1.

In the diagrams in plate 2 the abscissae are in years A.D. and the ordinates are the horizontal distance in kilometres (a_t) of the glacier front from the outermost point of the historical moraines or the trim line zone in the centre line of the glacier lobe (see fig. 8). Thus an upward trend of the curve indicates a glacier recession. Because of the variability of the shape of the glacier fronts, occasionally changes in form will be expressed as false minor advances and retreats in the diagrams. Attention must be paid to the behaviour of all glaciers in an area rather than to that of an individual glacier.

The trim line zone marking the maximum extent of the glaciers in historical time is clearly developed at nearly all glaciers. On gneissic rock this zone is especially clear, strongly contrasting in its fresh colour



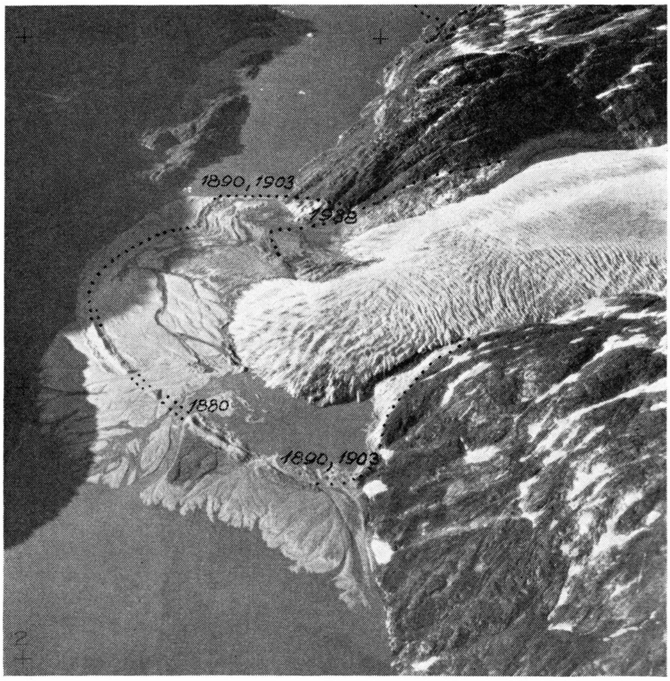


Fig. 8. Above. Significance of the parameters used in plate 2, a_t (length) and M_o (height a.s.l.), illustrated for "Søndre Qôrnoq glacier" (loc. 14A, plate 2). Below: Aerial photograph of "Søndre Qôrnoq glacier". Marginal positions in 1880, 1890, 1903 and 1938 indicated. Geodetic Institute's route 502 H–N, no. 8294 (23.7.1948). Copyright Geodetic Institute.

with the surrounding weathered rock. The same is true to a lesser degree in glacial and fluviatile deposits, but in basalt terrain, the outer margin of the trim line zone is at times difficult to determine on aerial photographs because of the dark colour of the terrain. However, here as elsewhere the outermost fresh moraines correspond to the outermost and lowermost part of the trim line zone. The lowermost point of the trim line zone in the centre line of a former glacier lobe is marked M_o (see fig. 8) and its height above sea level is given for each example in table 2, pp. 162–187.

4.2. Sources of error

The data used in the determination of the frontal position of the glaciers have been obtained from written descriptions, maps, terrestrial and aerial photographs and from direct measurements of the position of the glacier front in the field. The material concerning the fluctuations of the glacier lobes in the Julianehåb, Frederikshåb and Godthåb districts has been published previously (WEIDICK 1959). Part of this material, brought up to date, is included in plate 2 and table 2.

Written descriptions must be regarded as the least accurate source, though occasionally a description can give a precise indication of the position of the glacier. This is the case with PAARS' description of the Inland Ice margin in Austmannadalen, Godthåb district (loc. 23, plate 2), being situated in 1729 at the edge of a waterfall (PAARS 1936, p. 187). The only waterfall in the vicinity lies at the outermost limit of the trim line zone.

Such an unconsciously exact determination of the ice margin position can be trusted. In contrast to this, the local tales of land buried by ice must be treated with some sceptiscism. A classical case is the description of a "Bear Sound" stretching across Greenland which was inundated by an advance of the Inland Ice. The story that Jakobshavns Isfjord once extended to the east coast must be regarded as an offshoot of this legend (LARSEN & MELDGAARD 1958, p. 28, WEIDICK 1959, p. 178). The information quoted by JENSEN (1889, p. 70) about the earlier existence of a connection between Evighedsfjord and Kangâmiut kangerdluarssuat ("Kangerdluarssuk fjord", loc. 50 and 51, plate 2) in Sukkertoppen district is similarly dubious. The glacier lobe separating the two fjords has now retreated so far for it to be seen that half of the front rests on a c. 100 m high rock threshold, which seems to continue under the other half as well.

As early as 1770 CRANTZ (I, § 11, p. 40) mentions and compares the expansion of the glacier lobes in the Alps with those in Greenland, and around 1770 FABRICIUS (1788, pp. 69–70) reported a measurement of

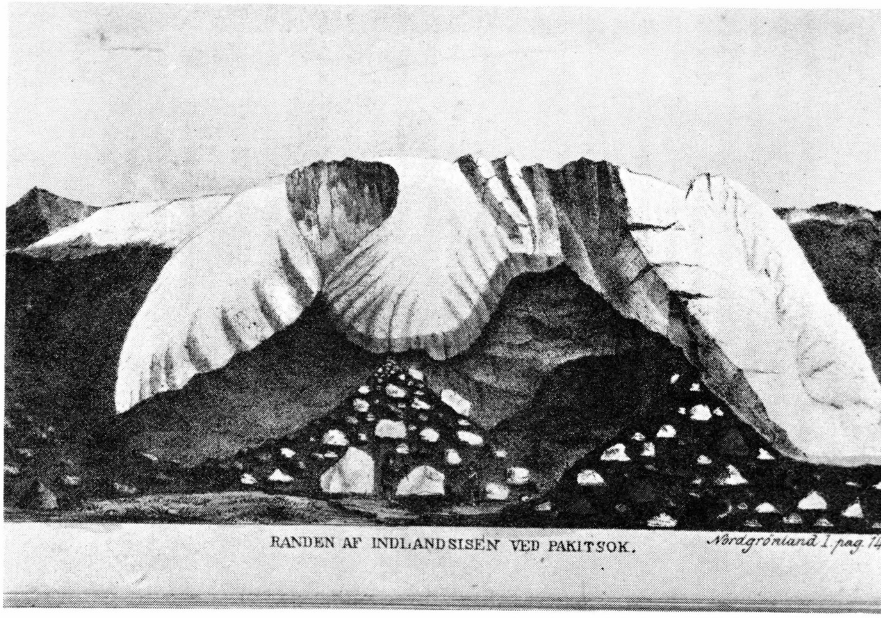


Fig. 9 a.

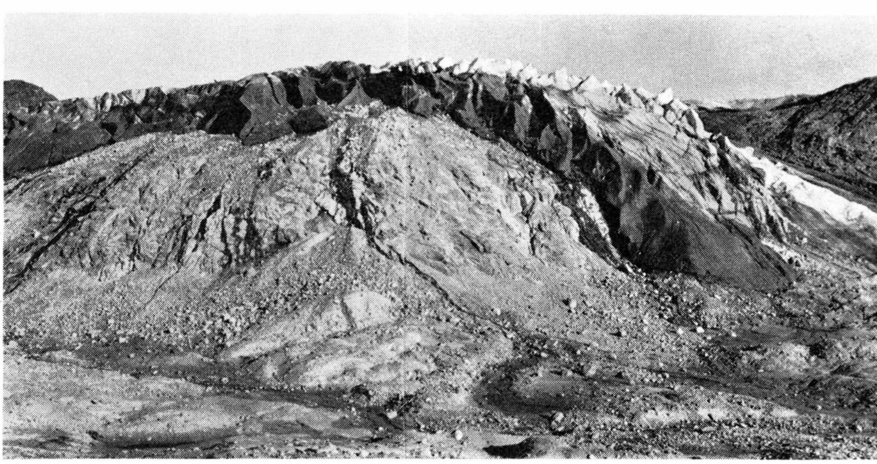


Fig. 9 b.

Fig. 9. The Inland Ice lobe at the head of Pâkitsup ilordlia (Qíngua kujatdleq, loc. 72 in plate 2). 9a. Drawn by Rink in c. 1850 (Rink 1857, p. 14). 9b. Photograph, 23.7.1961, Weidick. 9c. Sketch by Hammer drawn in 1883 (Hammer 1889, table III). The lobe marked "c" is the area glaciated since c. 1850. 9d. The Inland Ice lobe, shown from Hammer's position. Photograph, 21.7.1961, Weidick.

the advance of a glacier. Unfortunately Fabricius did not state exactly which glacier this was, but it is supposedly one located at the Inland Ice margin in the Frederikshåb district, in which he was the incumbent.

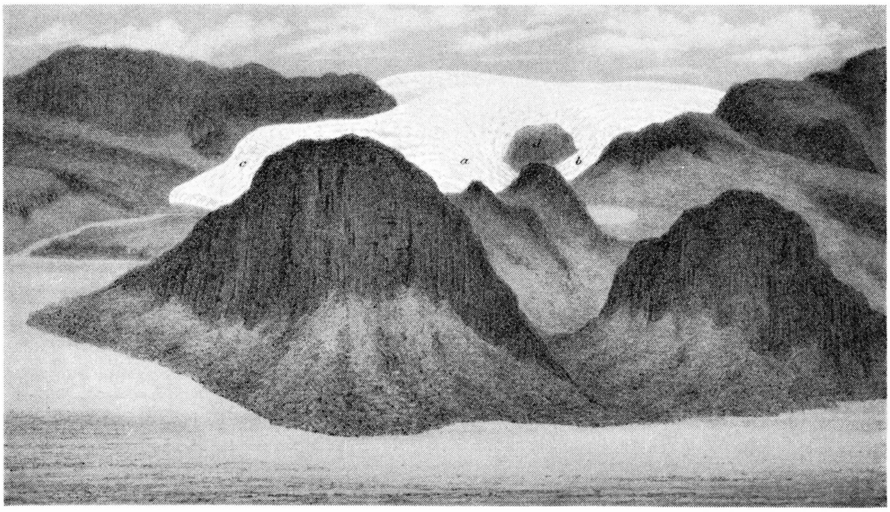


Fig. 9c.

Fig. 9d.

Thus it is possible that there was a real and even widespread expansion of glaciers in the 18th century which formed a background for the legends mentioned above.

On the early charts and maps errors in shape and scale are apparent from the inaccurate outlines of coast and glaciers. Moreover, the outlines of glaciers shown on them are frequently derived from older and less accurate sources. The best of the early maps are those drawn by the reindeer hunters and compiled by RINK in the middle of the 19th century. These maps are often strongly distorted in outline, but reveal a good knowledge of detail, which itself must indicate first hand acquintance with the area by the drawer of the map. The areas covered in this way were the little known eastern parts of the coastal area in West Greenland near to the Inland Ice margin. For most maps and charts the error in dating is possibly of the order ± 5 years.

Of the drawings, sketches and other pictures used those of an early date are an especially valuable source *e.g.* RINK's own drawings from the middle of the 19th century and the sketches of HAMMER (1889) and MOLTKE (STEENSTRUP 1901, MYLIUS-ERICSEN and MOLTKE 1906). However, a misinterpretation of pictorial evidence can take place. For example, HAMMER (1889, p. 15), after visiting Pâkitsup ilordlia, stated that the glacier lobe there had not altered position since it was sketched by RINK around 1850. A visit to this place in 1961 made it clear that HAMMER had misinterpreted RINK's sketch and that the glacier lobe had in reality advanced nearly half a kilometre between 1850 and 1883. Not until c. 1950 had the lobe (loc. 72, plate 2) retreated so much that its form was like that of 1850 (Fig. 9).

Terrestrial photographs are a reliable source though misinterpretations similar to that mentioned above are possible. It may be added that glacier photographs using a telescopic lens were taken as early as 1898 by STEENSTRUP of the glacier near Sarqaq on the south coast of Nûgssuaq peninsula (loc. 103 in plate 2). The dates of only a few of the photographs are known precisely. However, when the identity of the photographer is known, the date of his visit to the locality can often be established within a few years.

Aerial photographs are the most exact source of information available. The earliest pictures from West Greenland seem to have been taken in 1932 by the Geodetic Institute in the Julianehåb district and by the German "Universal Dr. FANCK Expedition" to Umanak district. A large number of photographs from 1936 and 1937 from various parts of West Greenland, are now in the files of the photogrammetric section of the Geodetic Institute.

During the second world war aerial photography was continued by the U.S. Air Force and after the war again by the Danish Geodetic Institute. Several copies of the series made during the war are kept in the Geodetic Institute but, unfortunately, the exact date of these is not known. They have been tentatively assigned to the years of 1942 and 1943 in plate 2, though 1944 and 1945 cannot be excluded.

Whilst the measurements of distance to the glacier front from a fixed point is normally the most accurate source, in some circumstances it may involve great errors when there is some confusion over the reference point used. This was STEENSTRUP's conclusion (1883b, p. 224) after his attempt in 1879 to duplicate HELLAND's measurements of 1875 on the glacier Agssakait sermiat ("Asakak-Bræen") on the north coast of Nûgssuaq. Unless a map or a sketch is given together with the measurements, their values can not be regarded as being as exact as they seem.

Some control of the information compiled from historical sources has been provided by the lichenometrical determination of the age of the deglaciated areas, *e.g.* in Søndre Strømfjord (BESCHEL 1961).

The data from all the various sources is summarised in plate 2. Taking into account the possible sources of error outlined above it seems probable that the error in the location of the glacier front for the more recent data is only a few metres, whilst for the older it may often be as much as several hundred metres. However, by considering the cumulative trend of the fluctuations of all glaciers the effects of erroneous information, and of the anomalous behaviour of individual glaciers are minimised.

Neither literary sources nor lichenometrical dating extend back to the earliest phases of the glacier advances in historical times. However, one locality in the Thule area, has yielded a radiocarbon date for organic material in shear moraines from an historical advance of 520 ± 200 years B. P. (GOLDTHWAIT 1961, p. 108). This date, around A.D. 1430, agrees well with the dates for the early part of the historical advance in Europe (AHLMANN 1953, p. 40).

4.3. Results

A glance at plate 2 shows that a small number of the glacier lobes have had remarkably large fluctuations, a few others no fluctuations at all, whilst the great majority have had moderate fluctuations of about the same magnitude. This subdivision provides a basis for the following arbitrary classification of frontal fluctuations of glaciers.

1. First order fluctuations, *i.e.* fluctuations greater than 5 km.
2. Second order fluctuations, *i.e.* fluctuations between 5 and 0.5 km.
3. Third order fluctuations, *i.e.* quasi-stationary glaciers or glaciers which have been expanding so that their extent to-day is their maximum extent during historical time.

4.3.1. Inland Ice margin

4.3.1.1. First order fluctuations

All the glacier lobes with extremely large fluctuations are calving lobes with a high production of calf ice, which supposedly drain major parts of the Inland Ice margin (DANSGAARD 1961, pp. 89–91). However, high productivity alone is not a criterion for large fluctuations and other very productive glaciers, such as Eqalorutsit kangigdlît sermiat in the Julianehåb district (loc. 8), Sermeq avangnardleq in Torssukåtak (loc. 76) and Store Gletscher ("Store Qarajaq") in the Umanak district (loc. 125, plate 2) have all had nearly stationary margins throughout the last 100 years.

The magnitude and production of the glacier lobes with first order fluctuations are given below.

Eqalorutsit kitdlît sermiat, Julianehåb district (loc. 9, plate 2)

The fluctuation has been c. 5 km. The present width of the lobe at the front is c. 2 km and according to LOEWE (1936, p. 327) the production of calf ice here, together with that at Qôrqup sermia, Julianehåb district, and Narssalik, Frederikshåb district, totals c. 3 km³ annually. However, if JESSEN (1896, p. 99) is correct in his estimate of the surface velocity of Eqalorutsit kitdlît sermiat as c. 1 m/hour, the productivity of this glacier lobe alone must be about 3 km³ annually.

Kangiata-nunâta sermia, Godthåb district (loc. 26, plate 2)

The fluctuation has been c. 20 km. Its present width is 4–5 km and its production of icebergs, though not as great as that of Eqalorutsit kitdlît sermiat mentioned above is appreciable (see fig. 10).

Jakobshavns Isbræ, Disko Bugt (loc. 70, plate 2)

The fluctuation has been c. 26 km. The width of the lobe is c. 6 km and its productivity, after LOEWE (1936, p. 327) and BAUER (1955c, p. 99) is c. 16 km³ annually.

Umiámáko glacier, Umanak district (loc. 134, plate 2)

The fluctuation has been c. 5 km. The present width of the front is 2–3 km and its iceberg production, together with that of Rinks Isbræ, is estimated by LOEWE to be c. 19 km³. However, since the surface velocities of the two glaciers are in the ratio of 1:3, and the width of the front of Umiámáko glacier is only about half that of Rinks Isbræ, it is probable that the production of Umiámáko glacier is little more than c. 3 km³ annually.

Upernaviks Isstrøm, Upernavik district (loc. 142, plate 2)

The fluctuation has been c. 22 km, the frontal width of the glacier lobe is 6–10 km and its productivity is about 12 km³ annually (LOEWE 1936, p. 327, BAUER 1955c, p. 99).

MERCER (1961a, pp. 856–857) has shown that calving glaciers, in order to overcome the need for long extensions of their floating tongues in response to changes in their mass balance, tend to place their fronts at widenings in the fjord. Consequently a small change in length will

Fig. 10. Kangiata-nunâta sermia area. In the foreground and nearest the fjord are the trim line zone and moraines from the 18th century (a–a) and from the readvances in 1850–1890? (b–b). Deposits from the outer zone are to be seen on the left (dotted) on the uppermost part of the hill sides (c–c). Geodetic Institute's route 506 B–N no. 6270 (21.8.1948). Copyright Geodetic Institute.

greatly alter the ablation area. The fronts of many of the glacier lobes mentioned above seem to have been situated near such points of widening in the fjords for long periods, in agreement with Mercer's principle. Thus the long halt in the recession of Kangiata-nunâta sermia, from 1850–1921, may be explained in this way.

The best known of these glaciers, Jakobshavns Isbræ, showed a tendency to halt its recession around 1883–1893 and again around 1921 but because of the regular form of the fjord, Mercer's principle does not seem to explain these halts and Bauer (1955c, pp. 55–56) has suggested that they were climatically controlled.

In general, all lobes have experienced a major recession in the last 100 years. Furthermore the curves in plate 2 show that the southernmost

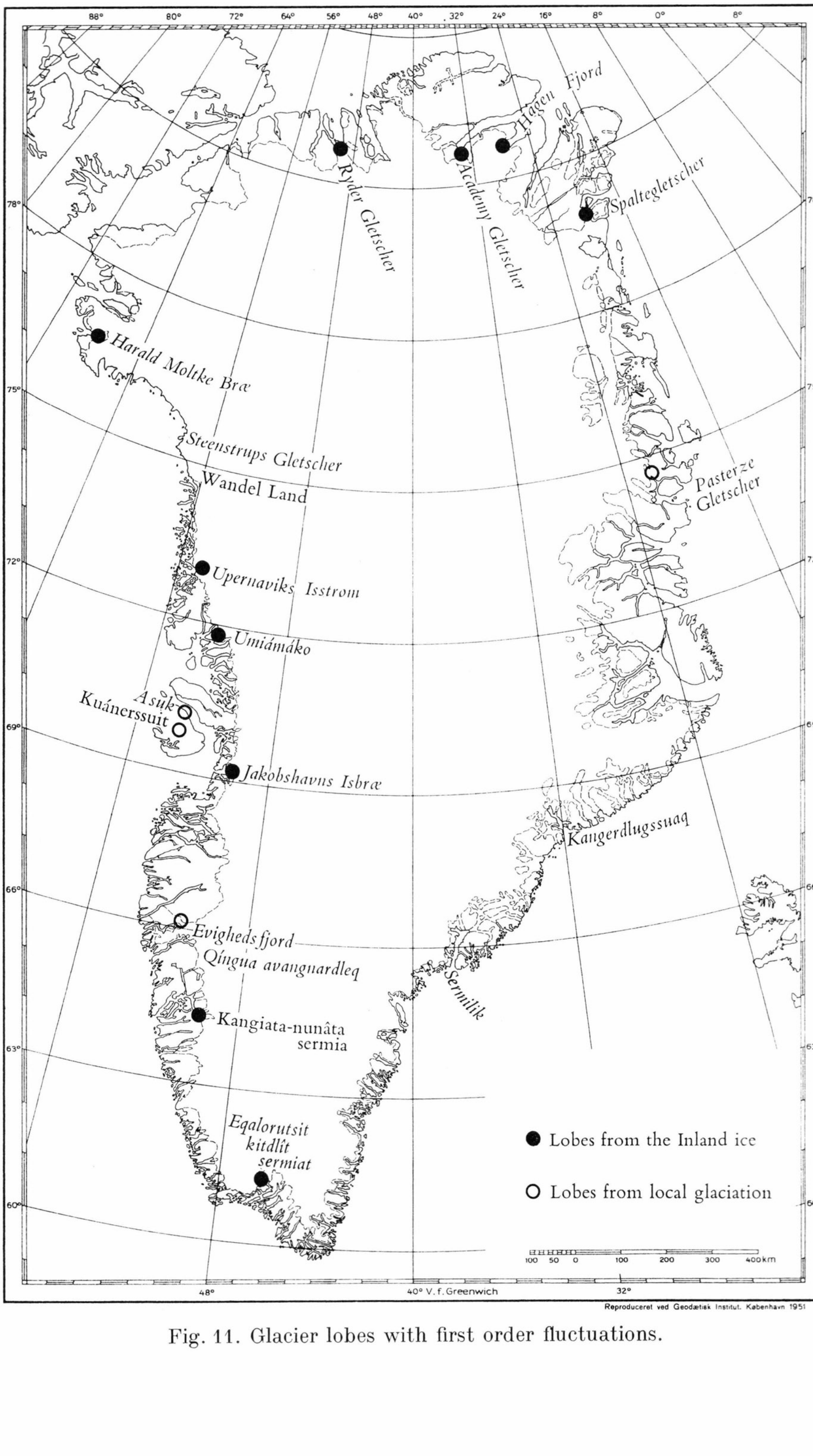


Fig. 11. Glacier lobes with first order fluctuations.

glacier lobes seem to have been receding since before 1800, whilst the three northernmost lobes did not begin to recede until after the middle of the 19th century, with their main recession occurring in the 20th century.

First order fluctuation have occurred elsewhere in Greenland as shown in Fig. 11. L. KOCH (1928, pp. 196–202) reports great fluctuations of the Inland Ice margin in Melville Bugt, though because of the unconfined nature of the margin there the fluctuations are difficult to describe quantitatively. Nevertheless KOCH's descriptions and maps seem to indicate that between 1916 and 1920 the Inland Ice margin there had an extent close to its historical maximum. Aerial photographs from 1948 and 1953 show a subsequent first order retreat of some sectors. A trim line zone on skerries and nunataks in the same area indicates also a thinning of the ice margin (f. ex. around Wandel Land).

The best known glacier fluctuation in North Greenland are those of the 7 km wide *Harald Moltke Bræ* near Thule air base, which has retreated 5–5.5 km since it reached its greatest historical extent around 1930. DAVIES and KRINSLEY (1962, pp. 119–130) summarise the data about this glacier as well as of other lobes in North Greenland. According to them *Academy Gletscher* had retreated 12 km by 1956 from a maximum position in 1920. *Spaltegletscher* also had a first order recession of c. 18 km (DAVIES and KRINSLEY 1962, p. 127), the recession occurring mostly between 1907 and 1938. Since nothing is known of its position between these two years it is possible that here also there was an advance or period of no change around the 1920's. The same can be said for a lobe in Hagen Fjord (DAVIES and KRINSLEY 1962, p. 127).

In general the frontal fluctuations of the North Greenland calving lobes are difficult to describe. Recession has occurred more by a process of terminal disintegration with the calf ice production of several years calving on a single occasion (KOCH 1928, pp. 199–200, AHNERT 1963, pp. 537–545).

Very little is known about the major lobes of the Inland Ice in East Greenland. The two which BAUER and HOLTZSCHERER (1954, p. 36) include as "outlet channels" of the Inland Ice are *Kangerdlugssuaq* and *Sermilik*, both of which according to literary evidence and aerial photographs, have had only a small recession in most recent times (HOLM and GARDE 1889, DE QUERVAIN and MERCANTON 1925 and THORARINSSON 1952).

4.3.1.2. Second order fluctuations

Of the 52 lobes of the Inland Ice margin described here a total of 35 have had fluctuations of the second order, *i. e.* 67 %. Of these, 20 calve in the sea, though one of them now only partially (loc. 18 in plate 2)

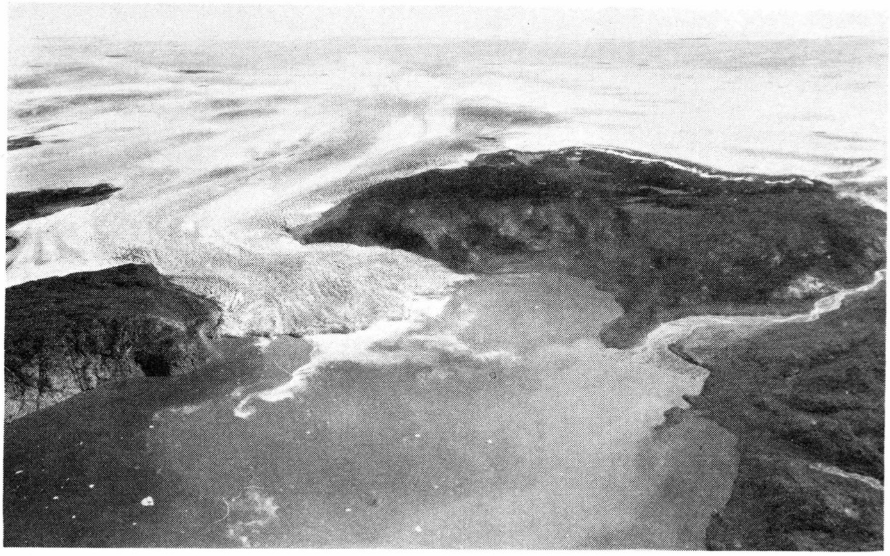


Fig. 12. Eqip sermia (loc. 74, plate 2) seen from the west. Geodetic Institute's route 518 A–Ø, no. 1493 (18.7.1949). Copyright Geodetic Institute.

whilst the other 15 glaciers calve in lakes or have fronts resting on the land. In both the Julianehåb district and Disko Bugt calving lobes generally show a tendency for greater fluctuations than lobes ending on land, though this difference cannot be demonstrated from other areas.

In the sector from the Julianehåb district to the Godthåb district, only *the glaciers of Austmannadalen* (*loc. 23 in plate 2*) had a major part of their trim line zone deglaciated before 1800, the majority of the glacier lobes elsewhere still having an extent around 1850 close to their historical maximum. Areas deglaciated before c. 1800 are also reported from the Inland Ice margin in the Julianehåb district itself.

The readvances which culminated between 1890 and 1900 seem to have stopped or delayed the general recession which began earlier in the 19th century. In some places in the south the readvance of 1890–1900 covered nearly all of the trim line zone (*Sermilik or "Sermítsialik" glacier*, no. 12 in table 2 and "*Søndre Qôrnoq glacier*", no. 14A in table 2, shown in fig. 8).

Further north in Disko Bugt, the readvances of 1890 can be seen to have given glaciers generally an extent near to their historical maximum. However, the lobe in *Påkitsup ilordlia in Qíngua kujatdleq* (loc. 72, shown in fig. 9) is peculiar in that it advanced beyond the earlier maximum, over vegetated areas.

In only a single case in West Greenland has a readvance in the 20th century given a lobe nearly its maximum extent for historical time. This

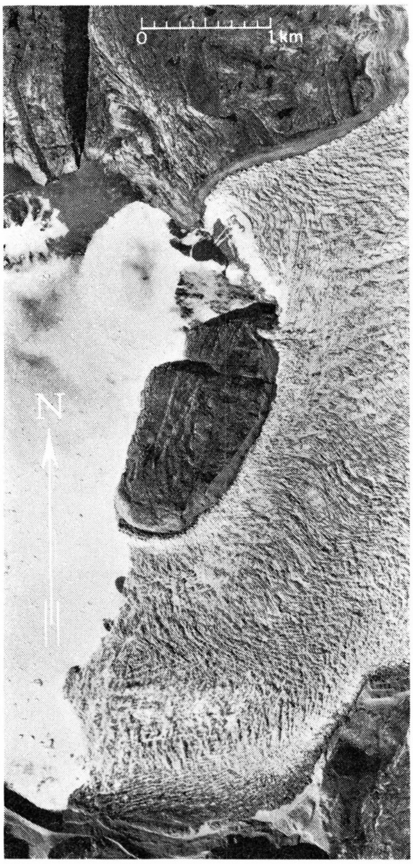


Fig. 13a

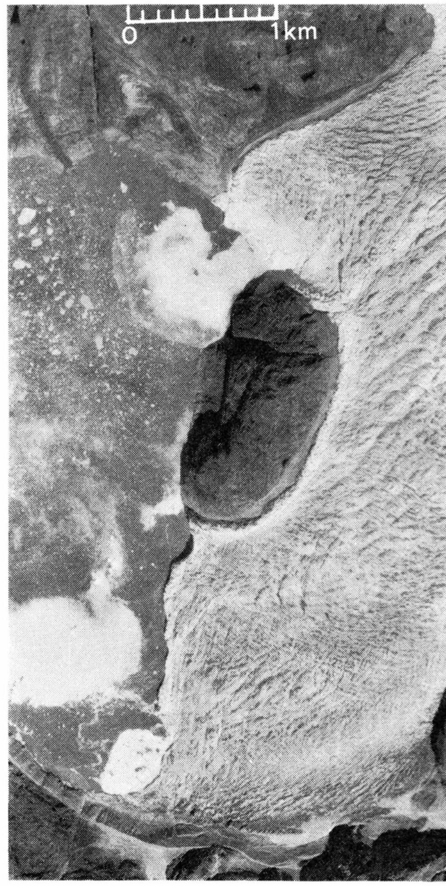


Fig. 13b

Fig. 13. "Kangigdleq glacier", Itivdliarssûp kangerdlua, Umanak district (loc. 128, plate 2). 13a. Photograph of the 29th July 1953. Geodetic Institute's route A102/94, no. 159. 13b. Photograph of the 25th June 1959. Geodetic Institute's route 238 F, no. 148. Copyright Geodetic Institute.

happened at *Eqip sermia* (loc. 74, shown in fig. 12 and 36) around 1920 (Bauer 1955c, p. 53). Its advance can be correlated with a halt in the recession of other lobes, *e.g.* Jakobshavns Isbræ.

In spite of the large amount of data available, little is known about the fluctuations of the glacier lobes in the Umanak district since the annual fluctuations of the calving glaciers there are often so large that they mask longer term fluctuations of the fronts. However, it seems apparant that, on the whole, the ice margin was almost at its most advanced position around 1890–1900. The late date of the retreat from there is suggested by the very fresh looking moraines along the Inland Ice in this region. For the lobes of Upernavik district also, a small and short recession around and after the middle of the 19th century was interrupted by halts or readvances between 1890 and 1900.

After c. 1920, a general recession of all glaciers set in, the rate of retreat reaching a maximum between 1920 and 1940, since when it has slowed down. The most recent information from the northern parts of the area, in Disko Bugt and Umanak district, from the late 1950's and the beginning of the 1960's suggests that the glaciers there are showing signs of readvancing. Whilst this may be due to the annual variation of a nearly balanced ice margin, at one locality, Kangigdleq (loc. 128 in plate 2) a small area has been reglaciated between 1953 and 1959 (see fig. 13).

4.3.1.3. Third order fluctuations

In this group two categories are distinguished:

a) Where the ice margin has been relatively stable, with fluctuations of between 0 and 500 m which are in phase with those of the first and second orders, and

b) Where the ice margin has had fluctuations out of phase with the "normal" first and second order fluctuations.

9 lobes belong to the first group, of which 3 calve in fjords, the rest ending in lakes or on land. These lobes are the localities 28, 31, 33, 33 (B_3), 35, 36, 39, 67 and 76 in plate 2. Fluctuations of this category have occurred widely along the Inland Ice margin and must be considered as the normal type. On aerial photographs such places show only a small or no development of the trim line zone. In areas visited on the ground, *e.g.* in Søndre Strømfjord area, Disko Bugt and Nûgssuaq, the glacial moraine landscape from this time shows a condensed development of the moraine succession known from the first and second order fluctuations.

Because of the decrease of ablation with altitude (see fig. 7) it is expected that the recession of the Inland Ice margin, even at altitudes only a few hundred metres above sea level, must be significantly less than at sea level. Though large parts of the Inland Ice margin at high elevations will have third order fluctuations because of this effect alone, such fluctuations are characteristic also of many other parts near to sea level, *e.g.* most of the examples quoted in plate 2 with the exception of Orpigsôq (loc. 67), and large parts of the Inland Ice margin which lie near sea level in Egedesminde district, Disko Bugt and Nûgssuaq.

The advances of 1880–1900 seem often to have been the historical maximum for this group. For example, J. A. D. JENSEN in 1884 reported that the Inland Ice margin in the area around Isordlerssuaq nunataks at Søndre Strømfjord was ploughing up fresh vegetation (JENSEN 1889,

p. 65). From the same area, NORDENSKIÖLD (1914, p. 633) reported that at the more northerly lobe *Isúnguata sermia*, the surface right up to the ice was covered with vegetation and only at the ice margin itself was there occasionally a bare zone. JENSEN (1881, pp. 140–144) also reported that at another lobe (*Inugpait qûat*, loc. 36 in table 2) the trim line zone in 1879 was either lacking or had a maximum width of only 100 feet (30 m).

On Nûgssuaq peninsula, it seems apparent that the ice margin was almost at its maximum extent around 1890–1900. Comparison of the observations made on the trim line zone at the same locality by DRYGALSKI (1897, p. 120), BARTON (1897, pp. 218–219) and the time of the author's visit in 1961 confirms this assumption. Whilst the trim line zone in 1893 and 1896 was only a few metres high, in 1961 it was 30–50 m above the present ice margin.

For Upernavik district there is RYDER's report (1889, pp. 212–213, 225), from his visit to the south of Upernaviks Isstrøm in 1886, that the Inland Ice margin was a short distance from old vegetation. Also there were local stories of the ice margin having advanced recently. At "*Cornell Glacier*" in Ryders Isfjord (loc. 147), TARR (1897, pp. 257–268) reported that there had been only slight thinning of the ice margin along the glacier.

The three glacier lobes which comprise group b) are all producing calf ice. One of them, *Eqalorutsit kangigdlît sermiat* (loc. 8, plate 2) has a large production of calf ice, the other two, *Sarqardliup sermia* (loc. 68) and *Alángordliup sermia* (loc. 69) only very small. Eqalorutsit kangigdlît sermiat was at its historical maximum and was still expanding in 1955 (WEIDICK 1959, p. 61). The sector of the Inland Ice margin north of Nordenskiölds Gletscher which includes *Orpigsôq*, *Sarqardliup sermia* and *Alangordliup sermia* has recently been stationary near to its historical maximum or expanding. It is known that an earlier expansion of Sarqardliup sermia buried a nunatak ("Nunataranguaq", see fig. 14) in c. 1900 (HAMMER 1883, KOCH and WEGENER 1930, pp. 382–391). There are also the more dubious expansions of *Sermilik* glacier, Frederikshåb district and *Ujaragssuit pâvat*, Godthåb district (WEIDICK 1959, pp. 111 and 173), not listed here, in the middle or at the end of the previous century.

4.3.1.4. Inland Ice, conclusions

There is broad agreement between the phases of the fluctuations of the first, second and third order fluctuations, the oscillations of 94 % of the glacier lobes studied being in phase, despite their wide geographical distributions, the variations in their subglacial topography, size and dynamics and such secondary controls as MERCER's principle for calving

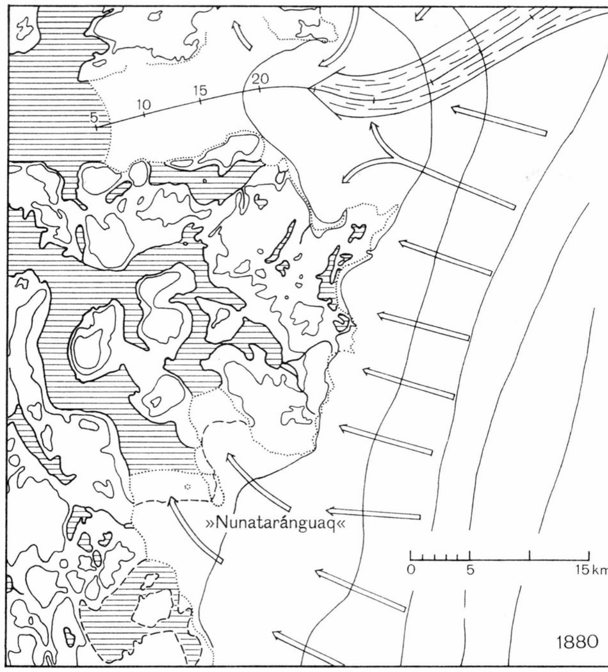


Fig. 14 a

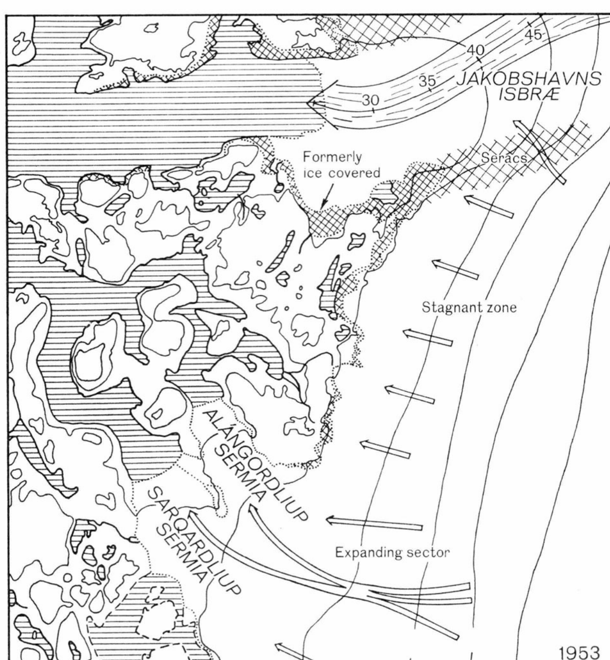


Fig. 14 b

Fig. 14. Jakobshavns Isbræ, Alángordliup sermia and Sarqardliup sermia glacier lobes, Disko Bugt. 14a. Conditions in the sector at 1880 when the ice margin still had an extent near the historical maximum. 14b. Conditions in 1953. Dense cross-hatching: Trim line zone. Open cross-hatching: Thinning, strongly crevassed area. Note the advance of the ice margin in the Sarqardliup sermia area. Based on the Geodetic Institute's map sheets 1:250,000 68V2, Christianshåb and 69V2, Jakobshavn. By permission of the Geodetic Institute.

glaciers (p. 36). The major trends of the fluctuation therefore cannot be the result of local factors but must be climatologically controlled, and, of the two major variables, precipitation and temperature, only the changes of the latter are uniform in trend over the whole area.

Of the 40 glaciers with fluctuations of first and second order, 39 have been receding since the middle of the 19th century, whilst one (Eqip sermia, loc. 74 on plate 2) showed strong readvance around 1920. This retreat was interrupted by periods of readvance or halt around 1890 and 1920. During the readvance around 1890 many glacier lobes reached their maximum historical extent, whilst the readvance of 1920 was in general of lesser extent. The temperature records (see p. 18) indicate that cold periods occurred in the years 1880–1890 and 1913–1916 suggesting a delay in the reaction of the glacier lobes of between a few years and two decades.

The deviations from the general trend which occurred along isolated sectors of the Inland Ice (category 3b) may be explained by the capture of a drainage basin by another sector of the Inland Ice. For example, it is possible that Eqalorutsit kangigdlít sermiat has captured parts of the drainage basin of the neighbouring Eqalorutsit kitdlît sermiat (loc. 8 and 9, plate 2). In the region south of Jakobshavns Isbræ, the sinking of the surface of the Inland Ice by 200–300 m will have allowed a threshold, which can now be seen stretching eastwards under the ice margin, to prevent Jakobshavns Isbræ from receiving ice from the southern border of its lobe (fig. 14). Its diversion to the lobes of Sarqardliup sermia and Alángordliup sermia would compensate for the general thinning of the Inland Ice and keep them in an advanced position. Whether the same mechanism is responsible for the behaviour of the margin south of Nordenskiölds Gletscher cannot yet be decided.

4.3.2. Local glaciations

4.3.2.1. First order fluctuations

Three of the 83 local glaciers listed in plate 2 belong to this category. The best documented of these is the glacier of *Qingua avangnardleq VI-N* (loc. 57, table 2). The other two glaciers are *Asuk A* and *Asuk B–C* on Disko island (loc. 78, table 2). The wide trim line zone and extensive dead ice deposits around the Asuk glaciers demonstrate their former great historical extent. The first observations made at these glaciers were in 1898 by Steenstrup (1901, pp. 268–269), who described the glacier front as lying some distance inside the outermost limit of the trim line zone and an area of dead ice. From his descriptions from Disko island in general it is clear that in 1898 dead ice areas were widespread at glacier margins, especially along the northern coast along Vaigat

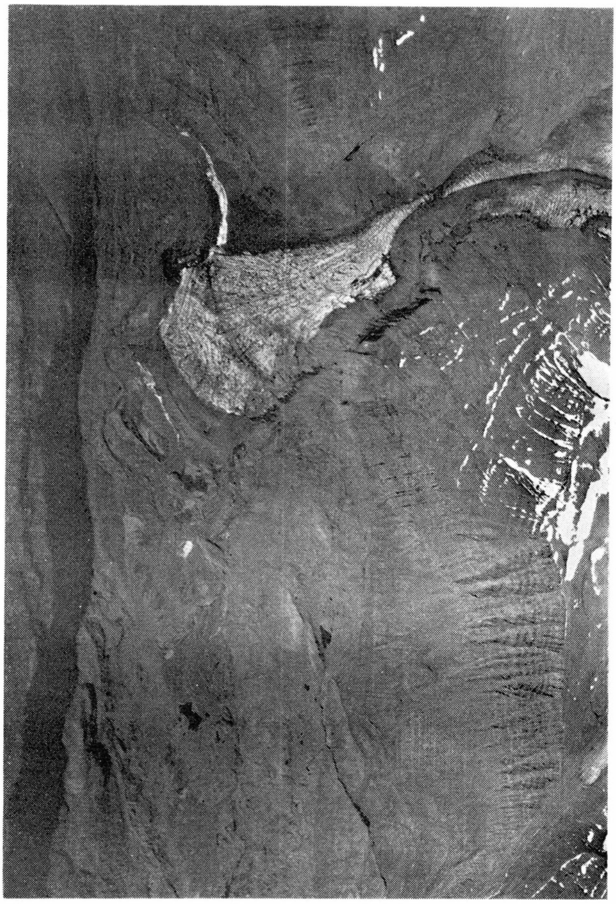


Fig. 15a.

Fig. 15. Sorte Hak, Kuánerssuit valley, Disko island. 15a. The valley is nearly deglaciated, but dead ice remnants are still present. Oblique photograph, Geodetic Institute's route B 37 B–L, no. 105, from the early 1940's. 15b. Dotted line, maximum extent of the glacier (midle of the 19th century?). Vertical photograph, Geodetic Institute's route A 70/134, no. 71 (20.7.1953). (a–a); esker.

sound. He also described how glacier lobes at *Ujaragssuit*, south of Qutdligssat, could be seen advancing over old dead ice. It seems likely that the widespread masses of dead ice were formed from glacier lobes which had been active earlier, in the middle of the 19th century at the latest. Another glacier on Disko island, in the Kuánerssuit valley in the central parts of the island, seems to have had a first order fluctuation. The glacier is shown in fig. 15, but it has not been included in the table because of the difficulty there of determining its historical extent. How-a

a

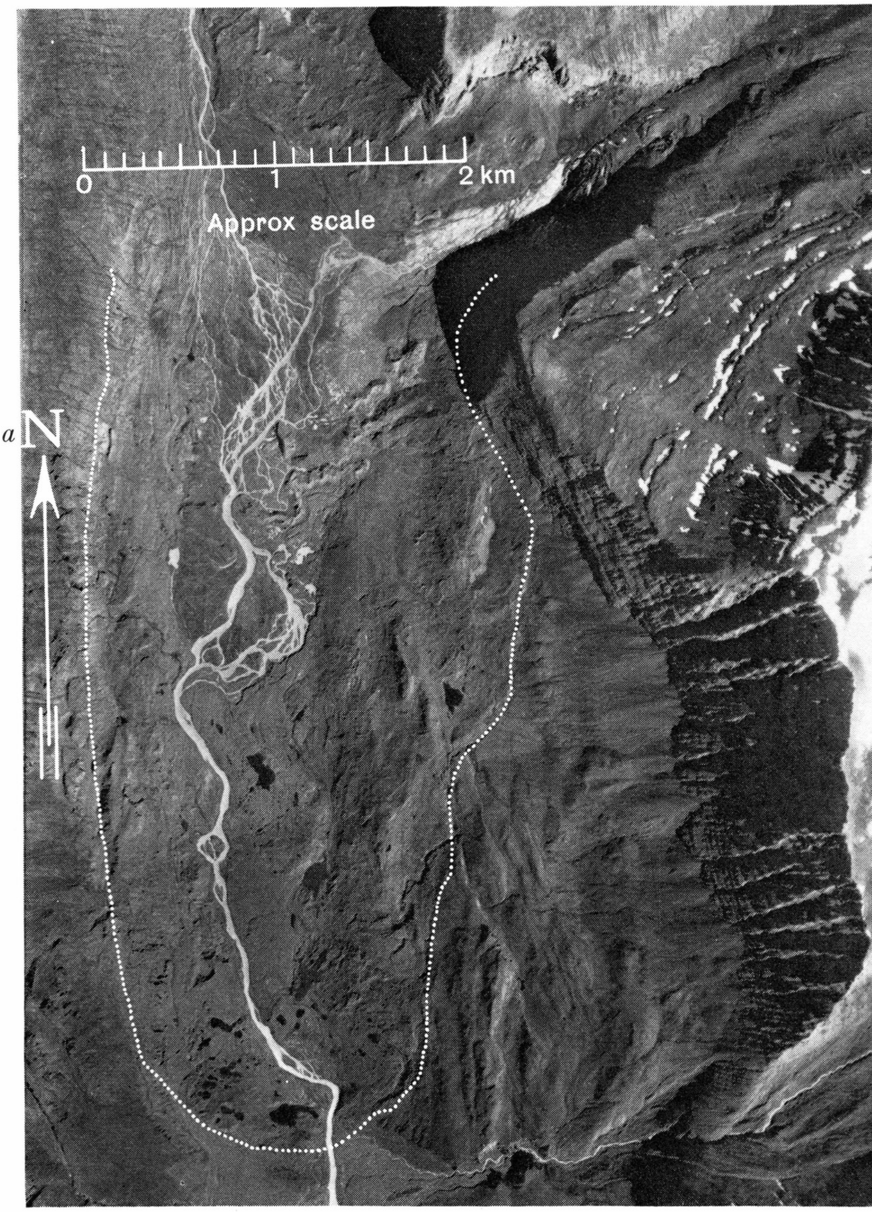


Fig. 15b.

ever, the descriptions from 1898 tell that the glacier lobe then was thinning, though it still reached nearly to the outermost historical moraines. In 1913 (Jost 1940, pp. 20–24) the glacier still filled a great part of the Kuanerssuit valley and it seems that it was not until after this date that a large part of the glacier became dead. However, the initial thinning of the glacier must have started already some decades before 1898.

A common feature of the glaciers mentioned here is that they are all individual lobes of former piedmont glaciations which filled valleys partly or totally by the confluence of their lower portions. All the glaciers are characterised by having their maximum extent for historical time at the middle of the 19th century or earlier.

From East Greenland, a first order fluctuation is known for *Pasterze Gletscher* at the head of Tyrolerfjord (see fig. 11), which between 1869 and 1938 receded 6 km (FLINT 1948, p. 125, SHARP 1956, p. 87). Between 1938 and 1948 the recession continued at a slower rate. This glacier seems to be a lobe from the system of transection glaciers in the East Greenland alpine terrain. It ends on land and has formerly been confluent with other glaciers. This type of glaciation is similar to the piedmont glaciations in parts of West Greenland.

4.3.2.2. Second order fluctuations

These comprise 55 of the 83 local glaciation localities, *i.e.* 66 %. Very few of them are calving lobes. Of these 55, lobes from local ice caps represent 51 %, valley glaciers 27 % and cirques 22 %, the three types being about equally represented in all regions. These figures are approximate as transitional types also occur. Included in the total number of 28 lobes from ice caps are 8 lobes, which were formerly of piedmont foot form, but which are now of the common tongue shape. These types tend to have greater recessions than other glaciers of the group. An exception to this is *Sermínguaq* (loc. 41, plate 2), which in form is transitional between the common tongue-shaped lobes from ice caps and a piedmont type. This glacier, which has a very small surface inclination at the snout, has had only small fluctuations.

It can be seen from plate 2 that the majority of the lobes in this category are found in limited areas near the coast. The information available about these is summarised, region by region, in the following.

Julianehåb-Godthåb districts

Glacier 1a, *Sermítsiaq*, must have begun to advance sometime before 1833 and in that year the glacier lobe already has an extent close to the historical maximum. This position was held until it began to thin around 1900, though the greatest part was not deglaciated until after 1920. In the outermost part of the trim line zone there are moraines older than 1830, which possibly were formed in the 18th century.

Locality 1b, "*Sermeq*", is known to have reached its maximum extent at about 1888 after an expansion of the lobe during the 1870's and 1880's. A rapid recession followed after 1900, which apart from a short halt at some date, has continued until recent years.

Other glaciers in the Julianehåb district are known to have been still close to their maximum historical extent in the last decades of the 19th century and to have had their maximum rate of retreat between 1920 and 1940.

Fig. 16. Qingaq glacier, Godthåbsfjord. Geodetic Institute's route 506 D–S no. 6468. Dotted line, maximum extent in historical time. Copyright Geodetic Institue.

For the areas north of Julianehåb district, there is little information. The lobe of *Kitdlavât* in the southernmost part of Godthåb district (loc. 21, plate 2) reached a maximum some time around the beginning of this century and the maximum rate of retreat occurred here after 1930. This glacier as well as the other example from Godthåb district, *Qingaq* (loc. 25, plate 2, also shown in fig. 16), are lobes which descend steeply towards the fjords from small cirques and a great part of their nourishment is probably from avalanches. Unlike most other glaciers, whose maximum period of retreat has extended into the 1940's, Qingaq has been stationary since 1930. The probable explanation is that after 1930 the glacier has used up most of the material in the firn basins, and it has since been nourished mainly by a rather constant supply of avalanches. This glacier belongs to the type known as "reconstructed" or "regenerated" glaciers. Qingaq glacier seems to have reached its maximum historical extent before or in the middle of the nineteenth century, though it remained close to it up to c. 1900.

In the Sukkertoppen district the maximum rate of recession was between 1920 and 1940 following a period around 1900 when the glaciers were in an advanced position near to their historical maximum. Descriptions of any readvances before 1900 are scarce for the district, though an advance of the glacier lobes of *Sermítsiaq* (loc. 50–51, plate 2) before the 1880's is recorded. Another exception is GIESECKE's report of a general expansion of the glacier lobes around Hamborgerland (loc. 61, plate 2: *Sermersût*) and "*Sermilik*" (*Manitsup sermilia*)-*Sermilínguaq* (loc. 59–60, plate 2) in the first decade of the 19th century (GIESECKE 1910, pp. 50 and 133). DE QUERVAIN and MERCANTON (1925, p. 178) say that the glaciers in "Sermilik" and Sermilínguaq fjords in 1912 were in a "décrue générale". However, recessions between 1900 and 1920 were generally very slow and the curves from the district show signs of interruptions in the retreat by minor readvances before or around 1920. The lichenometrical dating by BESCHEL (1961, pp. 1058–1059) of moraines in the continental parts of the Sukkertoppen district also indicates that there were minor readvances of the glacier lobes around 1920. The moraines of the historical maximum in the area were dated to 1850 or 1870–1890. However, BESCHEL suggested that in the more maritime western parts the maximum occurred earlier in the 18th century though the later advances had almost the same dimensions.

The contemporary literature confirms that the historical maximum in the district was around or before 1850. "*Sermilik*" (loc. 60, plate 2) and *Íkátûssaq* (loc. 48, plate 2) were reported to be thinning already as early as the end of the 19th century and the same is said also of the glaciers in Evighedsfjord in 1884 and 1885 (JENSEN 1889, p. 72). It is possible that the somewhat later date for the maximum shown by lobes in the more continental parts of the district is a result of altitude, as discussed earlier (p. 42).

Disko island. It is clear from the numerous reports of the extent of the glaciers on the island that between 1850 and 1900 most of the glacier lobes were close to, though not at, their historical maximum (*e.g.* 85B, 86-1, 88b and 95, plate 2). The maximum rate of retreat for this region seems to have occurred around 1920–1940.

The detailed information about *Lyngmarksbræen* (86-1) shows that this glacier probably last advanced between 1850 and 1870. That this advance was more widespread in its occurrence is seen from STEENSTRUP's report (1901, p. 266) of newly formed fresh glacier lobes on top of the "dead glaciers" in Vaigat. It is evident from his descriptions that the dead ice itself must have been formed before the middle of the 19th century. It therefore seems possible that many of the glaciers on the island reached their maximum in the 18th century or earlier, and that the formation of extensive dead ice was a result of the first phase of recession from the maximum.

Nûgssuaq peninsula. The northern side of this peninsula is the classical area for the study of glaciers in Greenland. The glacier on the south side of the peninsula north of the outpost Sarqaq (loc. 103, plate 2) is well known because the mountain it lies on is used as a mark for sea navigation in Disko Bugt.

In general, the glaciers of the area were near their maximum extent in the 19th century though some of the glacier lobes were thinning already in the 1890's. For all glaciers the maximum rate of retreat was in the 1930's, after a powerful readvance had taken place at some localities in the 1920's. Though the glaciers on this peninsula have a common period of retreat they nevertheless demonstrate a rich variety of behaviour and it is necessary to outline the most important cases.

Loc. 103 (*Sarqaq*). Certain imprecise information by WHYMPER (1867) indicates that the glacier reached its maximum extent before 1867, possibly around 1850. By the end of the 19th century it had begun to thin.

Loc. 108 Sermiarssuit sermikavsât (fig. 17). The outermost moraines of the glacier were described by GIESECKE. He visited it in 1811 so their formation therefore predates this. In 1811 the glacier reached to the sea and its extent was not very different from the historical maximum. This position was maintained throughout most of the 19th century, but the glacier began to thin slowly after 1880. The recession of the present century was interrupted by minor readvances in the 1910's and 1920's.

Loc. 109 Agssakait sermiat (fig. 17). It seems from the early descriptions of the glacier that in 1811 it had not yet reached its maximum extent. Nor had it at the time of the next visit in 1850 when the glacier front was still situated some distance from the beach. In 1880 STEENSTRUP observed a steep frontal lobe of clean glacier ice moving over the dead ice remnant of the former glacier. This readvance continued, and between 1880 and 1892 the glacier readvanced to the coast at approximately 100 metres annually. It is clear that the readvance which culminated around 1890–1900 was initiated around 1880. Subsequently the glacier has been retreating continuously.

Loc. 112 Sorqaup sermia ("Lille Umiartorfik glacier"). In the last half of the 19th century the glacier seems to have expanded slowly. An advance, possibly initiated in the 1860's or 1870's, culminated around 1890 with the glacier still inside the earlier historical maximum.

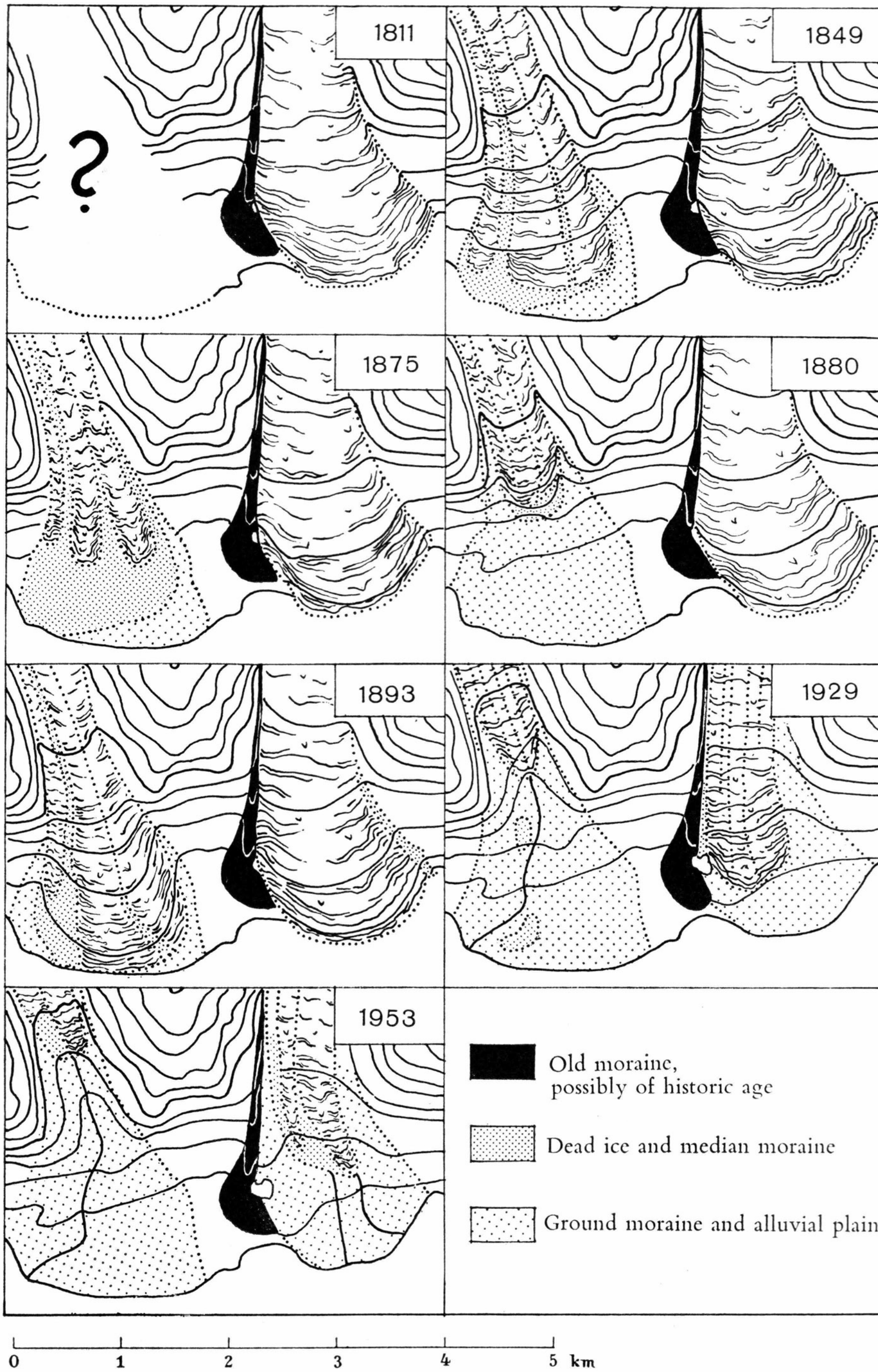


Fig. 17. Development of Sermiarssuit sermikavsât (to the right, loc. 108, plate 2) and Agssakait sermiat (to the left, loc. 109, plate 2). Map based on the Geodetic Institute's map sheet 70 V2, Umanak. By permission of the Geodetic Institute.

Loc. 113 "*Sorqaq Bræ*". Throughout the last half of the 19th century the glacier front had remained near to, but not at, the historical maximum. An advance of the glacier lobe began in the 1850's or 1860's and culminated around 1880.

It is seen from these examples that the maximum extent of the glaciers may have occurred during or before the 18th century, and that a general readvance commenced in the 1860's and culminated in 1880–1890. Only *Agssakait sermiat* deviates from this pattern with its phase of thinning around 1880.

Umanak and Upernavik districts. In these areas, also, the glacier lobes retreated most rapidly between 1920 and 1940, after a period of stability or slight recession between 1900–1920. However, at two glaciers on Upernivik Ø (119 IV and 119 VI), there were large readvances around 1930, taking them near to their maximum extent, (Møller 1959, p. 43 for 119 IV).

Earlier data, from the last half of the 19th century, are available only for the glaciers on the west coast of Upernivik Ø. Their extent was also nearly, but not completely, identical with that of the historical maximum.

In general the behavioural trends of the second order fluctuations follow those of the first order with some glaciers reaching a maximum before 1800 and others not until the last half of the 19th century.

4.3.2.3. Third order fluctuations

Fluctuations of this order were experienced by 25 of the lobes described, *i.e.* 30 % of the total number of lobes of local glaciations in plate 2. Of these, 3 are in the Julianehåb district, 17 in the Sukkertoppen district, 1 on Disko island, 3 on Nûgssuaq and 1 in the central part of the Umanak district. Glacier lobes with small fluctuations seem to belong to three categories:

a) Glacier lobes situated at high elevations which have had small fluctuations because of the altitudinal effect.
b) Glaciers with extremely low surface inclinations (4–10 % inclination).
c) Glaciers with extremely high surface inclinations (mostly 30–100 %).

Category a) includes the glacier lobes of *Napassorssuaq* (loc. 2), *Sarfartôq A and* B_1 (loc. 65) and possibly *Tunorssuaq* (loc. 85 A) and "Iviangussat" (loc. 111). The little that is known about them suggests that their fluctuations have been similar to the fluctuations of the first and second order. At loc. 65 at least, the extent of 1890 was its maximum for historical time.

Category b) covers the lobes of *Kiagtût sermiat* (loc. 6), *Umingmak* (loc. 53 IV-S), *Sarfâgfîp kugssinerssua* (loc. 106) and *Kûk* (loc. 107). These glaciers all seem to have had extents near their maximum throughout the 19th century. Since the last decades of that century they have been thinning. The best known examples are Sarfâgfîp kugssinerssua and Kûk. The outer parts of both glaciers had become stagnant at the beginning of the 19th century but at the time of STEENSTRUP's visit to them in 1879 and 1880 they were active again with steep, clean, and moraine poor fronts and had reached close to their maximum historical extent. In the 1890's these glaciers were thinning and were again partly transformed to dead ice. Thinning and dead ice formation, described by LOEWE (1935, pp. 14–16), continued throughout this century. At Kiagtût sermiat the front was fresh and steep in 1876, but in 1899 the glacier was thinning resulting in the formation of dead ice in the present century. The history of the Umingmak glacier in the Sukkertoppen district is only known back to 1902 when the front was still steep and fresh though the glacier had commenced to thin.

Category c) covers the glaciers nos. *21*, 44, 49(?), 53 III-N, *53 V-S*, 54-1, 54-2, 55-1, 57 V-N(?), 60a, 61 I, 61 II, 61 IVA, 61 V, 61 VI and 118. They are all either cirque glaciers or steep regenerated glaciers nourished by avalanches from firn plateaus. The italicised localities are extreme examples, the glaciers hanging on steep rock slopes with widths of only a few hundred metres and lengths of several kilometres. The movements of such glaciers may largely be by extended flow whereby there would be little decrease of movement with thinning of the glacier lobe. With continued thinning, these glaciers develop into different forms of avalanche fed glaciers with an abnormal recessional behaviour though, in general, thinning of these glaciers follows the trend of all other glaciers.

4.3.2.4. Local glaciations, conclusions

General agreement between the fluctuations of first, second and third order is found without regard to the glacier type or situation. The local glaciers reached their maximum historical extent before the 19th century, perhaps as early as 1750, just as did many of the lobes from the Inland Ice. Most of the local glaciation lobes kept close to this position up to the middle of the 19th century, though a few receded before the middle of that century, with the formation of large areas of dead ice. Between 1860 and 1880 there was a general reactivation or readvance of the glacier lobes. This was, in most cases, of less magnitude than older advances. Recession of the glaciers followed, with the maximum rate of retreat occurring between 1920 and 1940. Since 1940 the rate of retreat of most glaciers has slowed down. The general retreat was

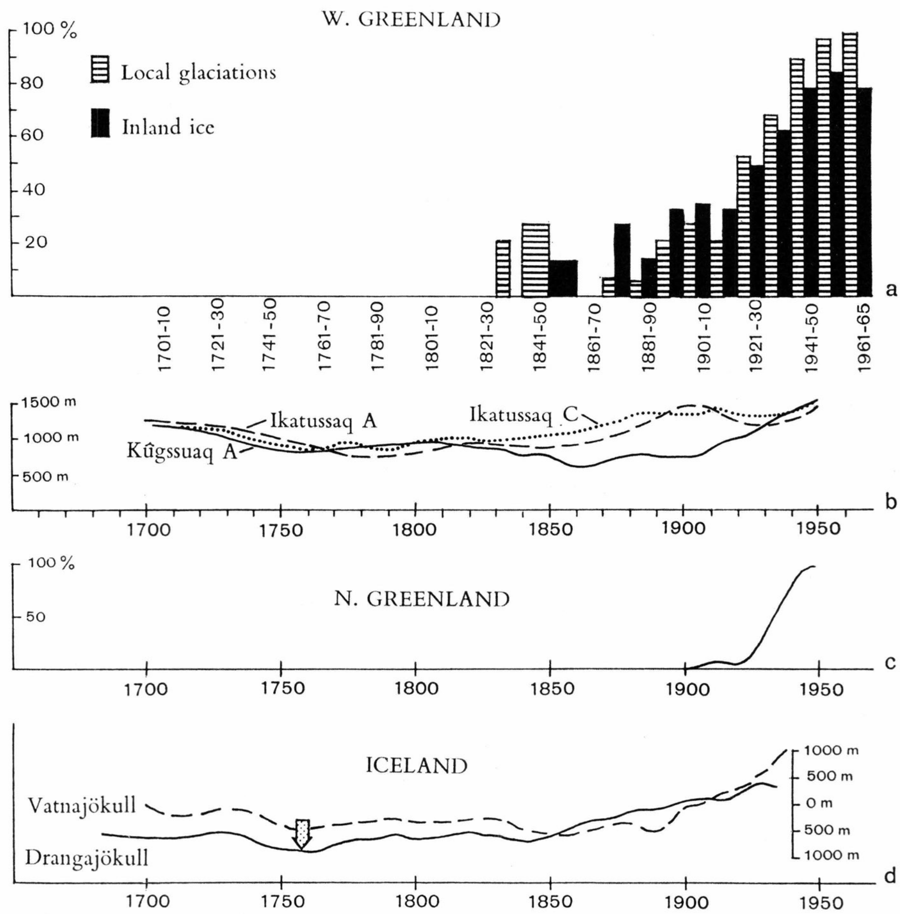


Fig. 18. Average glacier fluctuation curves. 18a. Approximate average positions for lobes from the Inland Ice (black columns) and from local glaciations (hatched columns) in West Greenland. 18b. Curves for glacier fluctuations in Sukkertoppen district, after BESCHEL (1961, p. 1058, fig. 2). 18c. Approximate average curves for lobes from North Greenland, calculated on the basis of the information published by DAVIES and KRINSLEY (1962). 18d. Approximate average curves for Iceland, after THORARINSSON (1943, p. 50, fig. 15c). Arrow: Period of maximum advance at Drangajøkull.

interrupted by short and minor readvances between 1915 and the 1920's which, as many of the curves (plate 2) show, were frequently only sufficient to stabilise the front. However, in the Umanak district this readvance had the dimensions of the older historical advances.

4.3.3. General conclusions

The fluctuations of the glaciers in West Greenland during the last few hundred years are summarised in fig. 18. In constructing the trend of fig. 18a the mean of percentage values of a_t (see fig. 8) for 10 year intervals was calculated from the values in table 2, and these decadal means expressed as percentages of the maximum historical fluctuation ($a_{t\ max}$). Thus the maximum distance a glacier has retreated becomes 100 %. The means of the percentages of all glaciers are plotted at 10 year intervals in fig. 18a. The trends so calculated are nearly identical to those of "the approximate average curves" for Iceland, given by Thorarinsson (see fig. 18d) which express the variations in terms of the length of the lobe. Comparison of the two show that the maximum fluctuation of the Icelandic glaciers is c. 2000 metres whereas for West Greenland the corresponding 100 % values is between 1000 and 2000 metres.

The occurrence of large stationary sectors, or sectors with an anomalous behaviour, is probably responsible for the fluctuation trend of the Inland Ice being lower than that for local glaciers (fig. 18a). The trend show that readvances around 1850?, 1890 and 1920 were widespread occurrences at both the Inland Ice margin and at local glaciers. Since these are the dates given by lichenometrical studies of a small number of moraines (Beschel 1961, fig. 2) it is possible that the older readvances dated by this method are equally representative of the general trend, *i.e.* that readvances around 1650 and 1750 occurred generally throughout the area. From the data that are available, it seems likely that in southernmost and most coastal West Greenland the advance of 1750 was often the maximum one in historical time.

It has already been mentioned that the summer temperature must be considered as having been the most important climatic factor responsible for glacial fluctuations during this period in Greenland. The simplest correlation of these with the known temperature fluctuations gives the following scheme. The cold period between 1807 and 1821 would be responsible for the readvances which began around 1820 and resulted in the formation of the moraines of 1840 or 1850. The advances around 1880–1900 (c. 1890) were probably the result of the cold period of 1860–1866 with a possible contribution from the short cold periods between 1880 and 1890. This readvance of c. 1890 was the last major one of widespread occurrence. The cold period of 1913–1916 was expressed by the readvances at the beginning of the middle of the 1920's. In the northernmost parts of the area investigated this advance increases in magnitude and duration, lasting there until c. 1930.

The most recent data available for the northernmost lobes of the Inland Ice (Disko Bugt and Umanak district) suggest that a new readvance might have begun around 1960. Climatological and biological evidence (Hansen 1961) indicates a tendency for temperatures since the 1940's to have been stable or even falling, and this could be its cause.

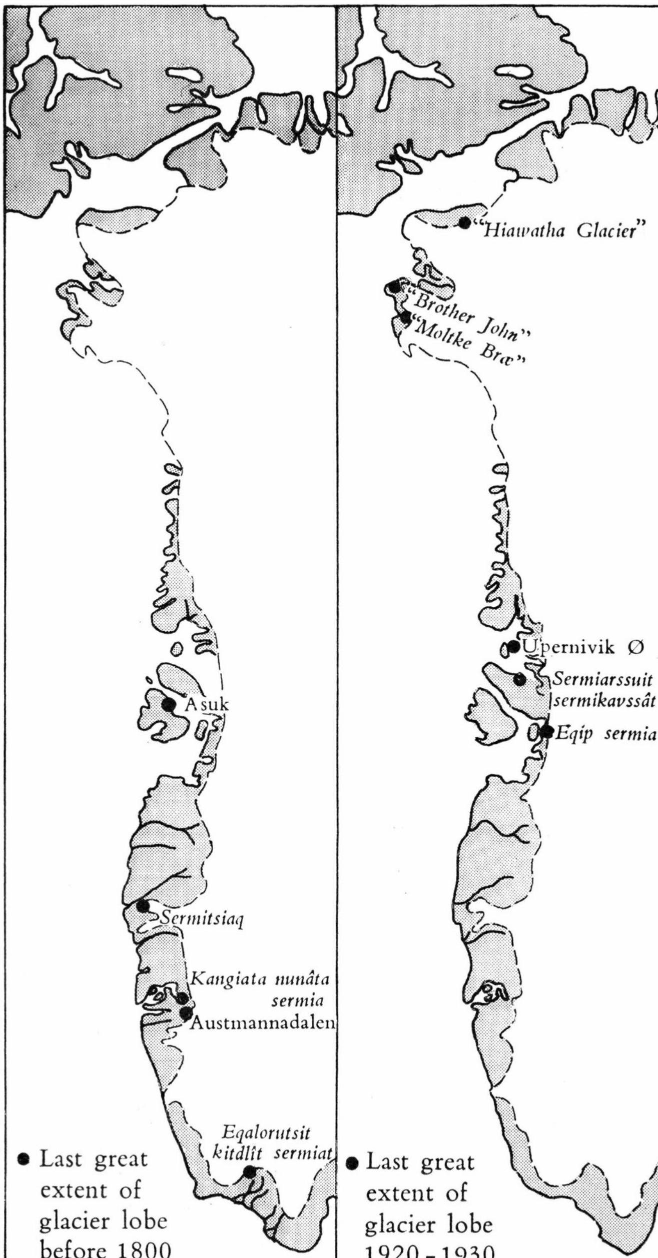


Fig. 19. Localities of glaciers referred to which had their maximum historical extent in c. 1750 or 1920–30. No significance is attached to difference in letter styles.

In addition, recent temperature determinations by Dansgaard (Dansgaard *et al.* in print) in firn strata from the Inland Ice between Umanak district and Station Centrale (Eismitte) which cover the period since the 1930's indicate a corresponding decline in temperatures in the central parts of the Inland Ice.

If this, the simplest correlation between climate change and glacier response, is correct the delay operating would seem to be from a few years to about 20 years, regardless of the type and magnitude of the glacier lobe.

The recognition of the increasing importance towards the north of the 1920–30 readvance in West Greenland agrees with Davies and Krinsley's (1962) investigations in North Greenland. Their data is summarised here in figs. 18 and 19.

In Peary Land, the small local glaciers and the Inland Ice margin seem at present to be stationary.

From East Greenland there is little information available. The summary given by Sharp (1956 see also Ahlmann 1948), suggests that the maximum extent of the glaciers occurred in the middle of the 18th and 19th century just as in many places in coastal and southernmost West Greenland. However, in East Greenland a later readvance around 1890 also seems to have had an extent near to that of the previous advances. Gribbon (1964, pp. 361–363) has dated the commencement of recession of *Tasîssârssik A* glacier near Angmagssalik from its maximum position to c. 1830 ± 20 years. This date is in agreement with that derived from the literary evidence for glacier *1a*, *Sermítsiaq*, in Tasermiut fjord, in the Julianehåb district. That many glaciers in East Greenland have been retreating strongly during this century has been stated by several investigators (see Sharp 1956).

4.3.4. Comparisons with surrounding areas outside Greenland

The areas discussed are shown in fig. 20.

1) Axel Heiberg Land. The present conditions of the glaciers here seem to be similar to those in Peary Land, with glaciers stagnant in a position very near to the historical maximum, or even advancing in most recent times (Müller 1962, p. 142).

2) Baffin Island. Falconer (1962) has summarised from the literary and photographic evidence the fluctuations of glaciers at 59 localities from the northernmost part of the island (Bylot area). Only small valley glaciers and hanging ice tongues have markedly retreated during the last decades.

Ward and Hale (1952) state that the southern margin of Barnes Ice Cap is nearly stable or slightly retreating, though Ives (1962), from evidence at the northern part of the same ice cap, concludes that it had

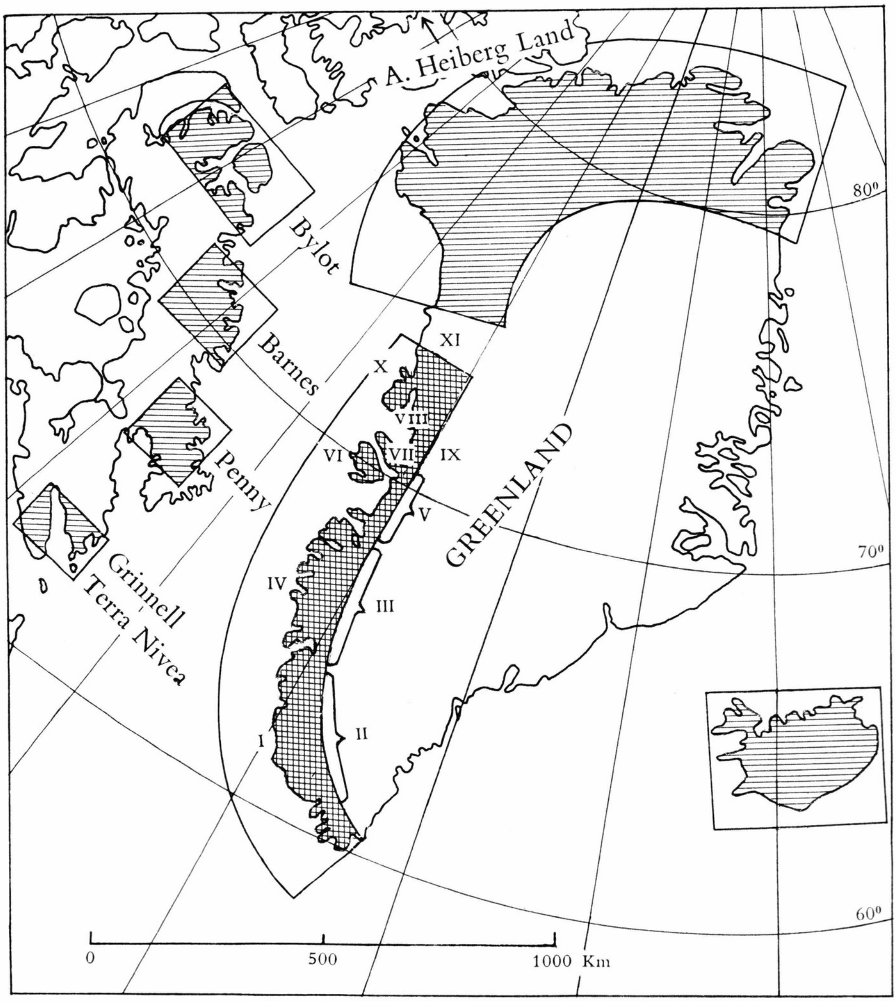


Fig. 20. Areas whose glacier fluctuations are mentioned in the text. Cross-hatched area: West Greenland with the position of the districts listed in table 2 (pp. 162–187) indicated by the roman numerals. Horizontal hatching: Areas whose glacier fluctuations are compared to the West Greenland ones.

a significantly greater extent 300–400 years ago. At the Penny Ice Cap, geomorphological (THOMPSON 1953) and botanical (SCHWARTZENBACH 1953) investigations at Pangnirtung indicate there was a general expansion of glaciers here, 150–200 years ago, followed by thinning, which began before 1900 and is still continuing (WARD and BAIRD 1954). The southernmost ice caps on the island, Grinnell and Terra Nivea, have receded during this century (MERCER 1956).

3) Iceland. The most complete record outside the Alps of fluctuations during the historical period is from Iceland. This is compiled by THO-

RARINSSON (1943) and recorded here in fig. 18d. Both Drangajøkull and Vatnajøkull readvanced in 1750 and 1850 but at Drangajøkull it was the first readvance which brought it to its historical maximum position whilst at Vatnajøkull it was the later. Subsequent readvances around 1890 and 1920 in both areas were of considerably smaller extent than these. In their type of reaction, the Icelandic glaciers show a close similarity to the few known from East Greenland and to those from southwest Greenland. However, it is apparent that even in West Greenland the readvance of c. 1890 was more significant than in Iceland.

These brief descriptions show that the regional variations in the fluctuations of the glacier lobes in Greenland are part of a more widespread trend related to a zonal variation of the activity index. High-polar glaciers, with low activity indexes in North Greenland, in Axel Heiberg Island and in the northernmost parts of Baffin Island mostly have small and late fluctuations. For North Greenland, the small magnitude of the fluctuations in general, and the particular importance of the maximum readvance in the 1920's has been explained by the increasing importance of variations in precipitation towards the north (DAVIES and KRINSLEY 1962, p. 128).

Temperate glaciers belong to another zone comprising southwest Greenland, Iceland, parts of East Greenland, Spitzbergen and possibly Jan Mayen (FITCH *et al.* 1962, p. 209). They have higher activity indexes and show a faster reaction to climatic fluctuations.

The tendency for the high- polar glaciers to reach their maximum extent in the last decades, whilst the maximum extent of the mostly temperate glaciers in another zone was reached around 1750 and 1850, is part of this zonal variation. It agrees with the zonal trend of glacier fluctuations, calculated by HAEFELI (1962, p. 53) and SHUMSKII *et al.* (1964, p. 449, fig. 4), which demonstrate that decrease in activity index results in a decrease in rate of response to climatic fluctuations.

4.4. Glacier fluctuations in area and volume

The difficulty of determining these has already been mentioned. However, the information available about frontal fluctuations suggests that by far the greatest decrease in thickness of glaciers has occurred since 1900. Whilst this is especially true for continental and high arctic glaciers, even for the majority of local glaciers near the coast, thinning before 1900 has formed a relatively small part of the total thinning. The exceptions to this have been discussed in the foregoing pages.

Accordingly, the trim line zone in West Greenland represents an area which in over 90 % of the cases has been once covered at the end of the last century. The mapping of the extent of the trim line zone must therefore be important for the estimation of the total area glaciated and deglaciated and especially for the estimation of the volume of ice lost. However, to do this for the Inland Ice involves a source of error of an uncertain magnitude, since it is not known how much land has been buried beneath sectors which have advanced since 1900.

THORARINSSON has suggested that neither the Inland Ice nor the Antarctic ice sheet has taken essential part in the worldwide thinning of glaciers in this century, since the measured eustatic rise of sea level has only been < 1 cm per 10 years, a figure which he suggests is too small to allow for much change in volume of the two ice sheets (THORARINSSON 1940, p. 150). Further uncertainty has arisen because of the wide disparity between the estimates of the mass balance of the Inland Ice by different authors (LOEWE 1936, BAUER 1955b, BENSON 1959, 1961, 1962 and BADER 1961), the chief reason for the difference being the uncertainty about the estimates of iceberg production and other ablation losses. Whilst LOEWE's calculation indicates, like BENSON's, that the Inland Ice budget is nearly in balance, BADER calculated an annual gain in ice between 120 and 270 km^3, and BAUER an annual loss of c. 100 km^3. However, it is possible to give an estimate of the thinning which has taken place in the marginal zones in West Greenland by plotting the vertical separation of the trim line zone from the present ice surface (Δh_H) as a function of the altitude of the present ice surface (h). The areas where the trim line zone has been investigated over relatively long distances, *i.e.* northern part of the Julianehåb district, Jakobshavns Isbræ and Nûgssuaq are those where the fluctuations of the ice margin have been found to reflect fluctuations in the total volume of the lobes. Presumably in these cases, the successions of the marginal moraines are the same as those of the frontal moraines. Ignoring the anomalous advances of certain sectors of the ice margin in this century, on the assumption that they affect an insignificant part of it, a random survey of the function $\Delta h_H = f(h)$ along the Inland Ice margin gives a result which is about the same as that for local glaciers, if the magnitude of the frontal fluctuations has been the same in the two cases.

The height of the trim line zone has been measured for local glaciers in the Julianehåb district, the Sukkertoppen district and at Blæsedalen, Godhavn area, Disko island, and the results compared with similar measurements from lobes along the Inland Ice margin, from the Julianehåb district in the south to Nûgssuaq peninsula in the north. The results have been listed in table 3, pp. 188–189. Whilst the maps used for the local glaciers in the Julianehåb and Sukkertoppen districts and for the Inland Ice lobes are the Geodetic Institute's map sheets 1:250,000 with a 50 m contour interval, a new special purpose map at 1:20,000 with a 20 m contour

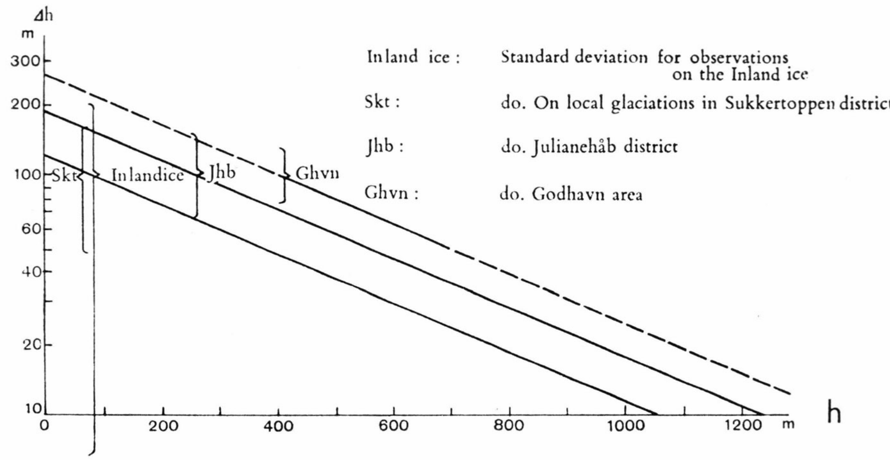


Fig. 21. Δh_H as a function of h.

interval exists for Blæsedalen. For each glacier lobe, the height of Δh_H was found for each 100 m interval of h and the standard deviation calculated for the sum of each 100 m interval within the four groups.

The results of the investigations are shown in fig. 21. These curves are expressed by the general formula,

$$\Delta h = z\, e^{kh}$$

where z and k are constants and e the base for the natural logarithm. In the particular case shown (for Δh_H), k_H is $\frac{-2.303}{1000}$ and z varies from 120 for the Inland Ice and the local glaciations in the Sukkertoppen district, to 180 for the Julianehåb district and 260 for Blæsedalen. The high value of z for Blæsedalen may be due to the small size of the sample area which would emphasize any local difference due to local climatic variation, quite apart from the fact that the curve is extrapolated from data from a narrow altitudinal range, between 400 and 800 m a.s.l.

The mean of the three curves for local glaciations lies close to that for Julianehåb, which area can thus be assumed as representative of the coastal part of Greenland. The curve for the Inland Ice lies below this, demonstrating the smaller amount of thinning which has taken place there.

The mean total thinning (S) of ice in historical time is determined by:

$$S \cong \frac{z \int_{h=0}^{h=1200} e^{\frac{-2.303}{1000} h}\, dh}{1200} \text{ metres}$$

$$\cong \frac{z}{3} \text{ metres.}$$

Thus for the Julianehåb district, the mean total thinning of the local glacier lobes is c. 60 m ice and for the Inland Ice c. 40 m ice. Since the main period of retreat, and hence of most of the thinning of the lobes has been within this century, the mean annual loss of the ablation area is seen to be approximately 1 m of ice for local glaciations and approximately 0.7 m of ice for the Inland Ice margin. These figures are close to those calculated by THORARINSSON (1940, p. 149) for ablation areas in other parts of the world, *i.e.* 0.5–0.9 m of water annually.

With an average of c. 0.7 m of ice lost annually over the ablation area of the Inland Ice, the total annual loss from the Inland Ice margin, assuming that the whole marginal area in Greenland has the same ablation conditions as West Greenland, would be:

$$286{,}600 \text{ km}^2 \times 0.0007 \text{ km} = 200 \text{ km}^3 \text{ per year.}$$

The figure for the area of ablation is taken from BAUER's calculation of the hypsographic curve of the Inland Ice (BAUER and HOLZSCHERER 1954, p. 35, fig. 23 and p. 39). The loss due to retreat of the margin is not included in the calculation and it is presumed that this, to some degree, compensates for the overestimate due to the thinning in North Greenland being less than the figure derived for West Greenland.

Whether the central parts of the Inland Ice are gaining or losing mass is still a controversial point. It is the opinion of BADER (1961, p. 9) that in the northern half of the Inland Ice ablation in the ablation zone is less than the accumulation in the accumulation zone and that the area has a positive budget and is gaining in mass. SHUMSKII (1965) from the geodetic measurements of the Expédition Glaciologique Internationale au Groenland 1959–1960 expeditions, calculated a loss near the central part of the southern dome of the Inland Ice (Station Centrale, Eismitte) of 50 g cm^{-2} annually, equalling 900 km^3 year^{-1}. He adds, that the result does not agree with the known present eustatic rise of sea level of 2.4 mm per year, but that the relationship may be partly invalidated by, for example, sinking of the bottoms of the oceans.

4.5. Surface characteristic

The deposits of the glacier readvances in historical time, as a whole, describe a single stage or "stade" in the development of the ice margin. The deposits of this unit are concentrated essentially in the outer half of the trim line zone, being the ice marginal deposits formed between c. 1600 and c. 1920. These have a vertical and horizontal spread which decreases with altitude (fig. 21). Thus the vertical spread of the Inland Ice margin and of local glaciers in the Julianehåb district decreases from values of c. 100 m where the mean maximum horizontal spread is 1000 to 2000 m.

In the event that these ice margin deposits of 1600 to 1920 had not been marked by their distinctive lack of weathering and their special pioneer vegetation, the principle criteria for their recognition as a unit would be the magnitude of their degree of development, and their zonal distribution parallel to the present ice margin. This last character is a function of the relationship of the past ice surface to the present, a relationship which is capable of definition by the equation given on p. 62. For the sake of convenience z (Δh for $h = 0$, fig. 21) is used to describe this character, and is defined as the "surface characteristic" of the zone of deposits. However, for a full description of the ice surface of a zone all terms in the equation have to be included (this concept is used in the later discussion of the deposits of older readvances). For the historical readvance of the Inland Ice, the surface characteristic z_H is typically 120m, within the limits shown in fig. 21.

4.6. Historical isostatic and eustatic fluctuations

Information compiled by BØGVAD (1940, pp. 28–29) indicates that West and East Greenland began to sink in the 17th century. This sinking continued until around the 1940's, since uplift has been recorded at several localities in both West and East Greenland (SAXOV 1958b, pp. 520–521, 1961, p. 413). In Disko Bugt the present annual uplift is considered to be c. 17 mm. A measurement by the author in 1961 of the altitude of a point in Qutdligssat, Disko island, erected in 1898 by PORSILD (STEENSTRUP 1901, pp. 267–268, PORSILD 1902, p. 121), suggested a rise of 0.6 m in this period. Though this value fits into SAXOV's curve for the sinking and subsequent uplift of the area it must be pointed out that the datum for both measurements was the upper limit of marine algae, for which there is an estimated variation of ±20 cm. However, in the south, at Igaliko in the Julianehåb district, STEENSTRUP (1881, p. 40) claims there has been no observable change in the shore line throughout the 19th century.

The probable explanation for these movements is that they are isostatic, related to the development of glaciation. The indications of early subsidence correlate with the early phases of readvance considered to have taken place in the 16th century. However, in general, subsidence seems to have continued for some decades after the initiation of retreat. The main loss of ice was between 1920 and 1940 whilst the commencement of uplift is put at about 1940, implying the existence of a time-lag of this dimension between the two phenomena. The late phase of subsidence is too large to be explicable by the eustatic changes in this period which are estimated to be less than 1–3 mm per year (GUTENBERG 1941, p. 731, SHUMSKII 1965, p. 322).

However, there is not enough information available yet for a detailed analysis of the problem. For its further study an island such as Disko would be an ideal place because of the existence of SAXOV's early measurements and of the supposed relatively high activity index of the glaciers there, which has resulted in large variations in the extent of the glaciers during this period.

5. TYPES OF ICE MARGIN DEPOSITS IN THE AREA DEGLACIATED IN HISTORICAL TIME

5.1. Shear moraines and dead ice formation

Deposits on the surface of the ice margin owe their origin largely to the operation of shear planes in the ice. An exception to this is the small amount of material transported on the ice surface from its source, as surface moraines. For the Inland Ice this is of minor importance.

The potential importance of shear planes for the transport of moraines was long ago pointed out by PHILLIP (1920), GRIPP (1929) and in Denmark by SIGURD HANSEN (1932). However, as an explanation of glacier movement proper, the shear plane theory was superseded by the theories of viscous flow, such as those given by LAGALLY (1930) and KOECHLIN (1944) which dealt with laminar flow in a body of constant viscosity.

The idea of a constant viscosity was later disproved by experimental investigations on the flow of polycrystalline ice (GLEN 1958). GLEN'S law for the flow of ice is stated by:

$$\dot{\gamma} = c\tau^{n}$$

where $\dot{\gamma}$ is the shear strain rate, c an empirically determined constant dependent on temperature, τ the shear stress and n a number between 3 and 4. Thus it can be seen that the apparent viscosity of the ice decreases with the increasing stress. As a simplification of the law, NYE (1951) assumed that the ice is an ideal plastic substance, behaving elastically below a certain limit and plastically above it. From the simplified law, NYE (1951, pp. 558–559, 1952a, p. 88) calculated the thickness of an upper "fragile zone" of a glacier, stiff enough for the formation of crevasses. It is presumed that the occurrence of shear planes is mainly within this tensile (fragile) zone and its maximum depth indicates a limit for their development. According to NYE the thickness of the zone (t) is:

$$t = \frac{k}{\rho g \sin \alpha} = \frac{1130 \text{ cm}}{\sin \alpha}$$

where k is the yield stress (10^6 dyne cm^{-2}), ρ the density of the ice (0.9 g cm^{-3}), g = 981 cm sec^{-1} and α the inclination of the glacier surface.

An attempt has been made to divide glacier lobes into two categories; those with and those without dead ice. The basis for the division is the present condition of the glaciers, with the presence of dead ice

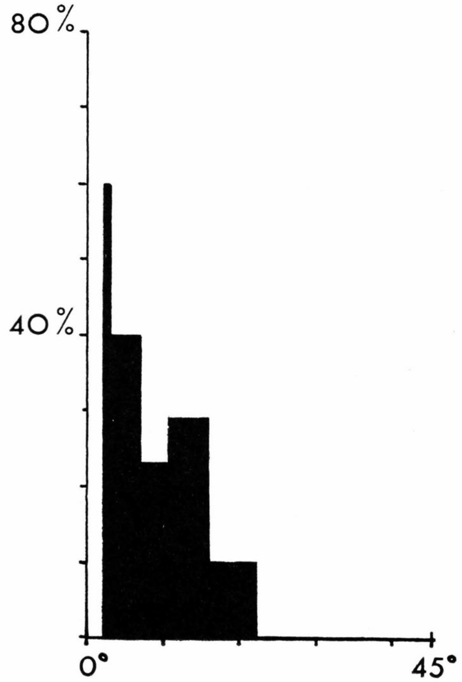


Fig. 22. Percentage of glaciers with dead ice (ordinate), grouped after their surface inclination (abscissae).

being deduced from the appearance on the aerial photographs of a moraine-covered lobe with dead ice topography. It is clear that transitional types are met with, but these have not been taken into consideration. 73 of the 135 glaciers listed in table 2 have been used in the division.

These 73 glaciers have been grouped in fig. 22 according to their surface inclinations; 2.0°–2.9°, 3.0°–6.8°, 6.9°–10.3°, 10.4°–15.5°, 15.6°–21.8° and greater than 21.9° as given in table 4, p. 190. The result (fig. 22) suggests that the probability of the formation of dead ice increases strongly with decrease in inclination of the glacier surface. This result is in spite of the fact that the morphological criterion for dead ice does not necessarily fit with the mechanical criterion that dead ice is without significant movement, and that all glaciers have been grouped together regardless of their thickness.

The formation of dead ice in the Thule area by the development of shear moraines has been described by BISHOP (1957, p. 18). He presumed that the outer parts of a glacier lobe with an evenly inclining surface will cease all movement when its thickness decreases to a certain critical value, which supposedly is equal to the thickness of the upper fragile zone of the glacier. At the limit between distal dead ice and proximal active ice shear planes will be developed which transport moraine material to the surface of the glacier. BISHOP stated the critical thickness to be 65–80 m. The inclination of the ice at this point is 3°–7°, and its actual thickness is less than given by NYE's formula. This difference may be due to the existence of a proglacial snow bank at the ice margin, or to the various errors in the identification of the limit between dead ice and active ice. Furthermore, nearly dead ice follows other laws of flow than active ice. It is believed (MELLOR 1964, p. 78) that "moderate concentrations" of dirt in polycrystalline ice increase its mobility, whilst "higher concentrations" will produce a substance more quasiviscous than the pure polycrystalline ice (SWINZOW 1962, p. 226). However, according to TSYTOVITCH and SHUMSKII, the viscosity of polycrystalline ice is 10^{12}–10^{13} poise and of frozen sand or loam $1–2 \times 10^{12}$ poise (SHUMSKII 1964, p. 102), between which there is no great difference.

The theory of BISHOP explains the way in which moraine material could pass into shear planes in the glacier ice and be transported to the surface, but WEERTMANN (1961, pp. 965, 967) raises the objections that it does not explain the fine dispersion of silt in the dirt bands of the ice and that in high-polar glaciers there is essentially no sliding over their base. WEERTMANN therefore presumed that the enclosure of material takes place at the base of the glacier by a freeze-thaw process.

That a freeze-thaw process also operates to some degree at the base of glaciers in West Greenland is suggested from Alángordliup sermia lobe in Disko Bugt by the presence of fragments of shells along ice crystal boundaries in the lobe. However, in general the theory of BISHOP seems sufficient to explain the development of dead ice at the ice margin.

Shear moraines are characteristically associated with frontal transversal shear and it is here that the greatest development of dead ice is found. However, examples of the formation of dead ice by marginal longitudinal shear do occur, *e.g.* Mellemfjord in the island of Disko and

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Fig. 23a. Marginal longitudinal shear moraines at Sarfâgfîp kugssinerssua (loc. 106, plate 2), Nûgssuaq, Geodetic Institute's route A76/141, no. 37 (5.7.1953). Fig. 23b. Frontal transversal shear moraines at the Inland Ice margin south of Søndre Strømfjord. Geodetic Institute's route 207V, no. 4772 (8.8.1952). Copyright Geodetic Institute.

Fig. 23 a.

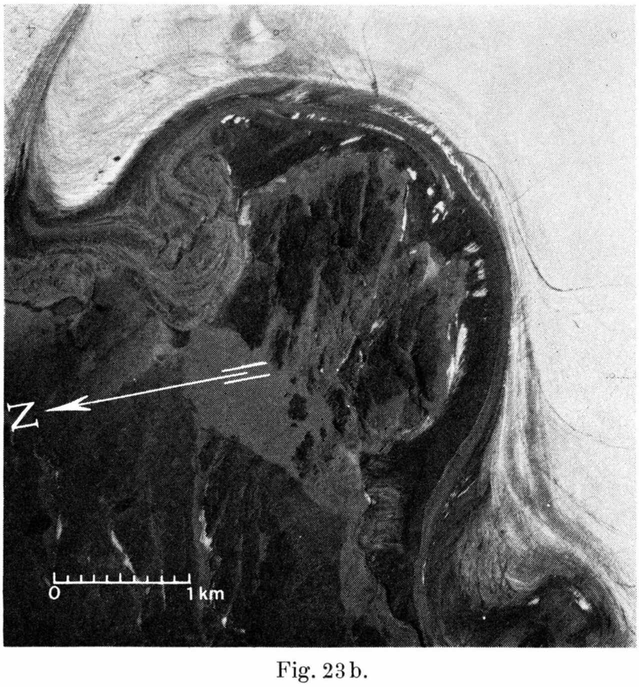

Fig. 23 b.

Sarfâgfîp kugssinerssua (see fig. 23) and Kûk glaciers on the north side of Nûgssuaq peninsula. In these last two glaciers active ice is embedded between very wide marginal and medial moraines with dead ice cores, and longitudinal shear presumably occurs between the dead and living ice. In both circumstances the formation of dead ice is dependent initially on the thinning of the glacier margin to a critical thickness for a specific surface inclination, below which the glacier margin loses most of its movement.

From the few examples available it seems that the reactivation of a glacier lobe which has formed dead ice by frontal, transverse shear, is by the formation of a new lobe which advances over the old dead ice whereas for a glacier lobe with dead ice due to marginal, longitudinal shear, reactivation of the dead ice itself can occur. The first step in this case will be the lateral expansion of all or parts of the inset living tongues in the glacier lobe leading to the direct incorporation of dead ice, or less active debris loaded ice, in the advancing glacier.

Frontal shear moraine deposits of moderately great extent have been visited by the author in Disko Bugt (Pâkitsoq and Jakobshavn area) and in the northern part of the Julianehåb district. In these areas the superficial moraine is generally formed of angular boulders, though rounded boulders and silt concretions with Quaternary fossils also occur. In this type of shear moraine deposit the finer fractions are more frequent than in boulder moraines favouring the preservation of ice under the moraine. It is probable that only a thin veneer of moraine is left when the dead ice has disappeared, forming a "drift border". Although less conspicuous than marginal moraines or terraces this can be an equally good indicator of former positions of ice margins.

5.2. Rock glaciers

The formation of dead ice and the development of surface moraines under certain circumstances is related to the formation of rock glaciers. These are especially frequent in the northern parts of the area examined, in Disko Bugt and Umanak district, occurring generally below steep rock faces. The southernmost example known from a coastal locality

Fig. 24. Rock glaciers and talus formation in Qôrorssuaq valley, south coast of Nûgssuaq peninsula near Torssukåtak. 24a. Rock fall on east side of valley, top of cliff 700–800 m and bottom at c. 100 m a.s.l. Photograph, 18.8.1961, Weidick. 24b. Sketch of rock glaciers in the valley. Angle marks the place from which fig. 24a is seen. Map based on the Geodetic Institute's 1:250,000 map sheet 70 V 2, Umanak. 24c. Aerial photograph of Qôrorssuaq. Route A 96/150, no. 75 (20.7.1953). Copyright Geodetic Institute. By permission of the Geodetic Institute.

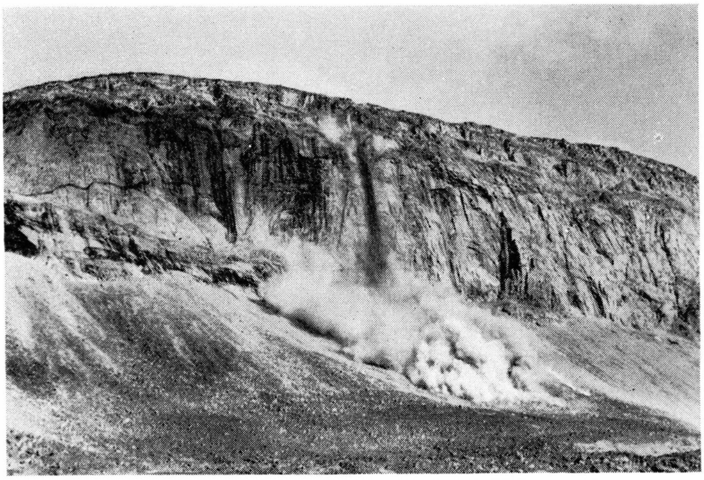

Fig. 24 a.

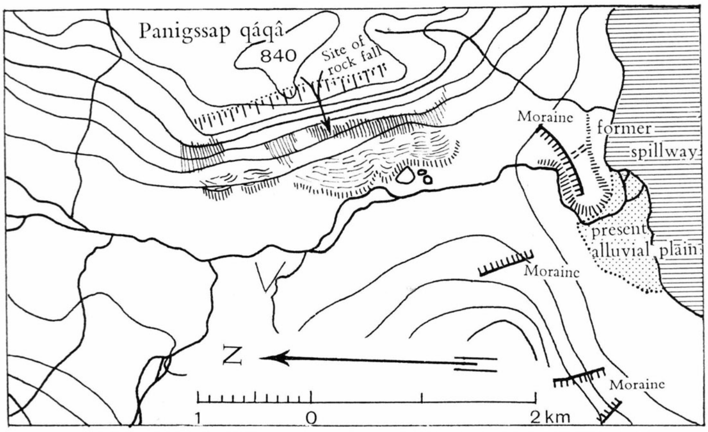

Fig. 24 b.

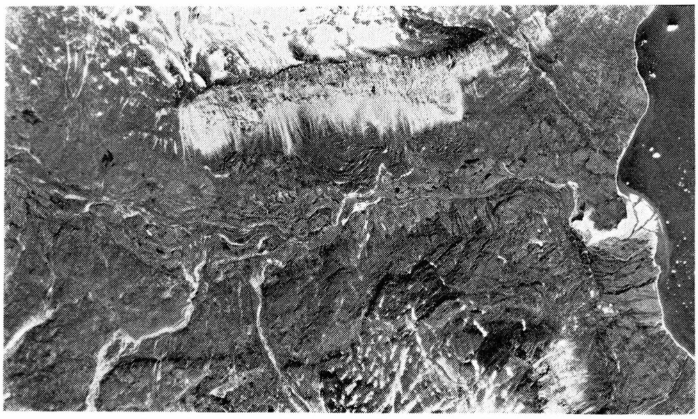

Fig. 24 c.

is that on the south side of Kællingehætten ridge near Holsteinsborg. All morphological transitions from moraine-like forms to genuine rock glaciers can be found, but typically in the area they are formed of coarse angular boulders with little interstitial material, and have ogive-like pressure ridges on their surface.

From observations made in Umanak district, BARTON (1897, p. 235) considered that there was a genetic relationship between moraines and rock glaciers, though DRYGALSKI (1897, pp. 150–153), after visiting the same localities, pointed out a genetic connection between them and rock falls and talus.

STEENSTRUP (1883a, p. 81) studied rock glaciers on Umanak island and on an old terminal moraine at Sermiarssuit sermikavsât (fig. 17) and concluded that both were dead ice remnants. Because of the presence of water bodies on the tops of the deposits of Sermiarssuit sermikavsât he maintained that they still contained dead ice.

LOEWE (1935, p. 13) concluded from a study of the dead ice remnants in front of Agssakait sermiat (see here fig. 17) left by the advance of c. 1890, that melting was virtually prevented by a cover of 1.5 m moraine with a climate as then, *i.e.* an annual temperature at sea level of c. $-6°C$, and an annual temperature range of 30°C and a thermal conductivity for the moraine of 0.8 cm^2 min^{-1}. In addition to this calculation there are ØSTREM's (1961, pp. 418–419) observations in Norway which indicate that ice cores in moraines can exist under favourable conditions for several thousands of years. However, the long term existence of relict ice is dependent on low negative annual temperature, presumably similar to that required for permafrost. As seen from the map in fig. 25, the requisite conditions for the latter seem to exist north of, or altitudinally above, an annual isotherm of $-4°$ to $-5°$. This means that even in the southern part of Greenland these conditions exist some 100 metres above sea level.

At rock glaciers visited in Blæsedalen on Disko island, Qôrorssuaq (see fig. 24) and Saputit on the south coast of Nûgssuaq peninsula, and also at the moraines in front of Sermiarssuit sermikavsât studied by STEENSTRUP, a dense cover of lichens was found on the boulders, showing that there has been no internal movement of these for several centuries. It was also noticed that fresh talus lay on top of some rock glaciers and that this was not developing pressure ridges (BESCHEL and WEIDICK, in press).

It is suggested that the development of rock glaciers took place in the following way:

Phase 1) Formation of extensive marginal moraines of coarse angular blocks by a valley glacier eroding easily weathered bedrock.

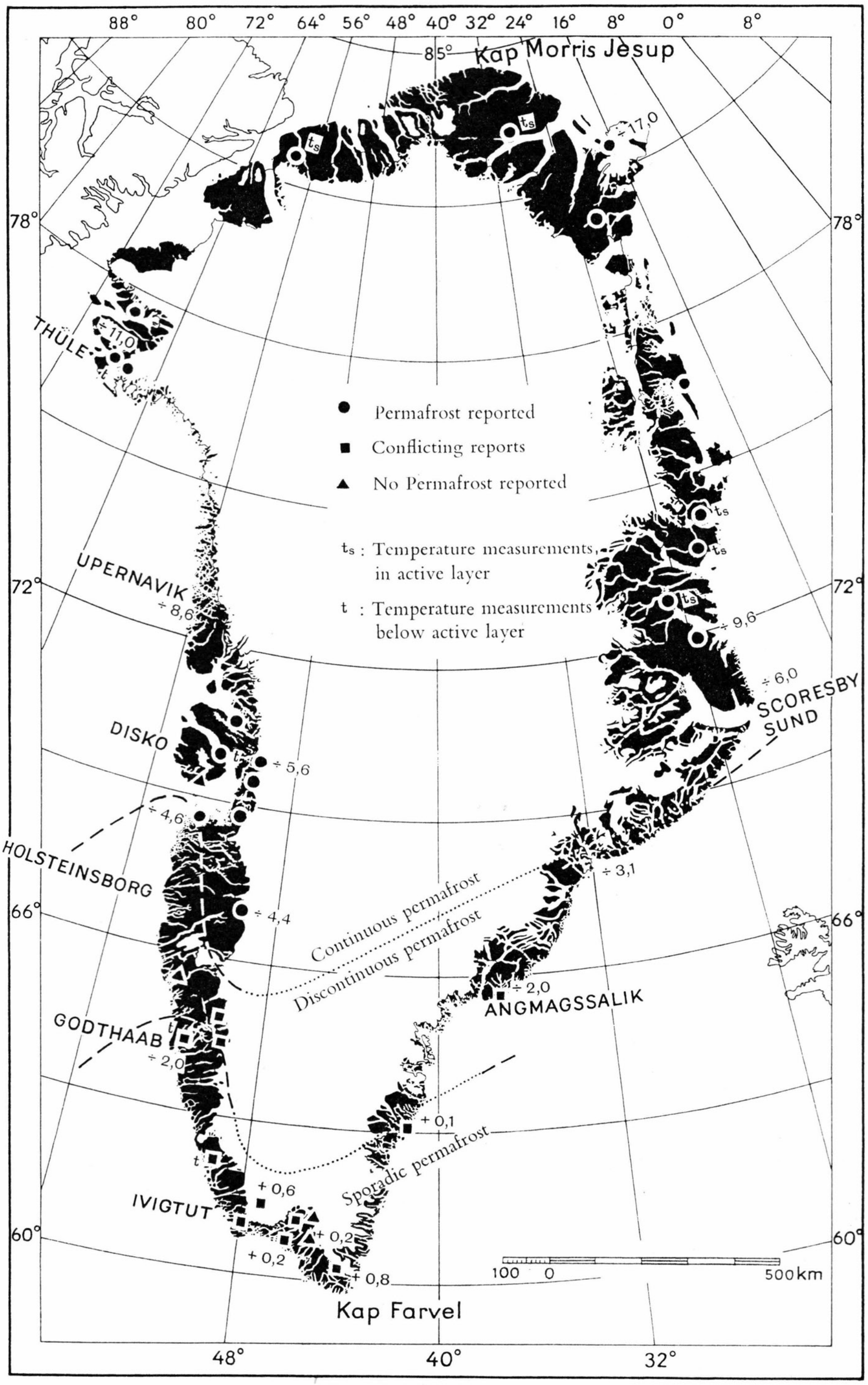

Fig. 25. Extent of permafrost in Greenland. Data based on S. Hansen (1952) and reports of G.T.O. (Greenland Technical Organisation, Ministery of Greenland). Values shown are mean annual air temperatures.

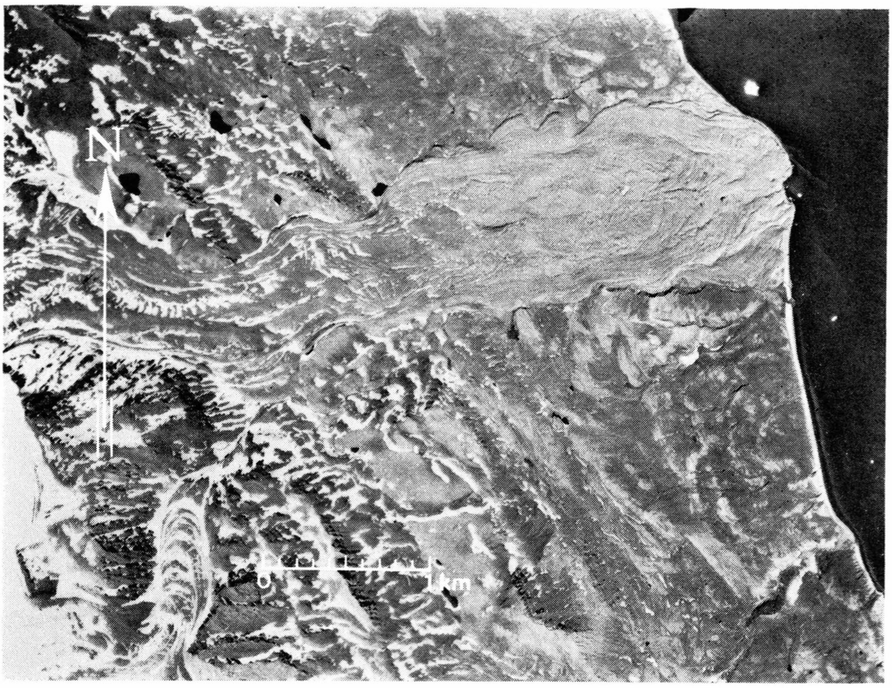

Fig. 26. Soil stream near Ûnartoq at Ujarasugssuk, Disko island. Geodetic Institute's route A80/169, no. 85. (6.7.1953). Copyright Geodetic Institute.

Phase 2) Thinning of the glacier margin resulting in the formation of zones of dead ice at the sides by longitudinal marginal shear. In these zones movement in a down valley direction ceases but an incipient creep down slope develops.

Phase 3) Disappearance of the glacier and downslope creep of the dead ice remnants with the formation of pressure ridges on their surfaces.

Phase 4) The disappearance of interstitial ice from between the boulders resulting in cessation of creep. From then on rock falls may add material to the surface but commonly do not lead to movement of the rock glacier. It is thought that the disappearance of the ice may take place more by evaporation than by melting. In the cases where the deposits are well-sorted large angular boulders, evaporation would be facilitated by the large pore spaces.

This scheme for their development is very approximate. It seems likely that in some cases ice becomes covered in time with a sufficiently thick layer of dust and fine fragments weathered from the boulders to effectively insulate the ice, prolonging the period of movement of the rock glacier. However, this has not been observed in the area.

Phenomena distinct from rock glaciers are the soil streams or mud flows, which are often seen in basalt terrain, especially on Disko island. Their source seems to be small glacier remnants in cirques, or perennial snow patches. Those examined in Blæsedalen, Disko island, and presumably most of the other examples seen on aerial photographs, have a high content of fine ash, which when moist has a low viscosity. An active mud flow at the locality shown in fig. 26 was observed by STEENSTRUP (1901, p. 280) to be capable of carrying large boulders, but not of being crossed by an adult man.

5.3. Marginal deposits

The most important marginal deposits in the context of this work are those indicating the position of former ice margins, *i.e.* marginal moraine ridges and moraine and fluvioglacial terraces.

The distinction by CHAMBERLIN (1894, p. 525) between dump moraines, submarginal moraines and push moraines can be used in Greenland only with much modification. The only criterion distinguishing depositional moraines is the absence of thrust features which the coarse texture of moraines in West Greenland makes it difficult to detect anyway.

Push moraines have been observed at Påkitsup ilordlia, Disko Bugt, where they were formed during the advance between 1850 and 1880 (see fig. 9, pp. 32–33, and at the present margin of the expanding Eqalorutsit kangigdlît sermiat glacier in the Julianehåb district (WEIDICK 1959, p. 61). In both cases the push element in their formation is shown by the incorporation of turf or lichen covered boulders in the fresh moraines.

Another occurrence of push moraines would seem to be the annual moraines observed in several places. At "Søndre Qôrnoq glacier", Frederikshåb district (WEIDICK 1959, p. 85, and here fig. 8) the extent of the glacier is known in 1903, 1938 and 1948. Between the positions of the ice margin in 1903 and 1938, approximately 20 moraine ridges can be counted, whilst between the positions of 1938 and 1948 there are approximately 10 ridges. The material in these moraines is fluviatile gravel of pebbles and cobbles with occasional silt concretions.

It is possible that most moraines at low altitudes in West Greenland should be considered as being primarily formed as push moraines. CHARLESWORTH (1957, I, p. 411) has said that there is a lack of push moraines in Iceland and Greenland, explaining this as being because "ice-free territories with abundant unconsolidated detritus are rare and the crystalline rocks are less favourable than Spitzbergen's Mesozoic

Fig. 27. Interlobate recent and prehistoric moraine landscape, interior part of Nûgssuaq peninsula. The locality is near the small nunatak to the right of fig. 30a. Photograph 12.8.1961, WEIDICK.

a

a

Fig. 28. Nunatak moraine. "Hammer's nunatak" (a–a), Qíngua kujatdleq in Pâkitsup ilordlia (loc. 72 in plate 2, cf. fig. 9). Photograph 23.7.1961, WEIDICK.

Fig. 29. Marginal terrace in front of Nordenskiölds Gletscher (loc. 39, plate 2). Aerial photograph, Geodetic Institute's route 512 C–S, no. 52 (19.8.1948). Copyright Geodetic Institute.

and Tertiary Strata". However, it seems that a lack of observations is also a factor.

At high altitudes the moraines are principally interlobate and lack well defined ridges being most often a wide zone of small moraine hills (see fig. 27).

In places nunatak moraines are present encircling isolated mountains bounded by slopes of moderate inclination. The moraines have either the form of interlobate moraine or of marginal moraine ridges such as shown in fig. 28. The latter type are best developed on the push side of the hill.

Marginal terraces are common along the Inland Ice margin and occurrences have been visited at Qaleragdlît imâ (WEIDICK 1963b, p. 88) in the Julianehåb district, Sermiligârssuk in the Frederikshåb district and Alangordlia-Sermilik and Ujaragssuit pâvat in the Godthåb district. Other examples have been seen on aerial photographs and one of these showing a frontal marginal terrace being formed at present is shown in fig. 29. Most of these terraces have small push moraines of fluviatile material at their glacier edge. In some examples visited, the terrace was formed principally of lacustrine or marine deposits with only a thin covering layer of fluvioglacial material, in others they are dominantly of fluvioglacial material (kame terraces), *e.g.* the gravel terraces of relatively wide extent at Hullet, Julianehåb district (WEIDICK 1963b, p. 30).

Wide terraces formed of fine sediments occur around the Tasersiaq lake–Sarfartôq river in the Sukkertoppen district. Detailed investigations of the formation of these terraces are under way by GOLDTHWAIT *et al.* (1964), but it is already known that they were formed essentially between 1865 and 1938 as a map made by J. KREUTZMANN in 1865 shows that the lake was still barred at this time by a glacier lobe from the Sukkertoppen Iskappe. Also in the Sukkertoppen district, a minor terrace formed as an alluvial plain in connection with a proximally situated terminal moraine at the maximum extent of the glacier, is seen in front of Tasiussaq A glacier (b–b in Fig. 55, p. 131). This glacier, like that in the Sarfartôq area, barred a lake that was still present in the last century but which emptied during the recent deglaciation.

Minor esker features have been seen occasionally in the deglaciated terrain but they mostly have lengths of only few metres. Longer eskers have been seen on aerial photographs at Kuánerssuit, Disko island (see fig. 15 and pp. 46–47) and on Alfred Wegeners Halvø in the Umanak district (HENDERSON personal communication) but neither of these localities have been investigated personally. In both cases the eskers are situated in wide valleys.

Among minor glacial features, fluted moraine surfaces occur in the areas deglaciated in historical time. The formation of fluted moraines seems to be connected with a low inclination of the ground and material consisting of coarse or fine gravel.

5.4. Glacial striae

The erosional features of most significance to a reconstruction of past glacial conditions are the minor ones, especially glacial striae and lunate fractures, which indicate the direction of ice movement.

One interesting point connected with these is the presence of directional marks immediately outside the historical moraines of local glaciers, which indicate ice movements from the Inland Ice, at right angles to the local movements, *e.g.* at Íkátûssaq and Íkamiut kangerdluarssuat north of Sukkertoppen town, and at Sermiarssuit sermikavsât on the northern side of Nûgssuaq peninsula.

6. PREHISTORIC ICE MARGIN DEPOSITS

The procedure used in mapping the deposits, by interpretation of aerial photographs and by fieldwork in key areas, has already been mentioned in the introduction. The collected information was compiled on the Geodetic Institute's map sheets at a scale of 1:250,000 and later compiled on a map at 1:500,000 reproduced here as plate 3.

Though isolated ice margin features and deposits were widely distributed over the area and could be found, for example, at the mouth of many fjords or valleys, mappable ice margin deposits were found to be concentrated in several zones parallel to the present Inland Ice margin. The first of these zones is situated immediately outside the area glaciated during historical time (the inner zone). The next occurs at a distance of 5–40 kms from the present ice margin, depending on the altitude and relief of the area (the outer zone). The third zone comprises nunatak moraines in the high coastal mountain regions of Holsteinsborg and Sukkertoppen. The width of the individual zones is mostly 1–2 km for marginal moraine systems and up to 5 km for terminal moraine systems in places where they are continuously developed.

In addition to these deposits and features from an expanded Inland Ice, a small number of deposits due to local glaciation were also mapped. The limitations responsible for the scarcity of the latter have been mentioned earlier. It is a characteristic of these deposits that they occur in much narrower zones than those of the Inland Ice, or even as a single moraine ridge (cf. fig. 30b), due probably to the greater influence of local topography.

The differentiation of Inland Ice deposits into fairly distinct zones allows their recognition and correlation over long distances. Furthermore, a third dimension to each zone is provided by marginal moraine systems in valleys, and nunatak moraines. Although ice flow and local meteorological conditions around nunataks will modify the ice surface in their vicinity, nunatak moraines can still be regarded as providing a valid picture of the altitude of former ice surfaces. Thus, the horizontal and vertical distribution of the deposits define the ice surface existing at their deposition.

Fig. 30a


Fig. 30b

Fig. 30a. Moraines of inner zone, Inland Ice margin, Nûgssuaq peninsula. Aerial photograph, Geodetic Institute's route 518 A–Ø, no. 1507 (18.7.1949). Copyright Geodetic Institute. 30b. Marginal moraines, formed at glaciation limit 400–600 m a.s.l. Nákâgajoq mountain group (a–a), Qôrorssuaq valley, Nûgssuaq peninsula north of Torssukátak fjord (cf. fig. 24 and pp. 123 and 134). Aerial photograph, Geodetic Institute's route 514 H–NØ, no. 2379 (15.7.1948). Copyright Geodetic Institute.

These surfaces, like that of the historical readvance, can be described in terms of their relationship to the present ice surface. In fig. 31 the vertical separation of deposits of the inner and outer zone from the present Inland Ice, Δh_i and Δh_o respectively, are expressed as a function of the height of the Inland Ice at the same locality, h.

As described earlier z (Δh for $h = 0$) is defined as the surface characteristic for the zone. For the zone of historical time this is 120 m, for the inner zone 350 m and for the outer 650 m. With a mean inclination of the ice surface of 7.6°, as at present, this would result in their terminal moraines being respectively 900, 2600 and 4900 m in front of the present ice margin.

The determination of the surface characteristic of the nunatak zone found near the Davis Strait, is very approximate since it requires an extrapolation of the data over 100 km to the present ice margin. However, as the deposits are situated 600–1000 m a.s.l. in the coastal mountains the surface characteristic for the nunatak zone must be between 1000 and 2000 m.

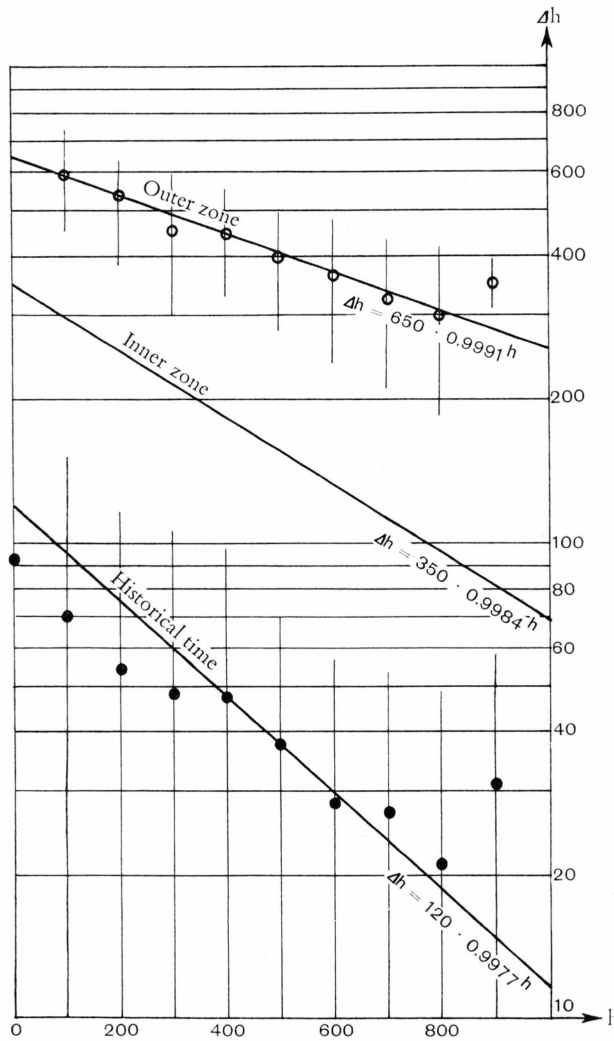

Fig. 31. Altitude (Δh) of historical and prehistoric zones of ice margin deposits expressed as a function of the altitude of the present ice margin (h) (cf. fig. 21, p. 62). The filled and open circles on the curves of the zone of "historical time" and "outer zone" respectively, indicate means of observed values, the standard deviations of which are given by the vertical lines. Units in metres.

Consideration of the appropriate surface characteristic greatly assists the correlation of disjunct systems of moraines.

Another approach to the relative and absolute dating of the moraines that can be used in coastal areas is the relationship of the moraines to past sea levels. Fig. 2 shows that the isostatic recovery of the land has been the dominant factor in the shift of the coastline during the post-glacial period until 4000 B.P. Therefore, where a beach ridge has been developed on a moraine it cannot be due to a eustatic rise of sea level after the formation of the moraine. Thus the uppermost beach ridge must be nearly contemporaneous with the deposition of the moraine, formed, for example, by the wave erosion of marginal moraines of calving glaciers. Where sorted deposits were laid down raised deltas or delta terraces result which can be similarly utilised.

The correlation of deposits from local glaciations is more difficult and often can be done only in a general way through the determination of the former glaciation limits, and where possible, by their relationship with raised beach ridges. The last method also partly links these isolated deposits with those of the Inland Ice.

7. INLAND ICE DEPOSITS–INNER ZONE

7.1. Definition of inner zone

The inner zone comprises a belt of ice margin deposits running parallel to the existing Inland Ice margin at distances of c. 0.5 to 10 km. In places where the former glacier has been a calving lobe, the distance between the westernmost ice margin deposits of this zone and the present glacier front can be up to c. 25 km. The form of the surface of the inner zone can be expressed by

$$\Delta h_i = z_i e^{k_i h}$$

where z_i, the surface characteristic is 350 m and k_i is -0.00161, *i.e.* $\Delta h_i = 350 \cdot 0.9984^h$, as shown in fig. 31.

7.2. Julianehåb district to Godthåb district

The Narssarssuaq stage described previously from the northern parts of the Julianehåb district (Weidick 1963b) fits fairly well with the characteristic of the inner zone mentioned above.

Around Kiagtût sermiat and Narssarssuaq the deposits of the Narssarssuaq stage are not cut by any sea level higher than the present one, though ruins of Norse farms on them indicate that their age is more than 1000 years. The location of the stage is indicated in fig. 32.

In the Frederikshåb district, Holst (1886, p. 56) mentioned the occurrence of old moraines at Fox Havn in Arsuk Fjord c. 5 km west of the present glacier front, but no details are known about them. Further north in the districts south of Frederikshåbs Isblink, current investigations suggest the existence of deposits which can be correlated with this zone (Kelly personal communication).

Sparse ice margin deposits are reported from the southern part of the Godthåb district by Graff-Petersen (1952) and further north ice margin deposits have been seen on aerial photographs around Nakaissorssuaq, Qajartoriaq and Isortuarssûp tasia. All the deposits which have been observed are plotted on the map in plate 3, but because of their scattered nature a closer correlation of them is impossible.

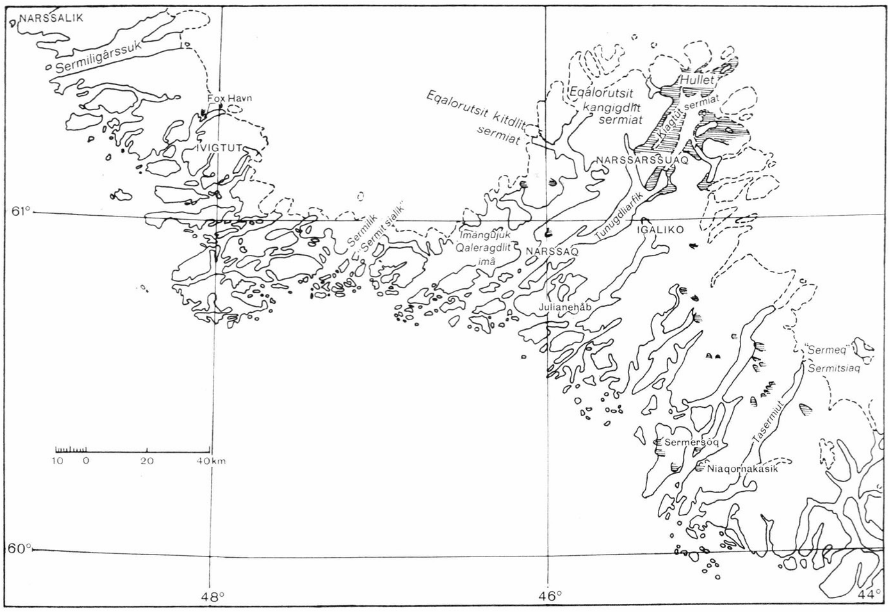

Fig. 32. Map showing the extent of prehistoric marginal stages in Julianehåb district.

In the northernmost parts of the Godthåb district, at Kangersuneq in the interior part of Godthåbsfjord, the aerial photographs show groups of moraines, formed by the glacier lobe of Narssap sermia, stretching towards the ice dammed lake Ujaragtôq (180 m a.s.l.). These ice margin features are situated 100–200 m above present glacier surface.

7.3. Sukkertoppen, Holsteinsborg and Egedesminde districts

Whilst there is only sparse evidence of the inner zone from the area between the Julianehåb and Godthåb districts, the zone is amply represented further north.

7.3.1. Interior part of Søndre Isortoq

Three areas with ice margin deposits which possibly belong to this zone, can be seen on the aerial photographs (see plate 3).

On the northern side of Søndre Isortoq valley a large system of marginal moraines surrounds one of the recent lobes from the eastern Sukkertoppen ice cap. Together with marginal moraines on the nunataks further east, these ice margin deposits indicate the former existence there of a great major lobe of the Inland Ice c. 13 km west of the present margin, which received tributaries from the local Sukkertoppen ice cap

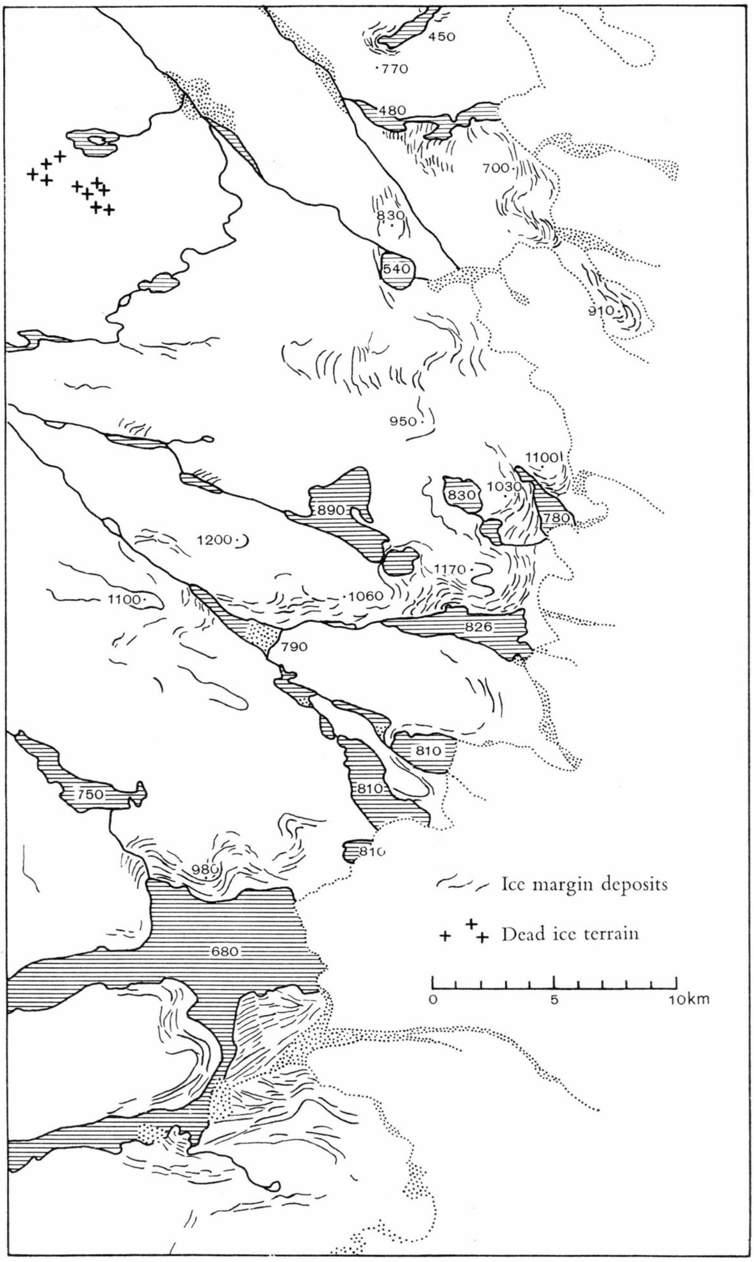

Fig. 33. Details of the inner zone south of Søndre Strømfjord (cf. plate 3), compiled from aerial photographs. Based on part of the Geodetic Institute's 1:250,000 map sheet 66 V2, Søndre Strømfjord Øst. By permission of the Geodetic Institute.

north of it. The old moraine system lies 400 m above the present front, which itself is only a few metres above sea level.

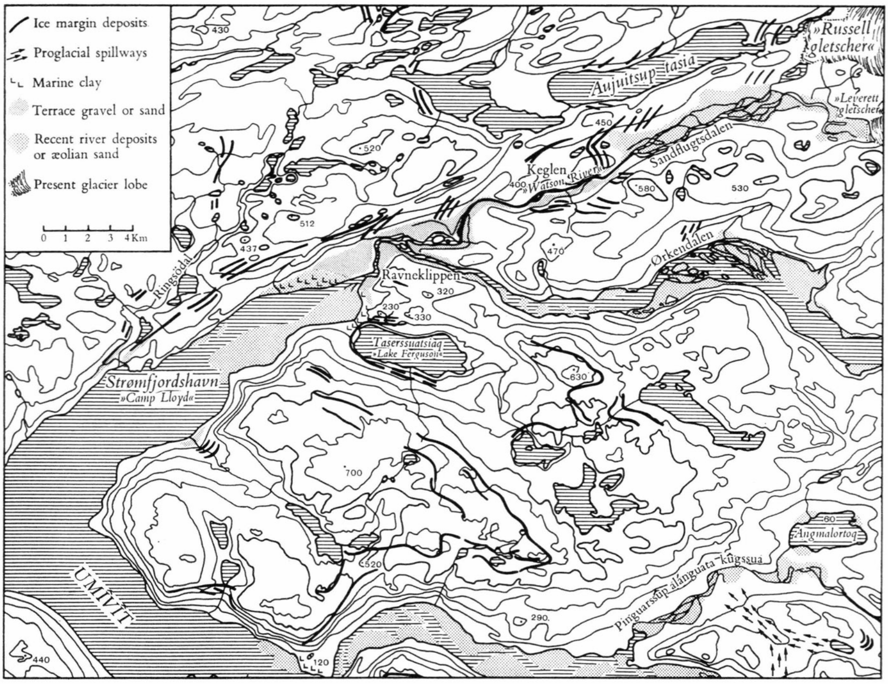

Fig. 34a. Sketch map of the area at the head of Søndre Strømfjord. Map based on the Geodetic Institute's 1:250,000 map sheets 66 V2, Søndre Strømfjord Øst and 67 V2, Nordre Strømfjord Øst. By permission of the Geodetic Institute.

7.3.2. Tasersiaq lake to Ørkendalen

The southernmost parts of this area are being investigated by Ohio State University expeditions and more detailed information than that given here from aerial photographic interpretation can be expected.

Individual sections of the Inland Ice margin from the southernmost part of the area have been described previously (Weidick 1963a). There, as well as in the area as a whole, the inner zone is developed as a 5–15 km wide belt in which the underlying bedrock is veiled by thick ground moraine, the surface of which is a closely spaced series of marginal moraine ridges running parallel to the present ice margin. Even where the outer limit is not marked by a ridge feature it can still be recognised as an attenuated drift border, in the sense of Flint (1947, p. 157). It has not been possible to map the area thoroughly from the aerial photographs, but fig. 33 gives an idea of the extent of the zone.

The ice margin in the southernmost part of the area is situated around 700 m and the upper limit of the inner zone in the same place is c. 1000 m a.s l. In the north, the present Inland Ice margin is 300–400 m a.s.l. and the upper limit of the zone is here at 700–800 m. a.s.l.

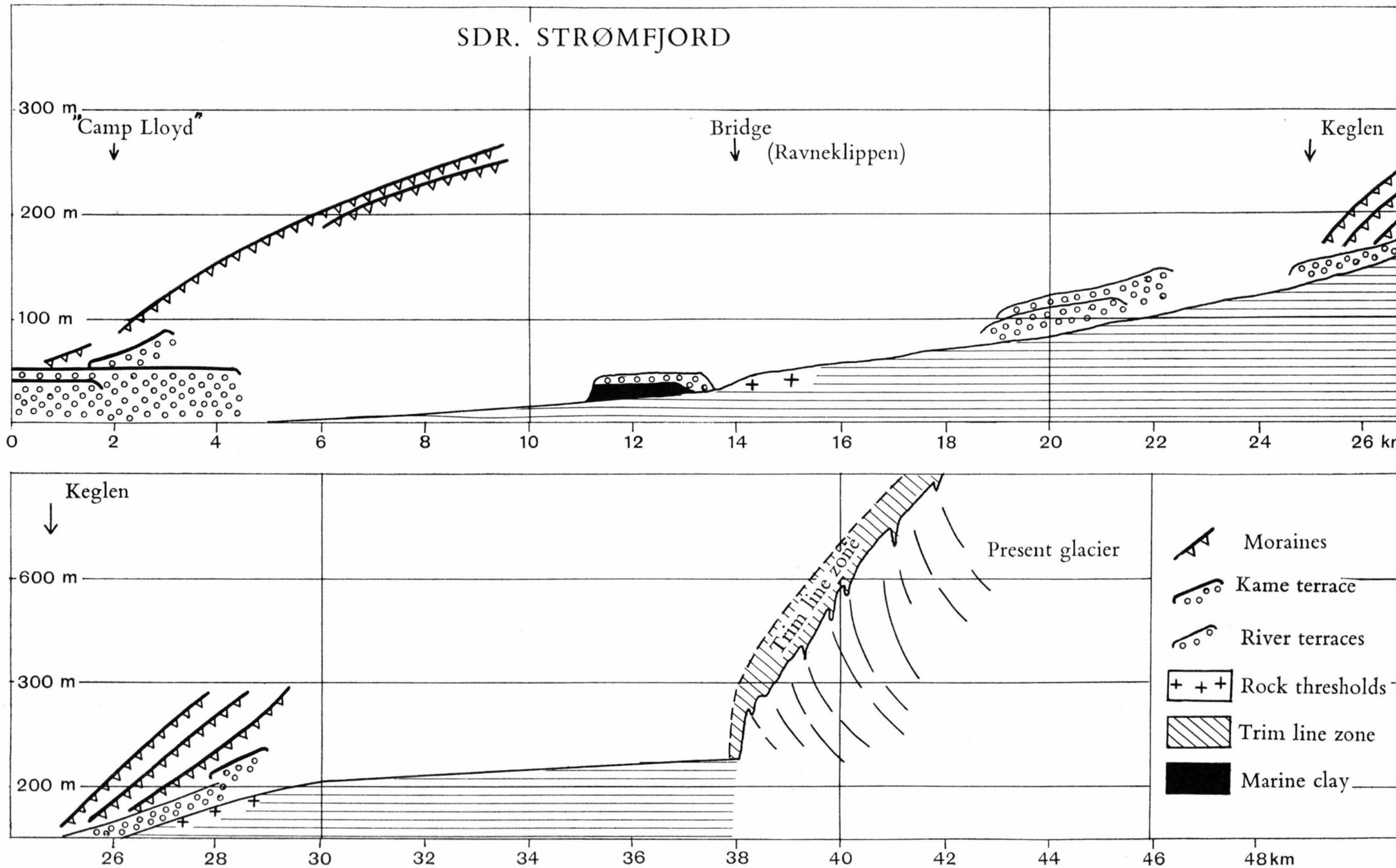

Fig. 34b. Profile along the valley at the head of Søndre Strømfjord (Sandflugtsdalen).

7.3.3. Søndre Strømfjord to Sandflugtsdalen

The area has been dealt with before by HOBBS (1941), BELKNAP (1941) and BØCHER (1949, pp. 60–61). Only BØCHER has tried to establish the relative chronology of the deposits in the valley. He discerns two stages in the retreat of the Inland Ice margin; one around "Camp Lloyd" (Strømfjordshavn) at the head of Søndre Strømfjord and one at mount Keglen, c. 25 km further east. These are shown on plate 3, and on the sketch map and profile in fig. 34. The surface characteristic of the deposits shows that only the mount Keglen stage belongs to the inner zone. Then the ice margin was situated 12–13 km west of the present margin of "Russell gletscher". The older marginal moraines of the zone are cut by a terrace, c. 30 m above the present river ("Watson River"), whilst the younger moraines grade into kame terraces and the river terrace mentioned above. Both the moraines and the kame and river terraces are lithologically similar, being formed of coarse gravel and pebbles with numerous rounded boulders.

Towards the north the inner zone is represented by isolated areas of moraine around the lake of Aujuitsup tasia and the glacier lobe of Isúnguata sermia.

7.3.4. Isúnguata sermia to Eqalungmiut nunât

On the aerial photographs numerous ice margin deposits can be seen between the two lobes from the Inland Ice, Isúnguata sermia (named "Otto Nordenskiöld Glacier" by HOBBS) and Inugpait qûat (plate 3). The deposits are grouped into two zones situated 5–10 and 10–15 km respectively from the present ice margin. The inner of these (the inner zone moraines) seem to be situated c. 300 m above the present ice margin, which is between 200 and 500 m a.s.l. The inner zone occurs only in the highest parts of the area. Further north, towards the glacier lobe of Usugdlûp sermia, only scattered and isolated ice margin deposits of this zone can be seen.

7.3.5. Interior part of Arfersiorfik fjord

The Inland Ice margin in the area in general is only 200–400 m a.s.l. and in two places lobes of the Inland Ice reach down to sea level: Usugdlûp sermia and Nordenskiölds Gletscher. Only a few isolated moraines in this sector can be referred to the inner zone (plate 3).

7.4. Disko Bugt

The coastal stretch in this sector is only 30–40 km wide, enabling easy access to the present Inland Ice margin and its older ice margin deposits. The maximum altitudes for the area are only c. 700 m a.s.l.

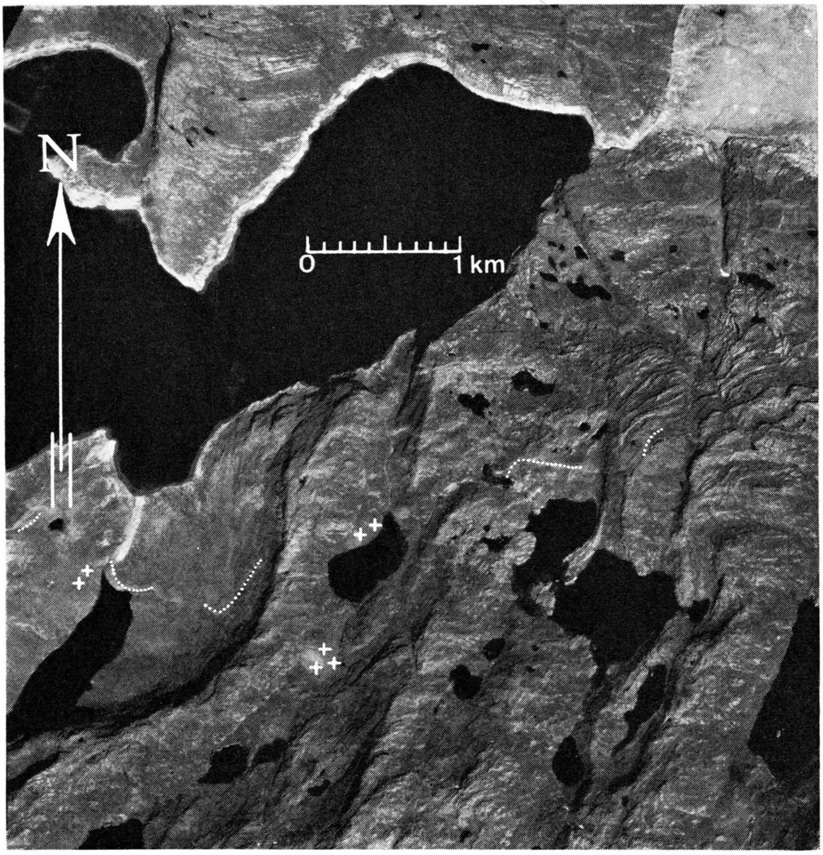

Fig. 35. Marginal moraine (dotted) and stoss moraine (crosses) south of Nunatap tasia, Jakobshavns Isfjord. Geodetic Institute's route 273C, no. 106 (12.7.1964). Copyright Geodetic Institute.

7.4.1. Jakobshavns Isfjord and Tasiussaq area

In the southeasternmost branches of the Tasiussaq fjord complex, two well developed moraine systems have been observed, both in the field and on aerial photographs, running across the peninsula of Qavdlunâp nunâ near Alángordliup sermia. The moraines consists of river gravel and are cut by terraces, the highest of which is c. 30 m a.s.l. In the terraces, and elsewhere at several places in the interior parts of Tasiussaq up to 35 m a.s.l., are deposits of clays which appear to be non-marine. It is therefore highly possible that the deposits and terraces are products of an ice dammed lake. This would require the blocking of the entrance to Tasiussaq at Qajâ (see fig. 47) by a glacier in Jakobshavns Isfjord.

Deposits observed on aerial photographs in the most eastern parts of Tivssarigsoq (fig. 35) and on the southern side of Nunatap tasia are a possible continuation of the system on Qavdlunâp nunâ.

At Qavdlunâp nunâ near Alángordliup sermia the moraines of this system at 200–240 m a.s.l. are poorly sorted gravel ridges, but towards the west end of Nunatap tasia there are great gravel heaps on the glacially upstream side of the valleys. They should be classified as stoss moraines (ANDERSEN 1960, pp. 64–65, 107) having been deposited behind the bedrock escarpments which are there parallel to the present ice margin. In the same area, around the west end of the lake of Nunatap tasia, there are clay terraces of prehistoric age which seem to belong to a phase in the retreat of Jakobshavns Isbræ from the position indicated by the stoss moraines.

Further towards the west on the southern side of Jakobshavns Isfjord, large marginal moraines were observed in the field, on the top of the mountain ridge south of Qajâ. They are composed of blocks without much matrix. South of Qajâ the moraines are situated at 80–100 m a.s.l. and they terminate abruptly at a steep "bird cliff", giving no indication of the level of the sea at the time they were deposited. The archeological site of Qajâ, situated close to the moraines, indicates a minimum age for them of 3000–3500 years (LARSEN and MELDGAARD 1958, FREDSKILD 1967).

7.4.2. Eqip sermia to Torssukátak fjord

Neither aerial photographs nor field investigations have shown anything other than isolated ice margin deposits of the inner zone between Jakobshavns Isbræ and Påkitsup ilordlia. However, north of Påkitsup ilordlia, a nearly continous system of moraines extends up to the nunatak of Qapiarfik, north of the glacier lobe of Eqip sermia. Parts of this moraine system have been investigated by DE QUERVAIN and MERCANTON (1925, p. 237), BOYÉ (1950, pp. 62–63) and BAUER (1955c, pp. 91–96). MERCANTON describes moraine at three levels in the Qapiarfik area, at 700 m a.s.l. and at 150 and 100 m above the present ice margin; the last two must be phases of the inner zone. BOYÉ in contrast to MERCANTON, states that at Qapiarfik, unlike the southern side of Eqip sermia, there are continous ice margin deposits showing evidence of an "englacement d'ensemble", with widespread frontal and lateral moraines indicating a complicated history of deglaciation of the nunatak. Plotting from aerial photographs shows numerous moraines in the area, in a zone 100–200 m above the present Inland Ice surface.

On the southern side of Eqip sermia all three investigators referred to above report the existence of extensive ice margin deposits at 100–200 m above the present ice surface. These are also clearly seen on the aerial photographs (fig. 36). BAUER (1955c, pp. 95–96) mentions the presence of three terraces with an altitude of 25 m at the front of Eqip

Fig. 36. Moraines of the inner zone, Inland Ice margin south of Eqip sermia (loc. 74 in plate 2). Geodetic Institute's route 272E, no. 252 (29.6.1964). Copyright Geodetic Institute.

sermia but it is not clear whether they are ice margin terraces or are of marine origin. East of these terraces he describes a "moraine ancienne" presumably of prehistoric age, which goes straight down to the present sea level without signs of marine disturbance.

North of Qapiarfik, no ice margin deposits are seen on the steep slopes surrounding the lobe of Kangilerngata sermia. The zone can be picked up again further north at Anap nunâ and around the glaciers at the head of Torssukâtak fjord. At Anap nunâ, the deposits are just outside the historical trim line zone at altitudes 100–200 m above the present ice margin. Deposits of the zone investigated in the field near the front of Sermeq avangnardleq showed no trace of reworking by the sea other than that occurring at present sea level.

7.4.3. Nûgssuaq peninsula

A continous belt of ice margin deposits crosses the peninsula, parallel to the present margin of the Inland Ice. Their height above the present Inland Ice margin remains constant at 100–200 m. An aerial photograph of their central section is shown in fig. 30, and the extent of the moraines can be seen in plate 3.

The moraines and kame terraces are often of considerable size, rising ten to twenty metres above the underlying rock surface, and consist mostly of moraine gravel with rounded boulders. The whole zone across Nûgssuaq has been called the Drygalski stage because its first description was by members of the German expedition of 1891–1893 led by Drygalski.

In part, the deposits can be split up into two phases, an inner phase (or stage), which at Torssukátak fjord can be seen to be cut by present sea level only, and an outer phase, which towards the fjord ends on a steep rock wall near a large lake, 2–3 km west of the Inland Ice, called Amitsup tasia.

In the central parts of the peninsula the moraines (as shown in fig. 30a) are broken into a series of short lengths which show all transitions in form between stoss moraines and marginal moraines, with an accompanying variation in composition. In front of the glacier in the central part of the peninsula and further north at the Store Gletscher in "Qarajaks Isfjord", the moraines are block moraines, whilst in the intervening stretch they are gravel with rounded boulders. At the southern margin of Store Gletscher also, the moraines are undisturbed down to the present sea level.

Besides the moraines on Nûgssuaq, Drygalski partly described their northern continuation (1897, pp. 57–60) and from his description, and the aerial photographs, the zone can be followed easily as far as "Ingnerit fjord" (Perdlerfiup kangerdlua). No inner zone ice margin deposits are known with certainty from north of there.

Thus in the northern sector of the area ice margin deposits of the inner zone stretch in a belt for 120 km from Eqip sermia to Itivdliarssup kangerdlua. There is also clear evidence that the deposition of at any rate an inner system, was contemporaneous with a sea level no higher than the present. In general, the surface characteristic of the zone in this sector is amazingly low, 100–200 m, and this value seems to be constant for higher elevations of the present ice surface, up to c. 600 m a.s.l. at the nunataks of Qapiarfik and Rensdyrnunatak. This steep profile of the inner zone surface may be due to the presence of bedrock thresholds beneath the ice beyond the heads of the ice-fjords of Torssukátak and "Qarajaks Isfjord", thresholds which apparently still operate and are responsible for the steepness of the present lobes.

8. INLAND ICE DEPOSITS – OUTER ZONE

8.1. Definition of outer zone

The outer zone can be characterized as a belt of ice margin deposits roughly parallel to the actual Inland Ice margin at distances of 5 to 40 km. The greatest distance is exceptional and is presumed to indicate former calving glacier lobes. The ice surface is described by,

$$\Delta h_o = z_o e^{k_o h}$$

where z_o, the surface characteristic equals 650, and k_o equals – 0.000921, *i.e.* $\Delta h_o = 650 \cdot 0.9991^h$, as shown in fig. 31.

8.2. Julianehåb and Frederikshåb districts

The Tunugdliarfik stage of the Julianehåb district has been described previously (WEIDICK 1963b) and only the main features will be given here. Its deposits do not lie far outside the younger Narssarssuaq deposits but their surface characteristic agrees with the curve for the outer zone given in fig. 31. These deposits are more weathered and their surface morphology more subdued by solifluction, than the deposits of the Narssarssuaq stage (see 7.2, p. 83). There is also a lithological difference, large angular boulders being more abundant in the deposits of the Tunugdliarfik stage. Only on the plain at Narssarssuaq air base do the deposits reach the coast and there the flattening of a kame terrace may indicate the position of the contemporaneous sea level, at 10–15 m above the present one.

Little has been published about the Frederikshåb district, but the recent investigations in the northern parts point towards the existence of ice margin deposits of the outer zone in this area (KELLY personal communication).

8.3. Godthåb district

As the map of plate 3 shows, not much is known about the ice margin deposits of the southern parts of this district. In the north however, numerous ice margin deposits have been observed in the field and on aerial photographs.

8.3.1. Bjørnesund (Aĝdlumersat)

KORNERUP (1890, p 100) records terraces up to an altitude of 192 m a.s.l. at Qáqarssuaq on the southern side of the entrance to Bjørnesund which VOGT (1933, p. 17) describes as being of marine origin, though this is doubtful. On the slope of Ũmánarssuaq mountain on the south side of Bjørnesund, terraces also occur at altitudes of 500–600 m a.s.l. Since the distance between Qáqarssuaq and Umánarssuaq is 16–17 km it is possible that both deposits are of glacial origin and belong to a stage of the outer zone.

8.3.2. Sermilik

Around Sermilik fjord and Sermilik glacier ice margin deposits are widespread, though scattered in their occurrence. Inland, minor ice margin deposits occur around the lake of Qajartoriaq (500–600 m a.s.l.) and on the northern slopes of Sinarssuk valley (800 m a.s.l.). Both localities are situated 10–12 km from the present lobe from the Inland Ice whose margin is at c. 600 m a.s.l. On the north side of Sermilik glacier, a c. 1 km long stretch of ice margin deposits is present at an altitude of 900–1000 m a.s.l., the present glacier surface being situated at c. 700 m a.s.l. Elsewhere in the area there are less well marked ice margin deposits, situated around 900 and 1000 m a.s.l.

Terraces at Sermilik fjord are numerous and well developed. They were first described by KORNERUP (1890, pp. 97–99) who compared their formation with that of the present alluvial plain in front of the glaciers of Frederikshåbs Isblink, thus interpreting them as ice margin terraces. According to him they are composed of marine sediments overlain by fluviatile sediments which have a flat surface situated at 13 m a.s.l. However, STEENSTRUP (*in* JESSEN 1896, p. 150, VOGT 1933, p. 17) gives the altitude of the terraces as between 43.3 and 51.5 m a.s.l., which agrees with the altitudes given on the Geodetic Institute's map sheet 1:250,000 (63 V 1, Færingehavn). On this map the maximum altitudes of the plains and terraces around the mouth of Sermilik fjord, at Marraq as well as Sanerâta timâ can be seen to be situated at or above c. 50 m a.s.l. Presumably KORNERUP has confused measurements made in feet and in metres.

The terraces at Marraq (fig. 37) and Sanerâta timâ slope down towards the west, away from Sermilik fjord. Further east, in Amitsuarssuk fjord, marginal terraces or moraines descend from 400 m a.s.l. at the head of the fjord to c. 100 m a.s.l. immediately east of Marraq. These two systems together presumably represent a stage in the position of the lobe of the Inland Ice. The deposits at altitudes of 900–1000 m a.s.l. mentioned above are a natural continuation of this stage towards the east.

Fig. 37. Marraq, seen from the south. Aerial photograph, Geodetic Institute's route 504 N–N, no. 49 (26.7.1948). Copyright Geodetic Institute.

8.3.3. Head of Ameralik fjord

The terrain at the head of the fjord has been described by numerous people since 1729 (Weidick 1959, pp. 141–151), though only a few made any observations about the Quaternary deposits. Kornerup (1890, p. 99) mentions three terraces at the head of the Ameragdla branch of the fjord, the lowermost of which was stated to be at 60 m a.s.l. and the uppermost at 106 m a.s.l. Kornerup seems to have interpreted these as ice margin terraces. Nansen (1890, p. 539) mentions a terminal moraine at the entrance of Austmannadalen and marine deposits in the valley up to 20 m a.s.l., which contained "Blaaskjel" (*i. e. Mytilus edulis*). According to the Geodetic Institute's 1:250,000 map sheet (67 V 2, Kapisigdlit) this terminal moraine has a very level surface at c. 50 m a.s.l. It is possible that the upper terraces mentioned by Kornerup are an older stage, and the terminal moraine in Austmannadalen a younger stage of the outer zone.

These ice margin deposits lie in a belt 15–20 km from the present margin and can be followed for 50 km from the lake of Kangerdluarssungûp taserssua in the south to Tungmeralik in the north. At the present ice margin where the glacier surface is situated between 100 and 500 m a.s.l., the deposits are at 800 m a.s.l.

8.3.4. Kangersuneq in Godthåbsfjord

The ice margin deposits in Ameralik fjord can be seen on the aerial photographs to continue north to Kangersuneq. Along the sides of Kangersuneq fjord they decrease in elevation from 800 m a.s.l. in the south at Tungmeralik, to 50–100 m a.s.l. in the north at Nâlagfik (10 kms north of Kapisigdlit). At Kapisigdlit kangerdluat they outline a broad lobe which extended towards this fjord. Investigations of the area in 1965 revealed the presence of well sorted pebbles and features like beach-ridges on the moraines between 50 and 80 m a.s.l. The moraines themselves consist of numerous rounded boulders in a sand to gravel matrix. Other clear evidence of marine action on the moraines was found further north, at Nâlagfik, where a continous series of beach ridges extends from the present sea level up to c. 50 m a.s.l. Above this are vegetation covered ridges up to a wide terrace between 70 and 80 m a.s.l.

8.3.5. Narssap sermia and Ujaragssuit pâvat

On the aerial photographs a c. 10 km long system of moraines belonging to the inner zone can be seen on the north side of Narssap sermia glacier. At higher altitudes, between 750 and 800 m a.s.l. west of this and northwest of Sarqánguaq, are fragments of moraines about 0.5–1 km long which from their situation belong to the outer zone. To this same zone is referred the great system of ice margin deposits which surrounds Ujaragssuit pâvat at 900 m a.s.l. in the east to 300 m a.s.l. in the west. This system can be traced towards the east where it forms two nunatak moraines at 750 and 900 m, and towards the west where it finally ends at the steep mountain slopes near Majuala, giving it a total length of 15 km.

At Majuala itself, JENSEN (1889, pp. 92–93) has observed terraces at altitudes of 6, 26, 57, 97, 123 and 130 m a.s.l. and in addition, a marginal terrace at 489 m a.s.l. All these terraces were reported to consist of gravel with rounded boulders, but whether they are of marine or glacio-fluviatile origin cannot be decided from the description.

8.3.6. Ilulialik and Narssarssuaq

Both localities are well known for their marine deposits and GIESECKE (1910, p. 140) states that these deposits are to be found in all bays on the north side of Godthåbsfjord.

In Ilulialik, the moraine systems drop from the north down to near sea level but in Ilulialik bay the only features are terraces at c. 15, 40 and 100 m a.s.l. Some of these are developed in marine sediments but the highest situated south of Eqaluit, is possibly a kame terrace.

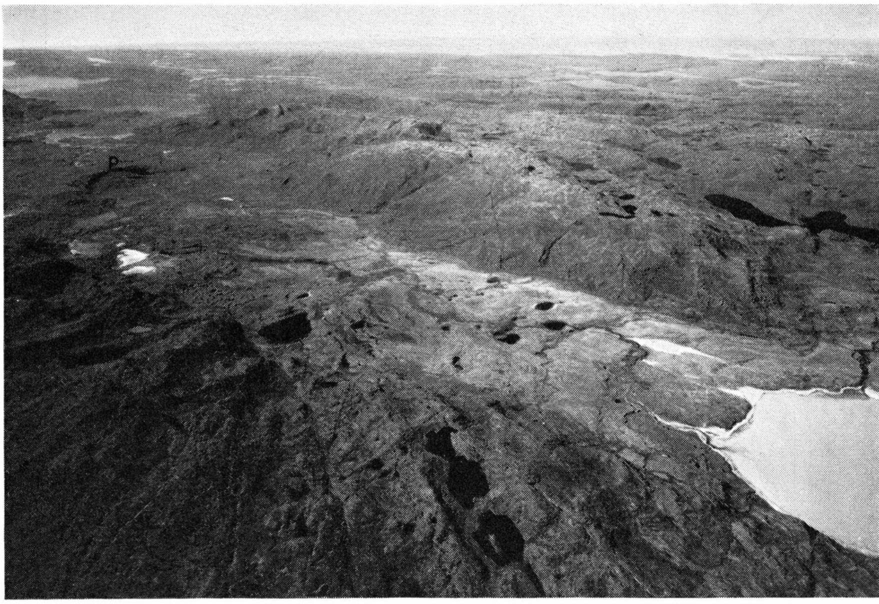

Fig. 38. Narssarssuaq plain north of Godthåbsfjord. In the right foreground Tasers-suaq lake and in the left background, Pingorssuaq mountain (p). Geodetic Institute's route 505 D_1–V, no. 5384 (4.9.1948). Copyright Geodetic Institute.

In Narssarssuaq (fig. 38, 39), the main moraine system can be followed only on the southeastern side of the valley. There, it ends in a system of kame terraces with the principal terrace lying between 60 and 80 m a.s.l. Just outside the kames and continuing across the southern part of Narssarssuaq valley as far as Pingorssuaq mountain is a broad zone of kettle moraine. The fluviatile deposits, which form the alluvial plain surface over much of the Narssarssuaq valley, must have been laid down in conjunction with the lowermost moraines, partly burying older dead ice near Pingorssuaq. This younger stage seems to have graded to a sea level of c. 50 m, as shown in the profile fig. 39.

In the cliffs towards the coast, and in the gullies in the plain, the sandy gravel of the terrace surface can be seen to grade downward into laminated sand and silt and finally, between 20 and 35 m a.s.l., into clayey silt. Extremely rich shell layers were found in the clay, at the base of the cliffs around the outer coast of the headland in the inner southern part of Qugssuk inlet (× in fig. 39). However, the chronological relationship of the deposits to these terrace surfaces above and others further inland is not clear.

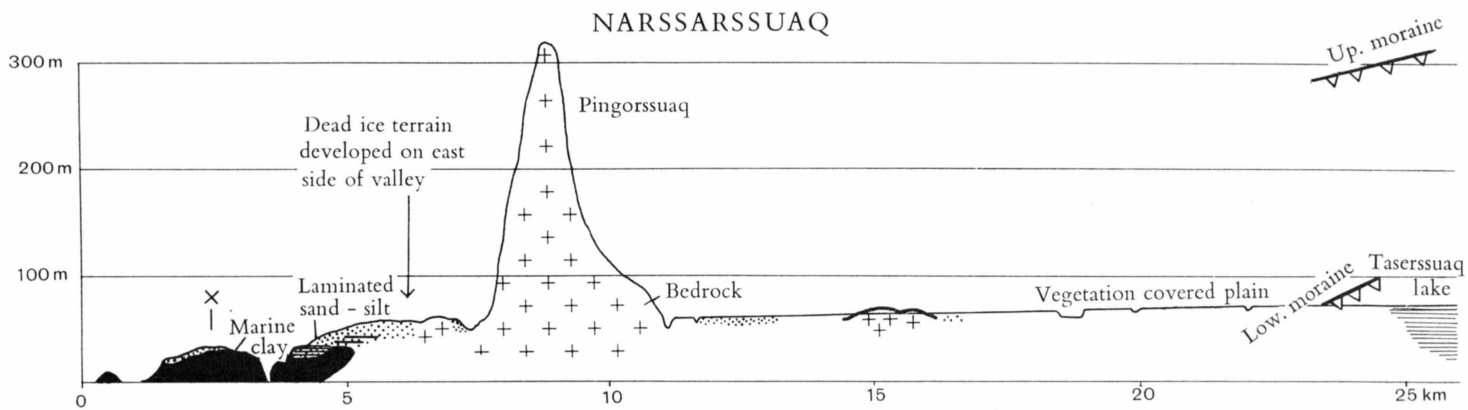

Fig. 39. Profile along the valley of Narssarssuaq. Vertical exaggeration × 20.

8.4. Sukkertoppen district

The district can be divided topographically into three units, a 1200 m high plateau north of Godthåbsfjord and east of Taserssuaq lake, the lowland area around the interior part of Søndre Isortoq fjord and the 2000 m high Sukkertoppen plateaus in the northern parts of the district.

8.4.1. Taserssuaq

The lake of Taserssuaq at 74 m a.s.l., is surrounded by ice margin deposits. Above the central parts of the lake, they lie at 600–700 m and from there they drop down towards the south and the north. At the southern end of the lake, towards Narssarssuaq and Ilulialik, they are between 100 and 300 m a.s.l. At the northern end they descend, at c. 500 m a.s.l., into a valley which leads westwards towards the lake of "Taserssuatsiaq", but there is no evidence of an extension of the system further west.

On the western side of the lake the ice margin deposits in the central part end at a valley which leads westwards to Fiskefjord (Niaqúngunaq). Since the bottom of this valley is below 100 m a.s.l. it is probable that the ice of the outer zone filled the valley and reached Fiskefjord. However, neither the aerial photographs nor JENSEN's (1889, pp. 83–84) report of a visit to the area suggest the existence of ice margin deposits in the valley.

On the northeastern shore of Taserssuaq (fig. 40) there is a well developed nunatak moraine system at 600 m a.s.l. The easterly branch of the lake, south of these nunatak moraines, leads to a glacier lobe from the present Inland Ice (the "Sarqaq glacier"). The ice margin deposits of the outer zone cannot be followed very far towards the east along this "Sarqaq branch" of the lake, but it seems likely that the ice margin of the outer zone was situated 700–800 m above the present one.

8.4.2. The area between Tasersuaq and Quvnerup qâqâ

The landscape here is an irregular plateau, situated mostly between 800 and 900 m a.s.l. but in places descending down to c. 400 m a.s.l.

A belt of ice margin deposits stretch from the northern end of Taserssuaq (see plate 3) towards the Majorqaq valley in the Søndre Isortoq area. The deposits are at 600–700 m until the "Qaersutsiaq" valley where they descend nearly to present sea level. Along the sides of this valley they are wider and thicker than on the plateau to the south and two separate stages can be distinguished. The westernmost of these

a

a

Fig. 40. Northern end of Taserssuaq lake, seen from the west. In the right foreground are ice margin deposits. The white dots (a–a) indicate the situation of the nunatak moraines mentioned in the text. In the background "Sarqaq glacier" (loc. 31 in plate 2). Geodetic Institute's route 505 D–Ø, no. 4681 (20.8.1948). Copyright Geodetic Institute.

descends nearly to sea level at the mouth of the valley whilst the eastern one reaches down to c. 300 m a.s.l. c. 6 km east of this.

The mouth of "Qaersutsiaq" valley has been visited by both JENSEN (1889, pp. 79–81) and HOLST (1886, p. 64). JENSEN describes terraces in the valley going up to 600 m a.s.l., built of non-fossiliferous clayey gravel containing rounded boulders. He supposed that they were river terraces, but did not exclude the possibility of their being ice margin features (*ibid.* p. 81). North of "Qaersutsiaq" extensive terraces which are clearly related to the former presence of ice dammed lakes between 400 and 450 m a.s.l., can be seen on the aerial photographs. They can be seen also to connect with the more easterly stage of the moraines in the "Qaersutsiaq" valley mentioned above. A nunatak moraine at approximately 500 m a.s.l. south of the valley is also connected to this inner stage.

Holst states that there are deposits of bluish-grey clay at "Qaersutsiaq" as well as at "Ilulialik", and that at the former locality an upper terrace level is developed on the clays at 31 m a.s.l. He says also that at "Ilulialik" this clay is covered by 1.5–3 m of arenaceous deposits consisting of sand below and gravel with rounded boulders above. In the clay were found shells, including *Pecten islandicus* and *Mytilus edulis*. As far as can be seen from his descriptions, the relationships between the glacial and marine deposits are very like those found in Godthåbsfjord (see pp. 96–97). It seems from the aerial photographs and the Geodetic Institute's maps that the outer moraines at "Qaersutsiaq" valley are cut at least by one terrace, at c. 50 m a.s.l.

8.4.3. Majorqaq area

The area includes the three east-west trending valleys, "Quvneq", Majorqaq and "Ilulialik". In all three valleys are ice margin deposits which slope down towards the west, but in none has a terminal moraine been seen. Presumably they have been removed by recent river erosion.

A more westerly situated stage is present in these three valleys also and is clearly a continuation of the deposits at the entrance to the "Qaersutsiaq" valley. In addition, the eastern stage is composite in character, especially around "Ilulialik". Its deposits have a trend parallel to the deposits of the more westerly stage, but are in two series situated 30–50 and 50–80 m lower respectively.

These moraines around Lûtiviup nunatarssua and "Ilulialik" are of special interest. Their distribution and form indicates that they have been formed by a lobe from the Inland Ice lying between the two large ice caps on the Sukkertoppen highlands. Their proximity to them indicates that the local ice caps have never been much more extensive than now (fig. 41).

The lower glaciation limit which is a prerequisite of an expanded Inland Ice could be the result of either lower temperatures or higher accumulation. Had accumulation been significantly greater than now, the local ice caps would have expanded also. On the other hand, if temperature decreased sufficiently to lower the glaciation limit by some 100 metres, there would only be an insignificant increase in the accumulation area of the ice caps, because of the steep form of the Sukkertoppen plateaus. Thus the implication is that the advanced position of the Inland Ice recorded by the ice margin deposits was due essentially to lower temperatures.

The same argument can be used to set an upper limit to the amount by which the glaciation limit was lower. If it had been more than 300 metres then the present ice caps around the highest points 870 m between

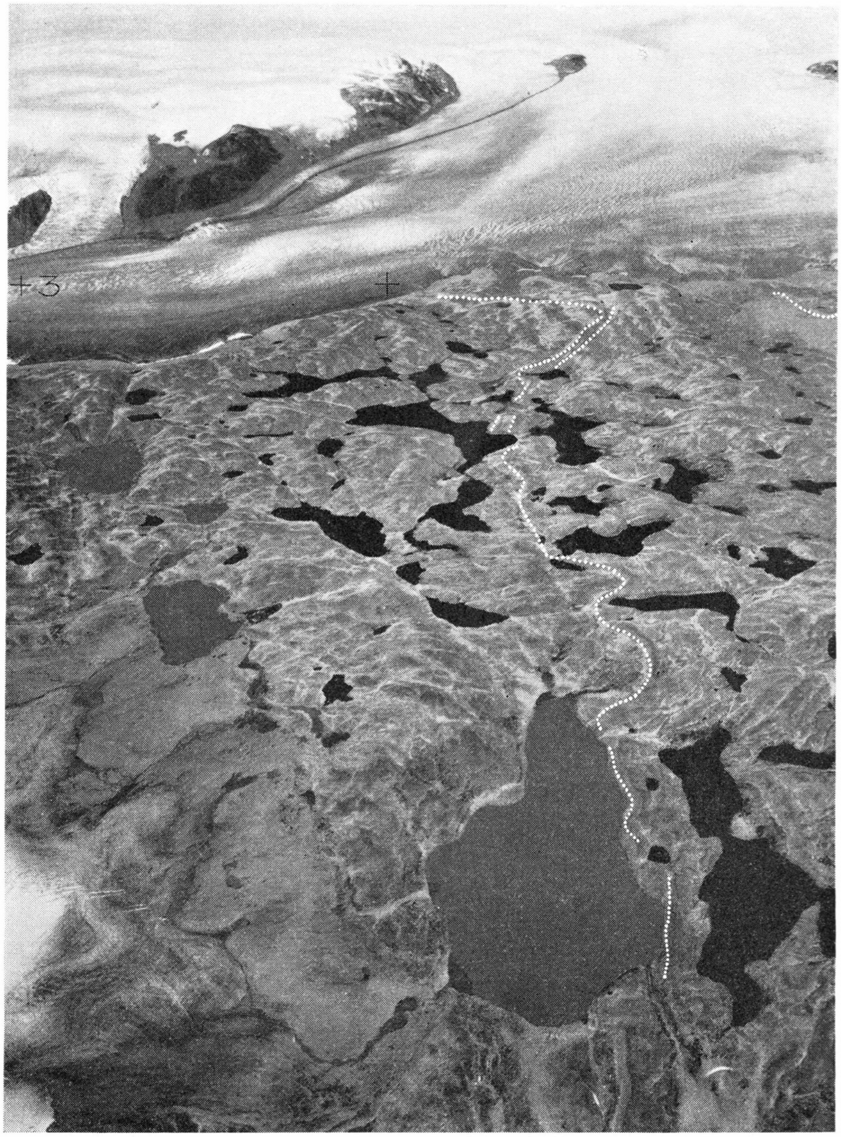

Fig. 41. The uppermost moraines at Lûtiviup nunatarssua, north of Søndre Isortoq, seen from the west. Geodetic Institute's route 505 D–Ø, no. 4720 (20.8.1948). Copyright Geodetic Institute.

"Ilulialik" and Majorqaq (Lûtiviup nunatarssua area) and 930 m on Quvnerup qáqâ (see plate 3 and fig. 42) would have been larger than now, destroying the ice margin deposits from the Inland Ice lobe.

It is not possible to give much information about the elevation of the Inland Ice at the time of formation of the outer zone in the area.

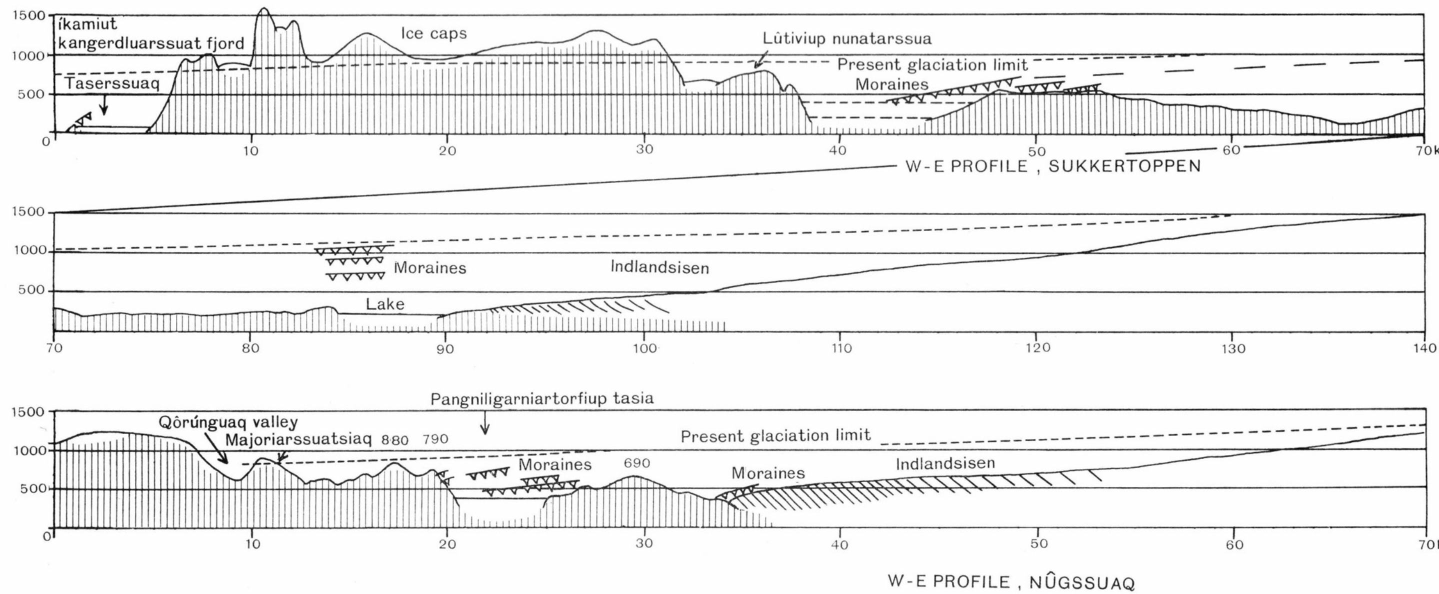

Fig. 42. Profile across Sukkertoppen district compared with a profile across the inner part of the Nûgssuaq peninsula. Instead of íkamiut, read Íkamiut.

The highest ice margin deposits on a nunatak north of "Kingingnerssuaq" (see plate 3) are at c. 1000 m a.s.l. whilst the present surface of the Inland Ice there is at 700 m a.s.l. As the profile in fig. 42 shows, the deposits are probably to be correlated with the two phases or stages of moraines recognisable at Lûtiviup nunatarssua further west.

8.4.4. The area from the Sukkertoppen ice caps to Qangâtap kûa

Under the present conditions the easternmost of the local Sukkertoppen ice caps is confluent with the Inland Ice above 900 m a.s.l. towards the south of the area. The region is largely a plateau decreasing in height towards the north from 1300 to c. 700 m a.s.l.

The ice margin deposits have been mapped from aerial photographs. They consist of frontal and marginal deposits in the valley and on wider plateaus, and of nunatak moraines around isolated minor hills or higher plateaus. From their altitudinal relationships the nunatak moraines can be seen to be in two groups, only the lowermost of which can be considered as belonging to the outer zone. This one defines a former surface of the Inland Ice, situated at c. 1200 m a.s.l. over Tasersiaq lake and at 700–800 m a.s.l. near the river Qangâtap kûa immediately south of the Umîvît branch of Søndre Strømfjord. The surface of the Inland Ice therefore must have been 400 to 800 m above the present Inland Ice margin.

8.4.5. Umîvît and Søndre Strømfjord

In Umîvît (see plate 3 and fig. 43) terraces of marine clay, up to an altitude of 40 to 50 m a.s.l., surround its inner part. Above this altitude the clayey-sandy deposits pass into gravel with rounded boulders. In the lower marine clay concretions with shells are abundant.

Above this system of marine and fluviatile terraces there are numerous moraine and kame terraces between 330 and 220 m a.s.l. The moraines descend towards the fjord, but they disappear on the steep mountain slopes before they come into contact with the marine terraces. However, it seems probable that the moraines surrounding Umîvît are older than a 45–50 m sea level.

The deposits in the interior part of Søndre Strømfjord proper are shown on the map and the profile in fig. 34. From the head of the fjord to Ravneklippen, the valley is dominated by terraces situated 15 to 30 m above the present river. The lowermost parts of these terraces consist of clay, containing shells, including *Portlandia arctica.* Near Ravneklippen, transitions from a lower clay to an upper fluviatile gravel are exposed in the profiles along the river at 25–35 m a.s.l. A sea level between 25

Fig. 43. Umîvît branch of Søndre Strømfjord. In the centre are two moraine ridges of the outer zone and at the left at sea level, a white coloured area of marine deposits (ma). Geodetic Institute's route 505 HV no. 4480 (17.6.1948). Copyright Geodetic Institute.

and 35 m a.s.l. is therefore younger than the moraines which extend along the valley out to Strømfjordshavn ("Camp Lloyd"). Terraces in the interior part of Søndre Strømfjord valley east of Ravneklippen are possibly connected with the deposits of the inner zone, described earlier.

Near Strømfjordshavn the eastern phase of the moraine system passes into a system of kame terraces around 60 m a.s.l. At this altitude, the material is still coarse, cross-bedded fluviatile gravel with rounded numerous boulders. The shallow inclination of these terraces relates their formation with a sea level at 50–60 m a.s.l.

As in other areas, several moraines were deposited during the recession of the Inland Ice from the outer zone, such as the especially well developed ones around Taserssuatsiaq ("Lake Ferguson").

8.5. Holsteinsborg and Egedesminde districts

The area is low and hilly with the present Inland Ice margin at 400–500 m a.s.l. in the south and 100–200 m a.s.l. in the north.

8.5.1. Isúnguata sermia to Ipiutârssûp nunâ

The southernmost part of the area around Isúnguata sermia was described by O. Nordenskiöld (1910, 1914). He reported that the gneisses of the area were deeply weathered even close to the ice margin, which he took to be evidence for the existence of ice free areas throughout a long period of time (Nordenskiöld 1914, p. 639). He mentioned the existence of eskers and "walls" (*i.e.* moraines) but gave no further information about them.

The aerial photographs show that between Søndre Strømfjord (Strømfjordshavn) and Isúnguata sermia glacier scattered moraine heaps rather then linear ice contact features proper are the dominant feature. Not until around Isúnguata sermia are fresher and more continous moraines met with. From their position relative to the present ice margin, it is possible that they belong to a younger phase or stage of the outer zone than the moraines further south. In the immediate vicinity of Isúnguata sermia, the deposits lie between 500 and 600 m above the present glacier front, which itself is situated near sea level. They imply that at the time of their deposition the front of Isúnguata sermia must have been 10–20 km further to the west (see fig. 44). Nothing is reported, or can be deduced from aerial photographs, about the possible relationships between marine and glacial features in the Nordre Isortoq valley west of Isúnguata sermia.

North of Isúnguata sermia, the ice margin deposits can be followed on aerial photographs as far as the northern part of Eqalungmiut nunât. In the central part of this stretch they are situated c. 700 m a.s.l. and 10–15 km from the present ice margin (cf. p. 88).

At the glacier of Inugpait qûat, the outer zone deposits descend towards the valley of Kûk. Here also, as in the valley of Nordre Isortoq, the valley sides are steep and the recent alluvial plains widespread and no remnants of older ice margin deposits are visible on the aerial photographs. A single terrace can be discerned c. 11 km north of the front of Inugpait qûat glacier. It has its surface situated between 50 and

Fig. 44. Isúnguata sermia glacier (loc. 35 of plate 2). On the left side of the picture are the moraines from the outer zone (dots). Geodetic Institute's route B 23 A–L, no. 92. Copyright Geodetic Institute.

100 m a.s.l., but the relationship between these deposits and the moraine system of the outer zone is unknown.

8.5.2. Usuĝdlûp sermia

The ice margin deposits plotted in this area on plate 3 have been mapped from aerial photographs. The morphology of the area between the lake of Tycho Brahes Sø and Usugdlûp sermia has been described by Jahn (1938). He reports the existence of abrasion terraces at 25–

35 m a.s.l. in the interior part of Arfersiorfik fjord which he correlated with a system of strandlines at the outer coast at 40–50 m a.s.l. Blue-grey shell-bearing marine clay was recorded from terraces at 35–42 m and 8–12 m a.s.l. The uppermost terrace of these was described as continuing under the present margin at Usugdlûp sermia. JAHN concluded that the 35–42 m terrace was formed during the maximum prehistoric (Holocene) retreat of the Inland Ice and the 8–12 m terrace after this. At Tycho Brahes Sø, the 8–12 m terrace is connected to the terminal moraines at the western end of the lake.

K. MILTHERS later investigated the southern side of Usugdlûp sermia (1948, pp. 393–395). In his unpublished diaries, he says that many moraine deposits are present in the inner part of the bay at Ugssuit in contrast to their absence in the outer parts of the fjord, which he suggested was due to the fast retreat of the Inland Ice there (MILTHERS 1947, 16th July). Two of the moraine ridges in the southeastern cove of Ugssuit bay are apparently older than clay terraces at the mouth of the river. FUNDER (1966) reports that the terrace is made of fossiliferous marine clay with *Mytilus edulis* amongst the shells present, and that the height of the terrace surface is 45–50 m. In addition to the moraine systems already known about in the interior parts of Ugssuit and at Usugdlûp sermia further north, FUNDER also reported the presence of another system situated c. 50 km west of Usugdlûp sermia, at the isthmus between the fjords Amitsuarssuk and Nuerssorfik at position 67°55′ N, 51°35′W.

The zone of ice margin deposits immediately south of Usugdlûp sermia can be seen on the aerial photographs to extend northwards from Ugssuit bay over Itivdliarssûp nunâ to Usugdlûp sermia. The highest point the deposits reach is c. 500 m a.s.l. near Usugdlûp sermia They slope down steeply towards the fjord and it appears that the front of the glacier lobe at that time was little more than 5 km west of the present one.

North of Usugdlûp sermia, a zone of ice margin deposits is present at c. 450 m a.s.l. It is not possible to decide from the aerial photographs if these moraines correspond to the terminal moraines observed by JAHN west of Tycho Brahes Sø, though their situation suggests that they are phases of the same stage.

Thus there is a nearly continuous zone of ice margin deposits referable to the same stage of the Inland Ice from Ugssuit in the south to Tycho Brahes Sø in the north, a distance of c. 20 km. In this area the outer zone deposits continue the trend shown by the deposits further south with the outer zone becoming lower, and flatter in section, when followed northwards from the Sukkertoppen ice caps. This means that the surface characteristic of the outer zone here is nearer to 450–550 m than to the typical 650 m.

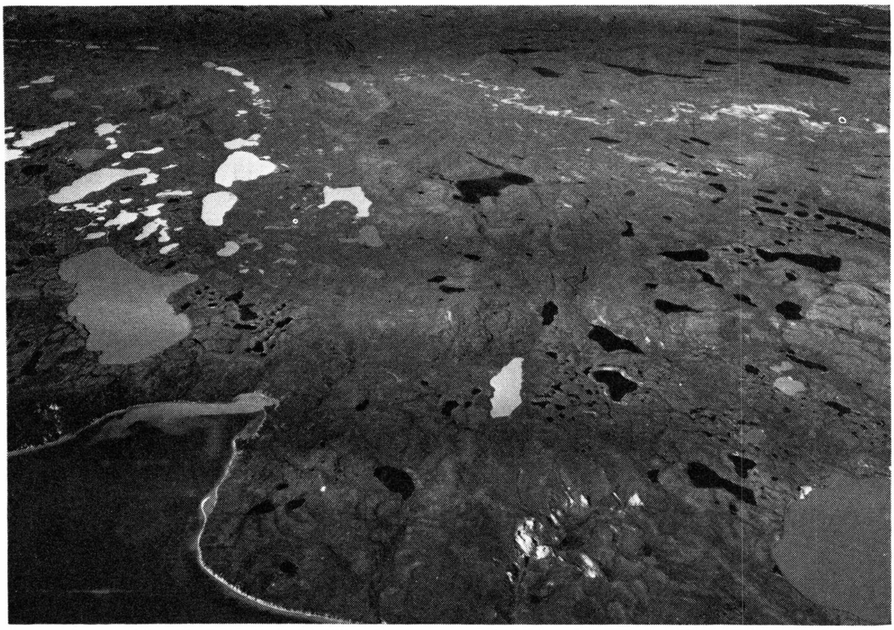

Fig. 45. Naternaq plain. In the foreground is Tasiussarssuaq bay. Geodetic Institute's route 510 F–V, no. 22 (19.8.1948). Copyright Geodetic Institute.

It may be that these deposits belong to an inner stage of this zone, such as that to be described from Påkitsoq (8.6.3 p. 121). Jahn's description, in suggesting that the deposition of the marine clay in the 35–42 m terraces predates the formation of the moraines would support this. However, Funder's observations conflict with this, and indicate that the terraces at Ugssuit are younger than the moraines around the inner part of the bay.

The classification of the moraines in the western part of Nuerssorfik fjord is also a problem. They can only be followed down to an altitude of c. 80 m a.s.l. and since parts of the system are at least older than a marine level of 50–60 m a.s.l. they may belong to an older stage of the outer zone.

8.5.3. Nordenskiölds Gletscher (Akuliarutsip sermerssua)

Whilst a great clay plain, Lersletten (Naternaq, Greenlandic: "like a parlour floor"), in the western part of the area has attracted some attention because of its size, the ice margin deposits of the area have not been investigated. Naternaq (fig. 45) must be considered to be one of the greatest areas of Quaternary marine deposits in Greenland, covering an area of c. 250 km^2. The bulk of the recent sediments consists of silts and clay covered by a thin layer of fluviatile deposits. Though

a

a

Fig. 46. Eastern branch of Tasiussarssuaq bay north of Nordenskiölds Gletscher, seen from the south. In the foreground (a–a) are nunatak moraines and in the background (north of Tasiussarssuaq), marginal moraines. Geodetic Institute's route 512 C–N, no. 52 (19.8.1948). Copyright Geodetic Institute.

the surface varies from 50 to 100 m, the relief is subdued, hence its Greenlandic name. RINK (1852, p. 60) states that in the northernmost parts of the area the clay contains concretions with shells and casts of the fish *Mallotus arcticus* and laminae of coal or charcoal-like material. More detailed description of the fauna of the clay is given by HARDER *et al.* (1949, pp. 66–67) for an area around Sarpiussat on the north side of the plain. There its surface is at 50 m a.s.l. and the marine deposits are in places underlain by moraine. Most of the clays there were assigned by them to the *Portlandia* clay. MILTHERS (1947, 23rd July) also found *Portlandia* clay in deposits at 100 m a.s.l. on the southern side of the plain. The eastern parts of the plain were visited by A. NORDENSKIÖLD (1871, p. 49, 1886, p. 172), but all that can be said from his description is that marine clays are the principle sediments there also.

The morphology of the plain suggests that it originated as an alluvial plain graded to sea levels between 100 and 50 m a.s.l., and that the lower height is related to the last phase of its formation. Aerial photographs show large parts of the surface of the plain to be covered by kettle holes and in the east, at Tasiussarssuaq bay, are moraines which can be correlated with the c. 50 m alluvial plain. At this time a glacier lobe must have covered the present Tasiussarssuaq bay.

Contemporaneously with the formation of the moraines, meltwater drainage channels were formed, draining northwards through Eqaluliata itivnera and across Naternaq via Kûgssûp qîngua. The drainage channels are somewhat lower than the surface of the plain and the sea level must have been below c. 50 m a.s.l. However, the form of the channels suggests it was not lower than 40 metres.

All the ice margin deposits around Tasiussarssuaq and Nordenskiölds Gletscher (see plate 3 and fig. 46) seem to have been formed by the same stage of the Inland Ice and may be a continuation of the main outer zone deposits further south. Here, as in the adjacent area to the south, the surface characteristic of the outer zone is only 400–450 m, though the zone extends c. 20 km further west than the present glacier front. It should be pointed out that in historical time also, this sector of the Inland Ice margin has thinned less than the rest of the Inland Ice in this century (see p. 42–45).

From the descriptions of earlier writers it seems that the *Portlandia* clay was deposited whilst the sea level was c. 80 m and possibly 100 m above its present level. Since the surface of the plain is covered with kettle holes, the Inland Ice margin must have subsequently advanced over the clays of the Naternaq plain, forming dead ice terrain as it again retreated and thinned. This is older than the readvance which was responsible for the formation of the moraines around Tasiussarssuaq.

8.6. Disko Bugt

In contrast to the regions further south the deglaciated coastal stretch is only c. 30 km wide. The highest parts are only 500–700 m a.s.l. and at several places fjords cut right across the area to the present Inland Ice margin.

8.6.1. Orpigsôq – Christianshåb area

Ice margin deposits of the outer zone are present in three well delimited areas, Orpigsôq, Kangersuneq and Laksebugt.

8.6.1.1. Orpigsôq

The richly fossiliferous deposits of this locality have been described by ENGELL (1910, p. 234), HARDER *et al.* (1949, pp. 15–50) and LAURSEN (1950, pp. 37–46). HARDER *et al.* (p. 46) and LAURSEN (p. 43) mention the existence of a terrace surrounding the lake of Orpigsûp tasia at 40 m a.s.l. (*i.e.* 4 m above the lake) which seems to be overlain by a minor moraine on its northwestern side. The terrace is reported to consist of glacio-fluviatile sand overlain by *Portlandia* clay and "shelly gravel", the last of which forms the surface of the terrace. According to LAURSEN's surveyed sections the beds dip towards the west. HARDER *et al.* also mention a higher terrace of 56 m a.s.l. at the lake whilst LAURSEN describes probable marine deposits at 93 m a.s.l. from a valley, c. 5 km to the west of Ilulialik lake. However, the latter are unfossiliferous clays and may be of glacio-lacustrine origin.

On aerial photographs of the area, ice margin deposits can be seen between 100 and 300 m a.s.l. which indicate that a lobe has completely filled Orpigsôq fjord as far to Kangersuneq (see plate 3). From the published descriptions mentioned above, it seems that they must have been formed whilst the sea level was higher than 40 m, and possibly 56 m, above the present. The moraine mentioned by HARDER *et al.* must be younger than the main system, but they have not been observed on the aerial photographs.

8.6.1.2. Kangersuneq

This area was visited in 1906 by HARDER (HARDER *et al.* 1949, pp. 50–56). He mentions that the head of the fjord at Kangersuneq is separated from the interior by a 200 m high rock wall, at the base of which are terraces cut in moraine material at altitudes of 30 and 50 m a.s.l. Moraine was also found at Nánikut (see fig. 47) on the northwestern side of the fjord. For the area around Nûgsutap kûa valley they concluded that a clear transition existed from basal moraine deposits into marine clays and finally into glacio-fluviatile deposits.

During a short visit to the area by the author in 1963, a moraine ridge was found on the southern side of the Serfarssuit peninsula, which must have been formed by a glacier lobe filling the interior part of Kangersuneq. The moraine, which consisted of boulder rich gravel, sloped down towards the west, from 150 to 20 m a.s.l. The lowermost section, 20–50 m a.s.l. appeared to be water sorted moraine heaps. The moraine is not present to the east, where the sides of the fjord are steep but a possible northern continuation may be sought in the belt of ice margin deposits east of the highest parts of the dissected plateau (*i.e.* the points of 446, 527, 530 and 572 in Fig. 47). The overall appearance of this belt is of an interlobate terrain of widespread moraine hills which only in the valleys have the more definite form of moraine ridges.

Fig. 47. Sketch map of the area between Christianshåb and Jakobshavns Isfjord. Based on Geodetic Institute's map sheets 1:250,000 68 V2, Christianshåb and 69 V2, Jakobshavn. Copyright Geodetic Institute. For phase, read phase or stage. By permission of the Geodetic Institute.

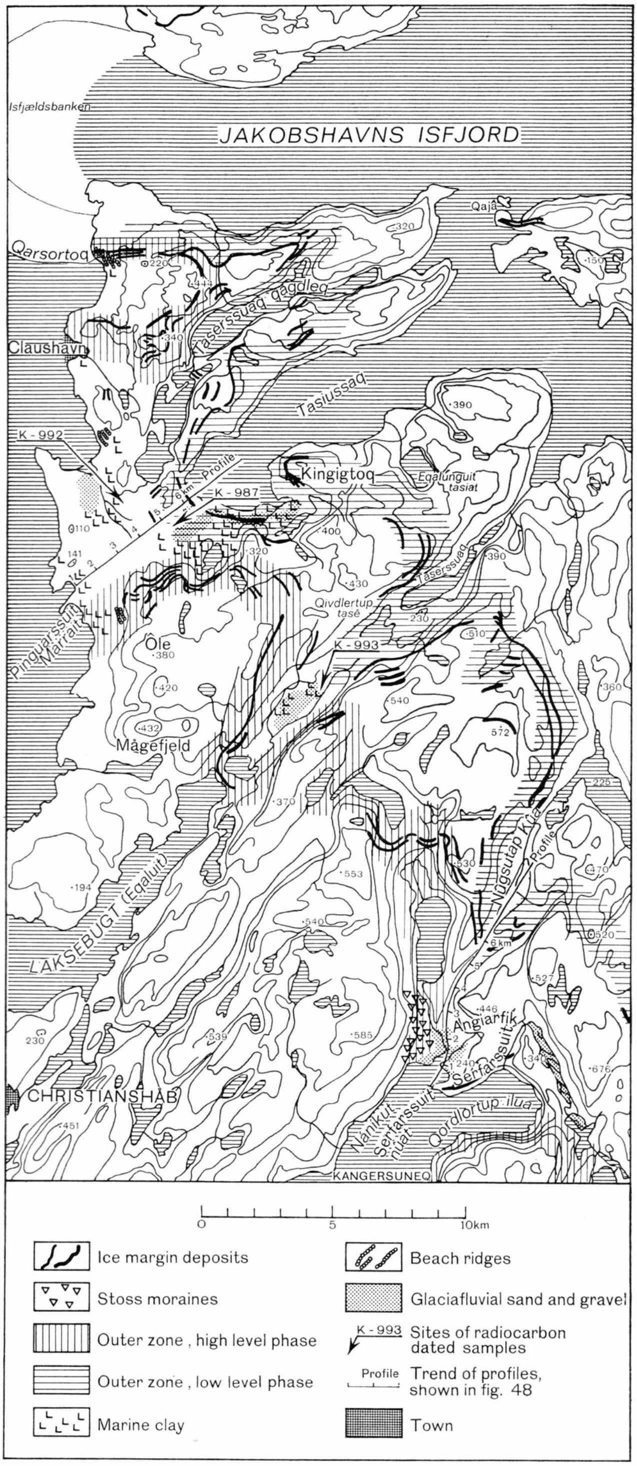
Isfjældsbanken
JAKOBSHAVNS ISFJORD
Qajâ
Qarsortoq
Claushavn
Tasersuaq qagdleq
Tasiussaq
K - 992
K - 987
Profile
6 km
Kingigtoq
Eqalunguit tasiat
Tasersuaq
Qivdlertup tasê
K - 993
Pinguarssuit Marrait
Ôle
Mågefjeld
LAKSEBUGT (Eqaluit)
Nûgsutap kûa
Profile
6 km
Angiarfik
Sêrfarssuit
Nânikut Sêrfarssuit nûat
Qordlortup ilua
CHRISTIANSHÅB
KANGERSUNEQ
0 5 10km
Ice margin deposits
Stoss moraines
Outer zone, high level phase
Outer zone, low level phase
Marine clay
Beach ridges
Glaciafluvial sand and gravel
K - 993 Sites of radiocarbon dated samples
Profile Trend of profiles, shown in fig. 48
Town

Fig. 47.

The best developed moraine ridges are seen in the Nûgsutap kûa valley where the moraines are connected distally with kame and valley terraces graded to a marine terrace at c. 50 m a.s.l. The gravel which covers the marine clays near the fjord probably represents an outwash deposit of this stage.

The moraines at Nánîkut, described by HARDER *et al.*, are stoss moraines which from their situation must be older than those at Serfarssuit. Both however, can be considered as belonging to the outer zone of ice margin deposits. These moraines imply by extrapolation that the Δh of the zone at the Inland Ice margin 10 km further east, is 600–800 m.

8.6.1.3. Eqaluit

The main topographic feature of the area is a 15 km long valley, connecting Eqaluit fjord in the west with Tasiussaq fjord in the east, which dissects a plateau, here 500–600 m high. The drainage divide in the valley is at the western end of the lake of Taserssuaq at a height of c. 100 m a.s.l. Only the western half of the valley has been investigated by the author.

West of Qivdlertup tasê coarse fluviatile deposits and gravelly ground moraine were seen at altitudes between 60 and 100 m a.s.l., the fluviatile deposits forming a terrace which sinks from 60 m a.s.l. at Qivdlertup tasê to c. 45 m a.s.l. approximately 4 km further west. Profiles in the terrace between 45 and 60 m a.s.l. have at their base marine clays, rich in shells of *Mya truncata* and *Macoma calcarea*, which towards the top become coarser and more laminated. These are overlain by glaciofluviatile gravel and boulders. The uppermost part of the clay proper is rich in shell bearing concretions but fossils in the lower clays are scarce though two examples of *Portlandia arctica* were found. The radiocarbon age of the shells of the upper part of the clay, from 50–55 m a.s.l., is 7650 ± 140 years B.P. (K-993).

This terrace is clearly an outwash plain associated with a moraine system which surrounds Qivdlertup tasê and Taserssuaq lake. By its nature, it relates this moraine system with a minimum sea level of 40–45 m a.s.l., the age of which should be younger than 7650 years B.P. A terminal moraine which lies on a rock bar at the entrance to the valley, separating it from Eqaluit fjord is cut by terraces at 50, 55 and 70 m a.s.l. and must be older than the other moraines.

Both the western and eastern moraine systems can be followed on the aerial photographs along both sides of Eqaluit valley, but become more difficult to trace on the plateau above, where they become irregular interlobate moraines.

8.6.2. Claushavn – Jakobshavn area

The area consists of a plain between Pinguarssuit, Marrait and Tasiussaq which is separated by a series of hills from Jakobshavns Isfjord to the north. Only the marine deposits of the plain have been previously described (GIESECKE 1910, p. 97, NORDENSKIÖLD 1871, p. 48, ENGELL 1910, pp. 232–233 and LAURSEN 1950, pp. 25–37).

These investigations, supplemented by a visit by the author in 1963 reveal that most of this plain is formed of *Portlandia* clay which is most fossiliferous towards the top. Overlying the clays are fluviatile gravels, which form the even surface of the plain at c. 50 m a.s.l. Besides the gullies formed by present day river erosion, the most striking morphological features of this plain are a distinct moraine ridge along the coast of Tasiussaq, and an area of well developed kettle hole topography (see fig. 47 and 49). Abandoned braided stream channels, now partly occupied by lakes, spread out from the moraine.

This moraine system is a part of an extensive system which constitutes a low level or eastern "phase" in the deposition of the zone. Towards the south it can be followed uphill to the area around point 430 (fig. 47), where it fades into an area of interlobate moraine terrain, but its connection with the inner moraine system in Eqaluit valley further south (see 8.6.1 p. 114) is obvious. Towards the north, the same moraine system can be followed along the northeastern sides of Tasiussaq as far as Jakobshavns Isfjord. A branch of the Inland Ice at this stage must have descended also into the lake of Taserssuaq qagdleq. This segment of the moraine system also developed an interlobate character towards higher altitudes.

A second moraine system exists in the area, which by its situation is both higher and further west than the other, constituting an older "high level phase". It is clearly a continuation of the moraine system found at the head of Eqaluit. During its formation the Inland Ice must have covered the plain and reached the coast. Well formed beach ridges developed on the moraine near Marrait at 65 and 70 m a.s.l. give a minimum age for this phase. These beaches lie topographically above the *Portlandia* clay of the plain which does not extend higher than 50 m a.s.l. Towards the north, between the plain and Claushavn, the "high level phase" is represented only by patches of moraine. In the western part of this area also are well developed beach ridges between 55 and 70 m a.s.l. which have the same relationship to the *Portlandia* clay as the beach ridges at the plain. During the deposition of this phase, the only major ice free area appears to have been between Claushavn and Qarsortoq on the lee side of a 300–450 m high hill mass which screened it from the lobe in Taserssuaq qagdleq. The moraines of the

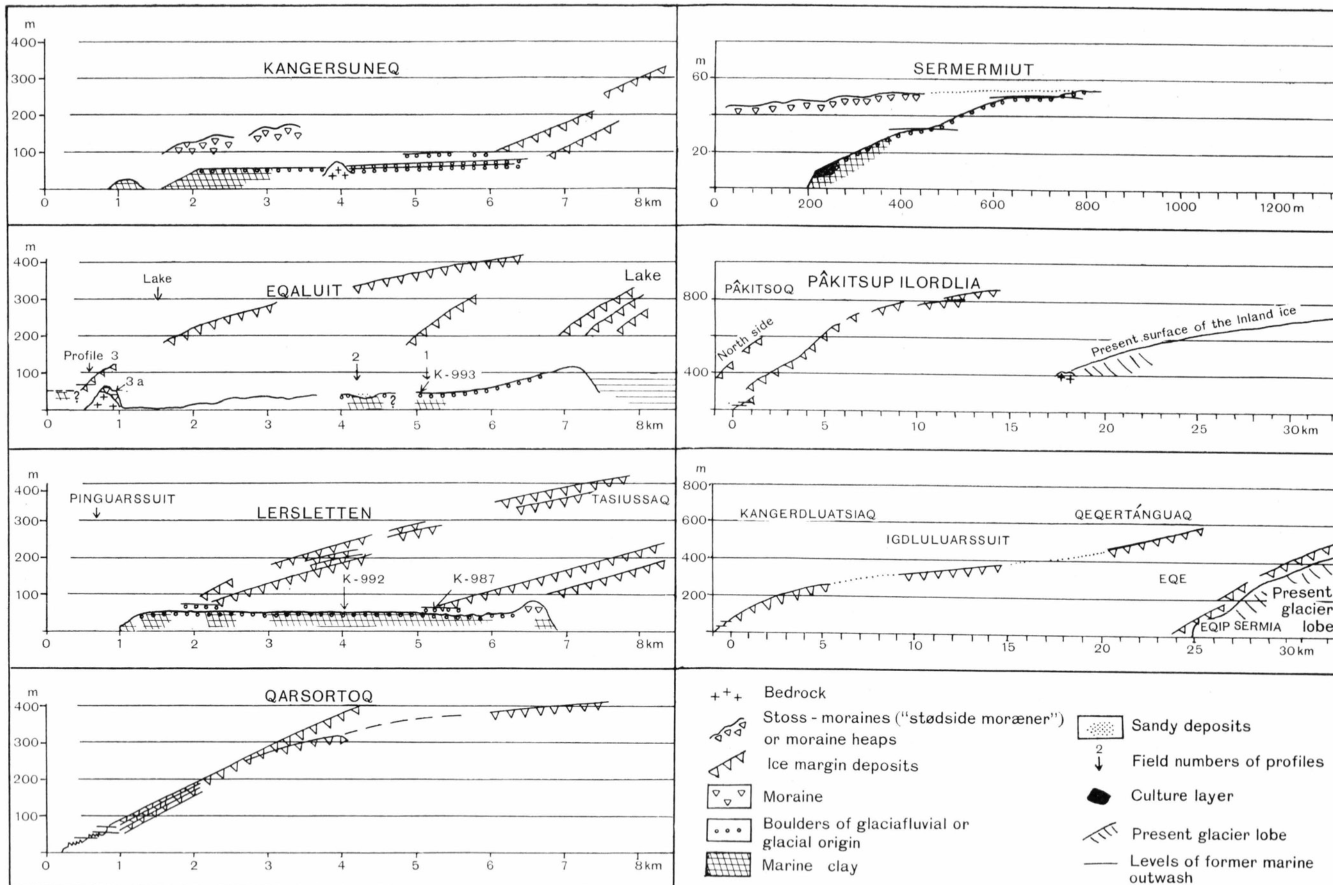

Fig. 48. Profile through the outer zone deposits in Disko Bugt.

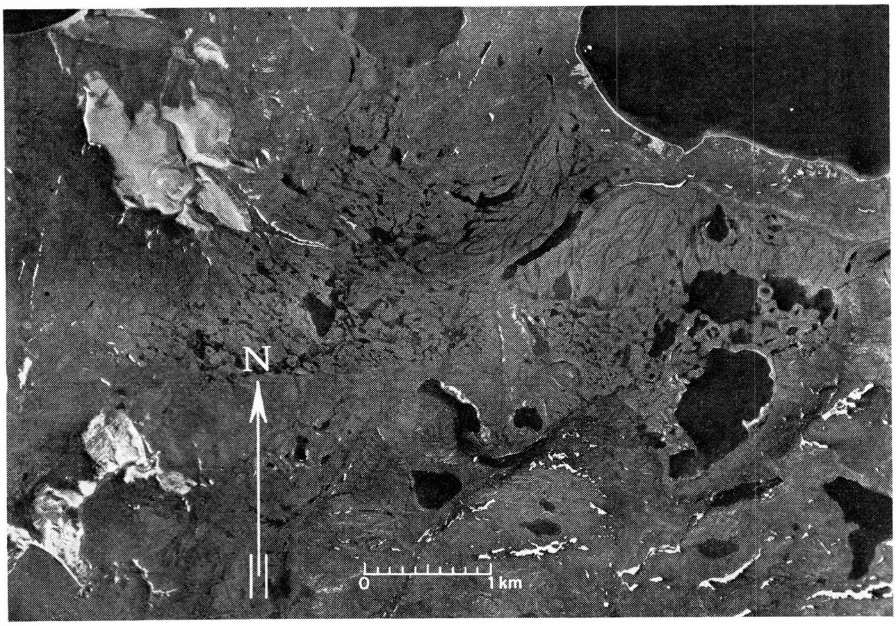

Fig. 49. The clay plain (Lersletten) south of Claushavn. Aerial photograph, Geodetic Institute's route A67/151, no. 68 (3.7.1953). Copyright Geodetic Institute.

outer system lie near the top of this, from where they descend steeply to the Claushavn and Qarsortoq areas. At Claushavn, the moraines, consisting of gravel with numerous rounded boulders, end at a kame terrace at 70 m a.s.l., just north of the village.

Around Qarsortoq, moraines from "the high level phase" and "the low level phase" join up into a system of parallel multiple moraines. The largest and highest situated gravelly moraine is cut by a terrace between 60 and 65 m a.s.l. The next moraine towards the north consists of the same material, but unlike the outermost moraine it keeps an unmodified sharp crest down to 50–55 m a.s.l., where it is cut by the highest of a series of beach ridges. These extend down to near the present sea level and cut across all the Qarsortoq moraines. The following two, at some places three, inner moraines have been cut by the beach ridges up to c. 55 m a.s.l. but are lithologically distinct from the outer moraines, being of angular boulders with little interstitial matrix. This change in the character of the moraines may be explained by there having been a reduction of the amount of reworked fluviatile material available between the earlier and the later glacial advances.

In addition to the lower beach ridges, Laursen (1950, pp. 18–21) mentions the existence of marine terraces at Qarsortoq at 145, 160, 165, 170 and 174 m a.s.l. and he puts the marine limit of the area at c.

185 m a.s.l. However, a close inspection of these features suggests that they are probably kame terraces.

The form of the Isfjældsbanken, the submerged ridge at the mouth of Jakobshavns Isfjord, suggests that the deposits of both "the low and the high level phases" contribute to it. On the northern side of the fjord there is a system of large moraine ridges but it is not possible to identify the two "phases" there. These ice contact features can be followed northwards almost continuously to Påkitsoq fjord. In the intervening stretch they reach altitudes between 500 and 600 m a.s.l. and it is clear that the Δh of the Inland Ice margin there must have been between 600 and 800 m.

Near the town of Jakobshavn, these moraines have been reworked producing "washed boulder moraines" up to 60–70 m a.s.l. (WEIDICK 1965). At the nearby locality of Sermermiut, the deposits surround an area of *Portlandia* clay which is overlain by the cultural deposits of the Sarqaq, Dorset and Thule cultures. At this place, the oldest Sarqaq layers are dated to 3000–3500 B.P. (TAUBER 1960b, pp. 18–19, 1964, pp. 221–222, LARSEN and MELDGAARD 1958, pp. 9–31, FREDSKILD 1967, pp. 39–40).

For the area as a whole, it seems that the "high level phase" of deposits along the outer margin of land was contemporaneous with a sea level around 70–80 m a.s.l., which may predate the deposition of *Portlandia* clay. Subsequent to this, the Inland Ice margin retreated beyond the clay plain and later readvanced to form the "low level phase", which according to the evidence at Tasiussaq was contemporaneous with a sea level of 35–50 m a.s.l.

Shells of *Mya truncata*, from c. 40 m a.s.l. in the uppermost part of the clay at Tasiussaq, were radiocarbon dated as 7110 ± 140 years B.P. (K-992). On the stratigraphic evidence the moraine on the surface of the plain should be younger than this. However, the oldest sediments in a lake which should post-date the moraine gave an earlier date, 7850 ± 190 years B.P. (K-987, KELLY personal communication). The reason for the anomaly is not obvious and only more detailed investigations will indicate whether the error is due to contamination of the material or to complication in the stratigraphy. However, in general, the dates fit well with other values for the 40 m level, given in fig. 2. An age of 7500 to 8500 years B.P. is therefore still regarded as a good estimate of the age of the "low level phase" moraines.

8.6.3. Påkitsoq area

The area stretches from Påkitsup nunâ around Niaqornaq in the south to Kangerdluarssuk fjord in the north. As the name Påkitsoq implies in Greenlandic the innermost part of this fjord is separated by

Fig. 50. Taserssuaq lake (a–a) between Jakobshavns Isfjord and Qíngua kujatdleq, Pâkitsup ilordlia, seen from the west. In the foreground are moraines of the outer zone. Geodetic Institute's route 518 A–Ø no. 1474 (18.7.1949). Copyright Geodetic Institute.

a narrow passage from the outer parts. The passage is described by RINK (1857, I, p. 125) but it was WHYMPER and BROWN (BROWN 1872, p. 173) who first seem to have realised the existence of a terminal moraine at this constriction. This was described later by HAMMER (1889, pp. 13–14) and SYLOW (1889, p. 26).

There have also been earlier reports of marine terraces up to 45 m a.s.l. containing *Portlandia* clay at Niaqornaq (STEENSTRUP 1883b, pp. 231, 235). A visit here in 1961 showed a continous series of beach ridges between 37 and 48 m a.s.l. and terrace notches at 12 and 60 m a.s.l.

At Pâkitsoq the surface of the terminal moraines was found to be of rounded boulders without matrix, but exposures in the two stream channels which cut through the moraine show that in its inner parts there is a matrix of sand. Well formed terraces at 25 and 30–35 m a.s.l. are cut in this moraine. Above this altitude the moraine is absent over a distance of c. 1 km and is not found again until c. 100 m a.s.l.

At Taserssuaq (south of Qingua kujatdleq, see fig. 50) lake, where the moraines reach c. 600 m a.s.l., their highest point between Jakobs-

Fig. 51. Inner part of the Kangerdluarssuk fjord, seen from the west. In the foreground of the photograph are marginal moraines, reaching the fjord (white dots). (a–a); Qíngua avangnardleq (loc. 73 in plate 2). Geodetic Institute's route 518 A–Ø no. 1485 (18.7.1949). Copyright Geodetic Institute.

havn and Pâkitsoq, the principal moraines are two closely parallel ridges of considerable size, rising to over 50 m above the surrounding bedrock over stretches of more than 100 m. Further east, around the shores of Taserssuaq, are a series of multiple, minor recessional moraines.

An equivalent continuous system of deposits was not found on the northern side of Pâkitsoq and only isolated features were seen between 50 and 200 m a.s.l. However, further northeastwards, a well developed system of moraines can be seen on the aerial photographs, at 600–700 m a.s.l. and at a distance of 3–4 km from the present Inland Ice margin, which is now at 400–500 m a.s.l. According to the definition of the outer zone, at this altitude it should have a surface characteristic of c. 500 m. The figure obtained by the extrapolation of the former ice surface from this moraine system over the present Inland Ice margin agrees with this.

The position of these ice margin deposits suggests that a branch of the Inland Ice would have extended into Kangerdluarssuk fjord, and deposits can be seen on the aerial photographs descending towards the fjord (see fig. 51), though HAMMER's description of this (HAMMER 1889, p. 13) gives no indication of their presence. Within Kangerdluarssuk fjord, the ice margin deposits show a clear division into high and low level stages separated horizontally by 2–3 km and vertically by 100 m at 600 m a.s.l. and 150 m at 300 m a.s.l.

8.6.4. Arveprinsens Ejland and Eqip sermia

The area comprises the stretch between Kangerdluarssuk fjord and Torssukátak ice fjord. Most of the area is a ramified fjord complex, the central parts of which, in the basin between Igdluluarssuit and Eqe, are known to have a level bottom at a depth of 400 m (HØPNER PETERSEN 1964, p. 33). Marine terraces are reported from various places around the fjords, at 2 and 16 m a.s.l. at Igdluluarssuit, between 5 and 58 m a.s.l. at Atâ, and between 9 and 81.5 m a.s.l. around the great lake north of Atâ (LAURSEN 1950, pp. 15, 16) and possibly at c. 25 m a.s.l. at de Quervains Havn near Eqip sermia (BAUER 1955c, p. 96, cf. p. 90).

Along the coast of Arveprinsens Ejland, and north of Atâ, ice margin deposits were found at c. 200 m a.s.l. They were formed of poorly sorted gravel with large rounded boulders, which included boulders of white and yellow sandstone and quartzite. This moraine could be traced towards the north as a system of single, double or triple ridges which ended at sea level in Kangerdluatsiaq between Torssukátak and Igdluluarssuit from where they were described by ENGELL (1910 pp. 190–191). There, marine terraces are cut in the moraine up to c. 35 m a.s.l. and from there up to c. 80 m a.s.l. the moraines are represented only by irregular boulder heaps. This last character may be due to resorting by the sea, as at Sermermiut and Pâkitsoq. Above 80 m a.s.l. the moraines exhibit their normal ridge form. The islands in Kangerdluatsiaq are also moraine covered and the uppermost level of re-sorting in these moraines was estimated visually from Arveprinsens Ejland to be situated between 30 and 50 m a.s.l. Beach ridges were also observed in the distance on Talerua near Igdluluarssuit at c. 30 m a.s.l. No continuation of the Atâ moraines north of Arveprinsens Ejland were observed.

In the south, a moraine system can be seen on aerial photographs along the shore at Angnertussoq which is clearly a continuation of the high level stage of the moraines in Kangerdluarssuk. These moraines decrease in altitude from east to west, from 600 to 200 m a.s.l., but do not continue further to the west because of the steep sides of the fjord. Their western continuation must be found in the moraines on Arveprinsens Ejland, at Atâ and west of Angnertussoq, which join the moraine systems of Angnertussoq with those of Arsivik north of Atâ. The ice of the low level stage of the moraine systems at Kangerdluarssuk seems to have had a lobe through the Eqip kûgssua valley, south of Eqip sermia, but its northerly continuation is not clear.

In the interior parts of the area MERCANTON (DE QUERVAIN and MERCANTON 1925, p. 237) observed an upper moraine level situated at c. 700 m a.s.l. on the nunatak "Ilulialik" (*i.e.* Qingârssuaq at Qapiarfik in plate 3).

In general, the deposits of the outer zone are well developed in the southern part of the area, where they can be correlated with sea levels between 35 and 80 m a.s.l. However, none are known north of Arveprinsens Ejland–Igdluluarssuit and therefore nothing can be said about the extent of this zone in Torssukátak fjord.

8.6.5. Nûgssuaq

The area comprises the stretch from Torssukátak fjord in the south to the southeastern branch of Umanak fjord in the north ("Qarajaks Isfjord"). The interior part of the peninsula is formed by deeply dissected plateaus which increase in height from 500–700 m a.s.l. near the Inland Ice margin, to over 1000 m a.s.l. further towards the west. In the area around Boyes Sø and further west, the plateaus are covered with ice caps.

In 1961, the north coast of Torssukátak from Ikorfat to Sermeq avangnardleq was visited. Besides a few ice margin deposits (see plate 3) terraces up to 10–15 m were found along the shores from Qeqertarssuk (20 km east of Qeqertaq) to near the Inland Ice.

Around Boyes Sø and Pangniligarniarfik (see fig. 52), ice margin deposits of great extent have been observed on aerial photographs, situated at c. 600 m a.s.l. in the east and c. 200 m a.s.l. in the west. The superficial appearance of the ice margin which formed these deposits can be reconstructed from the marginal and frontal moraines around the lakes and the nunatak moraines around the mountains. As the profile shows (fig. 42), still existing local ice caps are situated close to these moraines. This relationship, as at Majorqaq area (8.4.3, p. 101), indicates that the glaciation limit during the deposition of the moraine system cannot have been many hundreds of metres below the present one. The maximum possible value for the depression of the glaciation limit is estimated to be 400–500 m.

The ice margin deposits around the western end of Boyes Sø and in Qôrorssuaq valley are difficult to interpret without more detailed investigations. It is probable that even a slight depression of the glaciation limit in the area would result in a complex fusion of lobes from the Inland Ice with lobes from the local glaciations. Furthermore, slight varia-

Fig. 52. In the foreground, Boyes Sø (a–a) and "Pangniligarniarfik lake" (b–b), seen from the east. Geodetic Institute's route 518 A–V, no. 7629 (18.7.1949). Copyright Geodetic Institute.

tions of the altitude of the glaciation limit would result here in great variations in the direction of flow of the ice lobes, further complicating the form of the ice contact features.

At the seaward end of the Qôrorssuaq valley is a great terminal moraine and its associated proglacial fluvial plain, which is situated 60–70 m a.s.l. The alluvial plain becomes almost horizontal at 60 m a.s.l. and former spillways on the plain's eastern side are cut by a cliff remnant at 55 m a.s.l. It is therefore probable that the formation of the terminal moraine was contemporaneous with a sea level of 50–55 m a.s.l. Large terminal moraines from former local glaciations just east of the above mentioned one (see fig. 24 and 30b) are cut by a terrace at 60 m a.s.l. It is thus likely that these moraines are contemporaneous. The glaciation limit of the local glaciations at this stage can be estimated to have been between 200 and 500 m lower than the present one.

The terminal moraine of the Inland Ice continues towards the interior part of Qôrorssuaq valley as extensive block moraines, described earlier, but a connection between the glaciation of the Qôrorssuaq valley and that of Boyes Sø cannot be established from the evidence available. If such a connection exists, the terminal moraines in the interior parts of the Qôrorssuaq valley must be considered as recessional moraines formed during the disintegration of a ramified system of glaciers, an "Eisstromnetz" in the sense of KLEBELSBERG (1948, pp. 202–204), which formerly filled all the valleys of the interior part of Nûgssuaq.

In the area around Eqaluit qáqât on the northern coast of Nûgssuaq are moraine systems at 300–400 m a.s.l. which indicate a minor expansion of the local ice caps. This, on the basis of the argument given above, implies a greater expansion of the Inland Ice which probably filled Eqaluit with a major glacier lobe. The description of this valley (DRYGALSKI 1897, p. 118, BARTON 1897, p. 234) suggests that there is no possibility there of dating the moraines of this lobe by relating them to former sea levels. BARTON, in his description of Eqaluit, reports the presence of two lower lying marginal moraines on the south side of the valley, which may be interpreted as recessional moraines from the stage filling the valley.

Further north, in the Umanak district, the altitude of the plateau increases. It is therefore possible that a slight decrease of the present glaciation limit here would result in the formation of extremely complex system of confluent glaciers in the glaciation of the interior parts of the district. Extensive ice margin deposits (HENDERSON personal communication) in the outer parts of the ice fjords Kangerdluk and Umiámako may possibly belong to the outer zone because of their altitudinal relationship with the present Inland Ice surface. However, nothing is known about their relationship with former sea levels. A possible eastern continuation of these moraines gives a surface characteristic of 800–1000 m.

9. INLAND ICE DEPOSITS – NUNATAK ZONE

Nunatak moraines which characterise a level of ice margin deposits situated essentially higher than those of the outer zone have been observed on aerial photographs in two areas, the first between the Søndre Strømfjord and the Inland Ice margin to the south, and the second north of Søndre Strømfjord, between Pingup sagdlia and the head of Nordre Isortoq (see plate 3). The ice margin deposits are situated around the highest parts of the plateaus, especially at their eastern margins, and their inclination indicates that they were formed by ice coming from the east, *i.e.* the Inland Ice. The deposits form a zone, 500 to 200 m high at 1000 m a.s.l. and 400 to 500 m near sea level.

The situation of these deposits at about 1000 m a.s.l. at localities where the present glaciation limit is 1300 to 1400 m a.s.l. puts the depression of the glaciation limit at a little more than 300–500 m.

There are no earlier detailed descriptions of these deposits, though Belknap (1941, p. 220) mentions the existence of a belt of moraine in the interior parts of the above mentioned fjords in the Holsteinsborg district, and he compares this area with the deposits of what is here called the outer zone further east. The exact location of this belt is not given, but may have referred to a large area of dead ice terrain and ground moraine, which can be seen on aerial photographs immediately east of the nunatak deposits in the Holsteinsborg area. The geographical situation of these deposits seems to indicate their formation by dead ice remnants left by the Inland Ice on its retreat from the nunatak moraines. Jensen and Kornerup in their descriptions of this area (Jensen 1889, pp. 49–64, Kornerup 1881, pp. 181–194) point out its morphological resemblance to the Danish undulating till landscape. The esker of "Arssalik" is situated in this zone.

The deposits south of Søndre Strømfjord have been observed only on aerial photographs, and their distribution is shown on plate 3. By their altitudinal situation, they must be referred to the nunatak zone.

North of Søndre Strømfjord, the interior branches of the fjords of Ikertôq and Amerdloq were personally inspected in 1965. At the southern

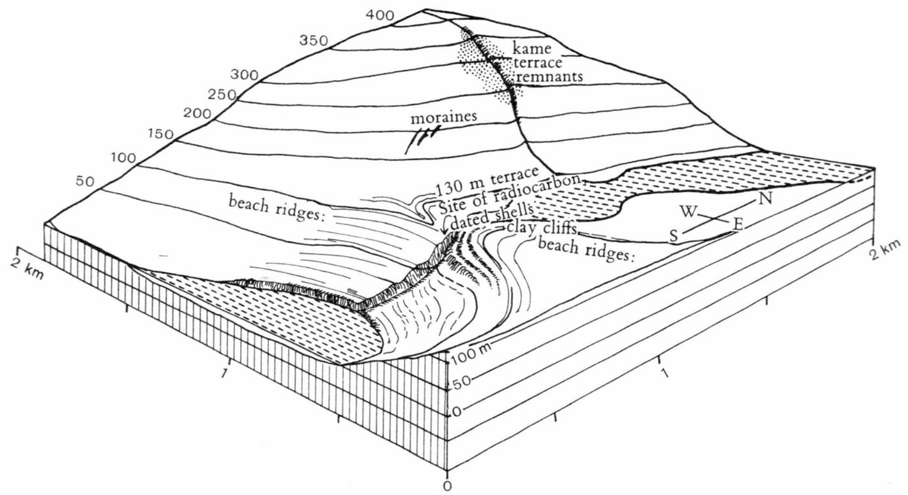

Fig. 53a.

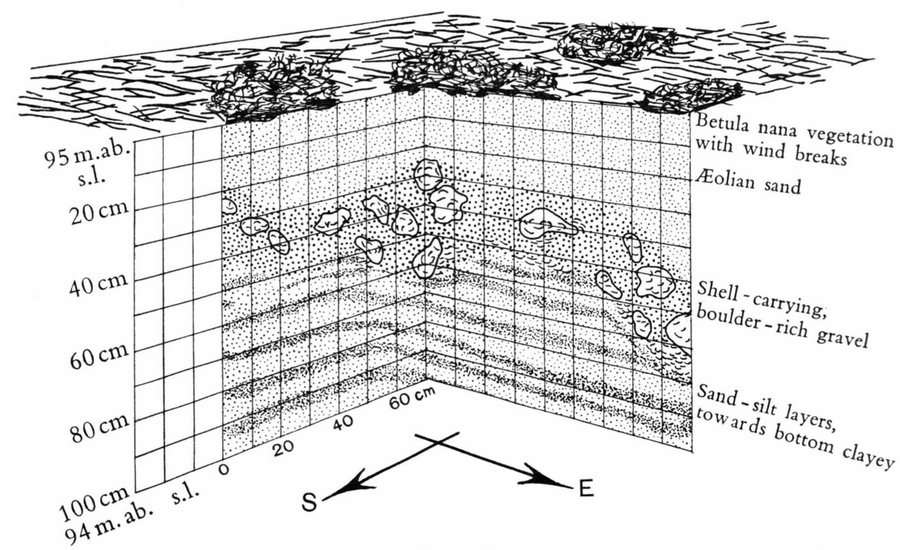

Fig. 53b.

Fig. 53a. Situation of the marine deposits in the interior part of Avatdleq, Ikertoq fjord. 53b. Section at 95 m a.s.l. in the beach ridges in the inner part of Avatdleq, Ikertoq. Position of section in fig. 53a as "site of radiocarbon dated shells".

branch of Avatdleq ice margin deposits were found to consist of gravel with numerous rounded boulders. In addition to the main system isolated moraines and terraces were present on either side of it, whilst towards the east, the system merged into a chaotic interlobate moraine terrain.

Towards the coast, it became a series of moraine hills which down in the valley were cut by a series of beach ridges between 130 and 110 m a.s.l. (see fig. 53a). Between 110 and 100 m a.s.l. the valley is occupied by a terrace, which can be seen in the cliffs at the southernmost lake in the valley to consist of unfossiliferous clays. The seaward end of this terrace is a slope covered by a continous series of beach ridges extending down to the present sea level, as shown in fig. 53a. In exposures cut in the slope by a rivulet from the valley, shell bearing sands and gravels were found, at 0–10 m and 95 m a.s.l., beneath a cover of beach gravel and at places also of dune sand. As the profile dug into the uppermost deposit showed (fig. 53b), the shells can hardly have been derived from older higher deposits and it is probable that they represent a littoral deposit, overlain by wind-blown sand and underlain by silts and sands.

Though the highest beach ridges, at 130–110 m a.s.l., are unfossiliferous their distribution suggests they are of marine origin. The moraine must therefore be older than a sea level of this altitude, 130 m a.s.l., but since the littoral features cannot be followed further into the valley it seems possible that they are not much older.

In the Akugdleq branch also, terraces were seen at 100–120 m a.s.l., but again marine fossils were absent. The terraces there are formed of gravel with numerous rounded boulders and they may be kame terraces. Not until a level of c. 50 m a.s.l. is reached are clearly developed beach ridges seen, built up of gravel and cobbles. Below 40 m there are fossiliferous marine sands and clays, which infill most of the valley out to the fjord. No continuation of the moraine system of Avatdleq fjord was seen here.

In the northernmost branch, Maligiaq, a system of moraine ridges has been observed on the northern side of Taserssuaq valley at altitudes between 600 and 400 m a.s.l. The system ends in a series of terminal moraines across the lakes northeast of the head of Maligiaq near a mountain called Iluliumanerssûp portornga. None of the moraines of this system can be seen to reach Maligiaq fjord, but the altitudinal position of the deposits indicates that a branch of the ice must have reached the inner part of Maligiaq. Possible marine terraces between 130 and 140 m may have developed after the ensuing retreat of this lobe.

Terraces, from 55 to 110 m a.s.l., have also been observed further west at Utorqait in the interior of Amerdloq fjord.

The moraine systems observed in Avatdleq and Maligiaq must belong to a younger stage of the nunatak zone, the main moraines of which (shown here in fig. 54) are situated 400–500 m higher in the area. The terminal moraines of the highest stage have not been found, but they can be expected to be found 30–50 km further west. The numerous moraines between this high stage and the lowermost, youngest stage,

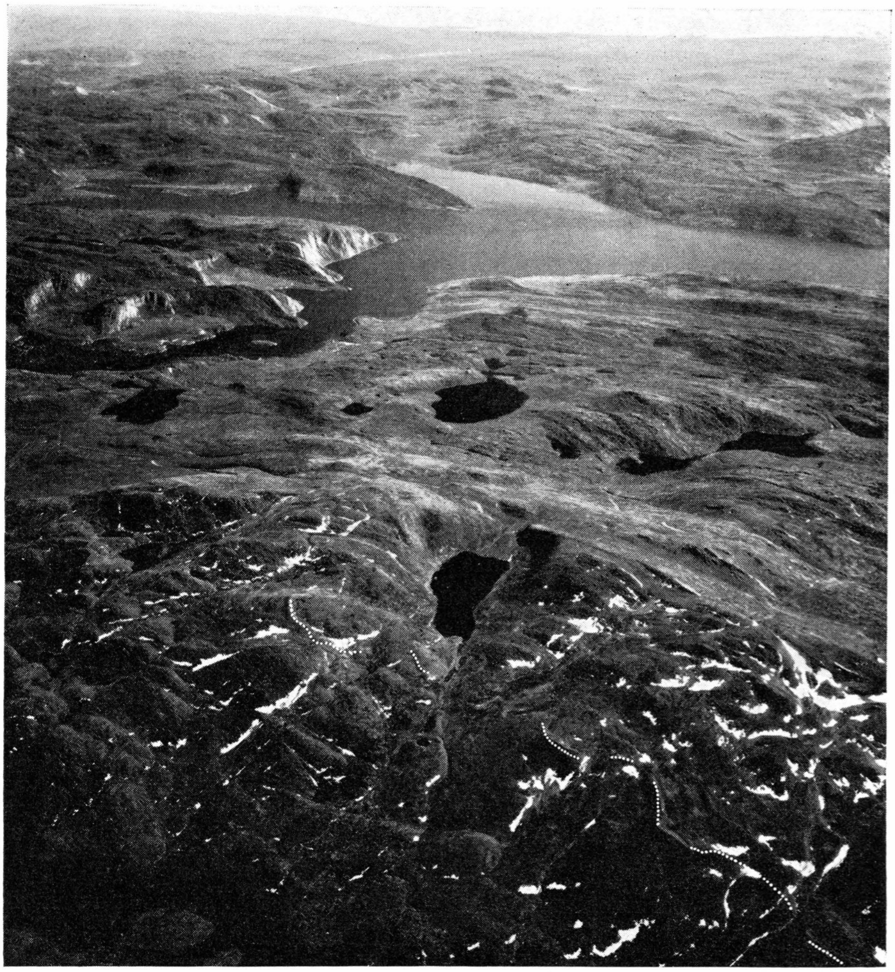

Fig. 54. Ikertoq fjord, seen from the north, with the highest deposits of the nunatak zone in the centre foreground. Geodetic Institute's route 508 C–S, no. 1618 (16.7.1948). Copyright Geodetic Institute.

and the relatively few observations hitherto made, do not allow a finer distinction of the stages in the deposition of the deposits and they have been grouped tentatively under the label "nunatak zone". The age of the deposits of the zone is unknown, but at least the youngest and lowest stage must, from the observations referred to above, be nearly contemporaneous with a sea level of 130 m a.s.l.

Although no nunatak zone moraines are known for the area further south and west, the schliffgrenzen (in the sense of Klebelsberg (1948, pp. 340–341)) which separate the jagged peaks from the smooth and rounded lowland may be correlated with them. These features are at

c. 800 m a.s.l. in the central parts of the highland of the outer coast and slope down westwards towards the coast, where they are between 200 and 400 m a.s.l.

North of the area of the nunatak zone deposits proper is a much discussed moraine feature at the entrance of Blæsedalen, Disko island. This moraine is taken to be either a terminal moraine from a glacier lobe in Blæsedalen or a marginal moraine from a lobe of the Inland Ice occupying the sound between Godhavn and Egedesminde. Its rectilinear form across the valley points towards it being deposited by the Inland Ice margin as a medial moraine between a valley glacier in Blæsedalen and the main lobe in Disko Bugt. An outwash feature at c. 100 m a.s.l. at this moraine may be of marine origin, but not until c. 55 m a.s.l. are clearly developed beach ridges present. Like the schliffgrenzen mentioned above this moraine may be related to the nunatak zone.

10. DEPOSITS FROM LOCAL GLACIATIONS

10.1. Types of deposits

A sharp distinction between local glaciations and the Inland Ice is not always possible as in some areas firn fields or lobes of the local ice are confluent with lobes from the Inland Ice. Such conditions of course will result in complicated ice margin features, as for example in the Sukkertoppen district and Nûgssuaq peninsula.

For areas nearer the western coast, this complication does not exist and the marginal moraines of cirque and valley glaciers can be recognised in the field. However, because of their limited extent, they are not often visible on aerial photographs. This difficulty is accentuated by their being frequently formed of coarse detritus whose vegetation cover differs little from that of the surrounding bedrock, unlike the fine grained redeposited fluviatile or marine sediments common in moraines from the Inland Ice margin.

Where the deposits from local glaciations have been mapped, attention has mainly been focused on determining the height of the glaciation limit which existed during their deposition.

10.2. Julianehåb district to Godthåb district

In this region there are many relatively elevated areas near to the sea and many existing local glaciers. The local glaciations have been described previously (Weidick 1963b) and it will suffice to mention here that the deposits in the southern part of the Julianehåb district have been divided into the following stages:

1) The Niaqornakasik stage, formed at a glaciation limit 700–400 m lower than the present one and contemporaneous with a sea level between 38 and 24 m a.s.l.
2) The Tunugdliarfik stage formed with a glaciation limit 300–200 m lower than the present one and contemporaneous with a sea level between 15 and 10 m a.s.l.
3) The Narssarssuaq stage formed with the glaciation limit 200–100 m lower than the present one and with sea level at its present level.

Fig. 55. Tasiussaq area, Sukkertoppen district, seen from the west. (a–a); Inugssuit taserssuat, (b–b); glacier loc. 62 in plate 2. Geodetic Institute's route B 19 A–L, no. 30. Copyright Geodetic Institute.

Little is known from the Frederikshåb district. In the area south of Sermiligârssuk, WEIDMANN (1964) states that between 500 and 1000 m there are numerous traces of marginal deposits of small glaciers, which developed after the disappearance of the Inland Ice.

Some details have been given for the southern part of Godthåb district, around and immediately north of Frederikshåbs Isblink by GRAFF-PETERSEN (1952, p. 267) who considered that since the deglaciation of the area by the Inland Ice, local glaciations have played a very minor role.

On aerial photographs, moraines of local glaciations have been observed on Nukagpiarssuaq, in the interior part of Bjørnesund at c. 500 m a.s.l., at Qáqat akulerît 12 km northeast of the head of Bjørnesund between 500 and 600 m a.s.l., and 5 km southeast of Isortuarssûp tasia at 1000 m a.s.l. (see plate 3).

It is considered that all these deposits were formed at a glaciation limit not more than 200–300 m below the present one. A slightly older stage may possibly be represented by moraines in the northern part of the Godthåb district at the localities of Qingaq and Augpalârtoq (DE QUERVAIN and MERCANTON 1925, p. 173).

10.3. Sukkertoppen and Holsteinsborg districts

The alpine topography of the region bordering the Davis Strait results in the existence there today of numerous valley and cirque

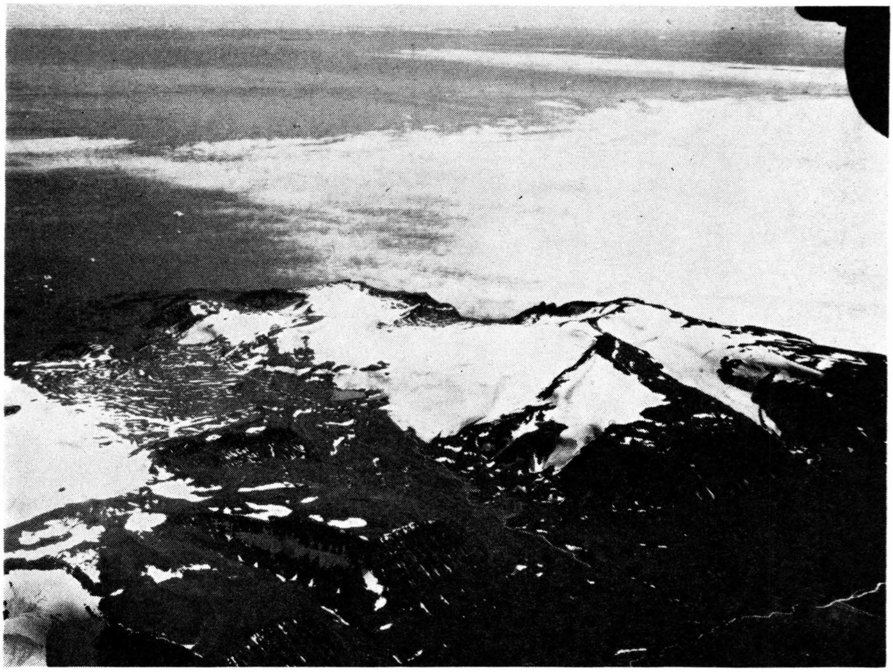

Fig. 56. A recently deglaciated valley, Tunorssuaq, Disko island. (a–a); loc. 85 A, (b–b); loc. 85 B in plate 2. Geodetic Institute's route 515 F–Ø, no. 12163 (15.8.1953). Copyright Geodetic Institute.

glaciers. Moraines from local glaciers, requiring a depression of the glaciation limit of 200–400 m, are seen on the south side of Nukagpiaq at 500–840 m a.s.l. and on the north side of the same mountain group at 600 m a.s.l.

Marginal glacial deposits also exist at Tasiussaq at Sukkertoppen town. The oldest of these are to be found at the western end of a large lake, Inugsuit taserssuat (fig. 55) where a terminal moraine has been cut by a terrace at 65 m a.s.l. South of Inugsuit taserssuat, around two existing glacier lobes, are several moraines formed when the glaciation limit was not more than 200 metres lower than now.

Moraines have also been found around glacier lobes in the interior part of the fjord Íkamiut kangerdluarssuat. The oldest and greatest of these terminal moraines is cut by a terrace c. 30 m a.s.l. and must have formed whilst the glaciation limit was little more than 400 m lower.

At Íkátûssaq, just south of the entrance to Søndre Strømfjord, local glaciation moraines have been attributed, on the basis of lichenometry, to the "hypothermal" (BESCHEL 1961, p. 1060). The distal parts of these moraines are cut by beach ridges at 80–90 m a.s.l. Immediately outside the moraines are well developed glacial striae indicating ice

Fig. 57. Sarqaq glacier, south coast of Nûgssuaq peninsula, seen from the southeast. Geodetic Institute's route 520 G–NV, no. 2189 (15.7.1948). Copyright Geodetic Institute.

movement from the east, a direction nearly at a right angles to that of the local glaciation which produced the moraines.

A somewhat greater extent of local glaciers is demonstrated by terminal moraines at the lake of Taseq qutdleq, 12 km northeast of Ikâtûssaq. On aerial photographs the terminal moraines can be seen to reach down to c. 50 m a.s.l. corresponding to a depression of the present glaciation limit of c. 400 m. In the same valley, 8–10 km further east are other terminal moraines representing stages when the glacier was smaller and the depression of the glaciation limit less.

10.4. Disko island

A large part of this island is at present covered by glaciers and ice caps. High basalt plateaus dominate the landscape and a depression of the present glaciation limit of only a few hundred metres would have a great effect on the glaciation of the island. At the localities of Tunorssuaq (fig. 56) and Blæsedalen on the southern part of the island are prehistoric moraines situated near to, but outside, the moraines from historical time.

10.5. Nûgssuaq

Remnants of prehistoric local glaciations have been found at several localities. At the glaciers near Sarqaq on the south coast of the peninsula (fig. 57), terminal moraines at 300 m a.s.l. lie just outside those from historical time. Further east, in Saputit valley, are systems of boulder moraines, the outermost of which indicates a depression of the present glaciation limit of 200–500 m. Still further east, moraines from local glaciations in Qôrorssuaq, mentioned earlier (p. 123 and fig. 30b) imply a depression of the glaciation limit of the same magnitude.

In the interior part of the peninsula, at the western end of Sarqap taserssua lake, the present glacier lobes are surrounded by older moraines in addition to those of historical time. However they show that in this area also, the glaciers were not much larger than in historical time. Again a relatively small depression of the glaciation limit there would result in the total glaciation of the valleys.

The possible existence of a "Daun stage" on the northern coast of Nûgssuaq peninsula, characterised by moraines immediately outside those of historical time, was stated by HEIM (1911, p. 227), when describing the area around Qaersuarssuk.

10.6. Conclusions for local glaciations

The most frequently found deposits relate to a period when the extent of the glaciers was little greater than that of historical time. In the Sukkertoppen and Holsteinsborg districts they have been referred by BESCHEL (1961) on lichenometrical evidence to advances around 500 and 2000 years B.C. No exact date can be given for the deposits which imply a greater extension of local glaciers to near sea level, as at Niaqornakasik in the Julianehåb district (see fig. 32), Inugsuit taserssuat and Îkamiut kangerdluarssuat in the Sukkertoppen district, since the terraces to which they are related have not been proved to be of marine origin. However, it seems that the local glaciations were never of very great extent and that the glaciation limit, whilst the glaciations were independent entities, was not a great deal lower than the present, with the exception of those in the Julianehåb district. Furthermore, the persistence of an extensive continental glaciation has delayed development of local glaciation until a relatively late stage in the deglaciation of the coastal stretch.

11. CONCLUSIONS

In this paper a study has been made of the connection between climatic fluctuations and the effects of these fluctuations on glaciers as shown by the glacial deposits. However, as MEIER stated (1965, pp. 795–802) each link in this chain involves assumptions that have not yet been clarified.

11.1. Deposits of historical time

Historical records indicate that readvance maxima occurred around 1850?, 1890 and 1920 at glaciers in West Greenland. These are in phase with glacier fluctuations in other parts of the world, which implies that the older readvance maxima of other places at around 1650 and 1750 also occurred in West Greenland. This assumption appears to be confirmed by the lichenometrical dating of moraines in the area to this period by BESCHEL (1961).

The periods of glacial advance which culminated in the maxima of 1850?, 1890, and 1920 seem to have been initiated essentially by periods of low temperature, with changes in precipitation being of secondary importance only. This must also be true for the older periods of 1650 and 1750. For all these readvances there was a time lag of a few years to two decades between change in the climate and the resulting glacial maxima. However, it must be realised that the climatic data, and especially the data on the mass balance of the glaciers, are too few to allow clear conclusions to be drawn about the short term periodic behaviour of the glaciers.

Each period of maximum extent of the glaciers left its mark as narrow lines of ice margin deposits. For a full development of these deposits the older advances must be the greatest and the youngest the least—a prerequisite that is largely satisfied in the southern and western parts of West Greenland. Further north the older moraines are to a greater or lesser extent covered by younger deposits.

In general terms, the deposits formed between 1650 and 1920 describe the behaviour of glaciers over a period—a "stage" (or stade) during which they responded to a major climatic fluctuation of several centuries' duration—the cold period from the 16th century to the end of the end of the 19th century. Secondary, short term fluctuations of ice margins in response to climatic fluctuations of a few decades' duration constitute "phases" within a stage.

The associated fluctuation of the height of the glaciation limit during this stage was over a height range of 100–200 m. However, its fluctuations during individual phases are not clear because of the general uncertainty about its exact position.

11.2. Deposits of prehistoric time

For the recognition of prehistoric climatic events the chain of processes must be followed in the opposite order, from the glacial deposits to the changes in climate causing them. As a first step the descriptive term "zone" is used and is defined as a continous wide belt of ice margin deposits. The dimensions of a zone indicate that the fluctuations in the extent of the Inland Ice margin involved are of the order of a stage. For prehistoric local glaciations, where the deposits do not have their zonal characteristics, the calculated approximate position of the glaciation limit is used as a criterion for distinguishing stages.

There can not always be a simple relationship between morphological zones and stages, *e.g.* different cold periods of similar magnitude would result in a marginal deposition in the same zone, or subsequent cold periods of greater magnitude would result in obscuring the deposits of earlier lesser cold periods. In these cases the stratigraphy of the deposits must be used to distinguish the stages.

11.3. Inner zone

The surface characteristics (p. 83 and fig. 31) of the inner zone is 350 m. However, it is questionable to what degree they should be classified as belonging to a single stage as in many cases their only common feature is their contemporaneity with a sea level more or less at its present position. The moraines of the Drygalski stage in Disko Bugt–Umanak district with their uniform continuity and wide extent are mainly referable to a single stage. So too are the innermost deposits of the Holsteinsborg and Sukkertoppen districts and the deposits of the Narssarssuaq stage in the Julianehåb district. With the possible exception of moraines in the Julianehåb district these inner zone moraines are probably younger than c. 4000 years (fig. 2, p. 15). Furthermore, at Qajâ (p. 90) the inner zone predates deposits of c. 3500 B.P., which provides a minimum age for at least this section of the zone.

11.4. Outer zone

The surface characteristic of the outer zone is 650 m (fig. 31). Its deposits have a narrow vertical and lateral distribution but a number of phases are recognisable and these can be grouped into two stages, the youngest of which is usually the more clearly developed. This younger stage is everywhere found to be contemporaneous with a sea level 35–60 m a.s.l. A corollary of this is that the uplift of the coast has been rather uniform along a great length of the west coast of Greenland. Reference to the inland part of the trend in fig. 2 suggests that the age of this stage can be considered as 7500–8500 years. Only in the Julianehåb district, where the rate of uplift is not known, is there some uncertainty as to the relation of the local Tunugdliarfik stage to the youngest stage of the outer zone. Because of this the name "fjord stage 1" is applied only to the deposits of the outer zone between the Godthåb and Umanak districts while the local name of Tunugdliarfik stage is used for the local deposits in the Julianehåb district.

At several places between the Godthåb district and Disko Bugt the older stage of the outer zone can be seen to have been formed whilst the sea level was a little over 70–80 m a.s.l. From the uplift trend of fig. 2 the age should thus be between 9000 and 9500 years old. This stage is called "fjord stage 2".

11.5. Nunatak zone

The deposits of the nunatak zone are restricted to the Sukkertoppen and Holsteinsborg districts. Its surface characteristic is believed to be between 1000 and 2000 m (p. 80). Several stages in the deposition of the nunatak zone are distinguishable. The youngest, Avatdleq stage occurred at the formation of a marine level of 130 m a.s.l. and before one at 95–100 m the age of which is 8250 ± 130 years B.P. (K-1037). However, its situation outside the fjord stage 2 deposits indicates that the age of this stage must be older than 9500 years. With a distance between the outer zone and the nunatak zone of c. 50 km and a maximum recession of the ice margin of c. 100 m annually, the retreat of the Inland Ice margin between the two zones must have taken under optimal conditions at least 500 years. Hence, the Avatdleq stage must be considered to be nearly 10,000 years old and the more westerly situated stages of the zone even older.

The glaciation limit during these stages is believed to have been situated little more than 300–500 m below the present one.

11.6. Local glaciations

Deposits from local glaciations must be relatively young because of their proximity to the present glaciers and deposits from historical time. The glaciation limit at the time of their deposition is considered to have been 200–300 m lower than the present one, and the sea level stood near the present level. Thus, these deposits are correlated with those of the inner zone of the Inland Ice.

A few older deposits of local glaciations, corresponding to a greater extent of glaciation than those above, can be related to terraces, which may be of marine origin, *e.g.* Tasiussaq and Íkatûssaq in the Sukkertoppen district. If so these deposits correlate with fjord stage 1 or 2. The depression of the associated glaciation limit is thought to have been 300–500 m.

It is not certain to which stage the deposits of local glaciations at Niaqornakasik in the Julianehåb district should be referred. The related terrace systems and the necessarily relatively low glaciation limit imply their correlation with the fjord stages or possibly even with deposits of the Nunatak zone.

A striking feature is the relative unimportance of the local glaciers to the glaciation of the area during the prehistoric period. This may be due to the glaciation limit already having risen to heights almost equal to those of today before deglaciation of the coastal stretch reached its present extent.

The glacial events in West Greenland may be expressed schematically:

Deposit	Age	Classification	Marine level	Depression of glaciation limit
Historical time	350–30 B.P. (c. 1600–1920 A.D.)	1 stage, several phases	0 m	100–200 m
Inner zone	4800–2500 B.P.?	Several stages?	0 m	200–300 m
Outer zone	7500–9500 B.P.	2 stages:		300–500 m?
		Fjord stage 1 (7500–8500 B.P.)	35–60 m	
		Fjord stage 2 (9000–9500 B.P.)	70–80 m	
Nunatak zone	$\geqslant$ 10,000 years B.P.	Several stages. Youngest:		300–700 m?
		Avatdleq stage (c. 10,000 B.P.)	130 m	

The position of the zones is shown schematically in fig. 58.

11.7. Correlation with other parts of Greenland

The only published survey of glacial deposits of regional extent in Greenland is that from northern and northeastern Greenland by DAVIES (1961) and KRINSLEY (1961). They recognise widespread deposits of the Inland Ice belonging to a glacial stage which they date to between 3700 and 6000 years B.P. However, these moraines are related to a marine level of c. 60 m a.s.l. and since the available radiocarbon dates suggest that the uplift there has been essentially contemporaneous with the uplift of West and East Greenland, the moraines are probably older than the date suggested. They should therefore probably be correlated with the deposits of the outer zone of West Greenland rather than with the deposits of the inner zone.

Whilst the behaviour of the Inland Ice in historical time suggests that some difference in the time of response to climatic fluctuations should exist between North and West Greenland, not enough is known for the implication of this in terms of the older glaciations to be predicted.

Though the inner zone has not defenitely been recognised in North Greenland, some "inner zone" moraines, mapped by KRINSLEY (1961) around the interior parts of Danmarks Fjord, may possibly belong to it. However, no dates are available for their formation.

In central East Greenland, LASCA (in print) concluded from his investigations in the Skeldal area, near Mesters Vig, that the area was deglaciated before the Allerød and that two later readvances of the glaciers occurred. These he dates, on indirect evidence, to the Younger *Dryas* time and to some time prior to 1500 B.P., possibly around 2600 B.P. Further south, in Scoresby Sund, JOHN and SUGDEN (1965) have described an older stage near Schuchert Dal area, which coincides with a marine level of c. 100 m a.s.l. They refer it on indirect evidence to the time of the formation of the Fennoscandian moraines of Scandinavia, *i.e.* c. 10,000 B.P.

11.8. Correlation with areas outside Greenland

Outside Greenland, glacier advances dated to the Sub-Atlantic have been described from Norway (ØSTREM 1961, p. 419), Iceland (THORARINSSON 1949, p. 250), the Alps (AARIO 1944, p. 28, HEUBERGER 1956, pp. 91–98, BESCHEL and HEUBERGER 1958, pp. 91–93), western United States (HEUSSER 1957, p. 144) and Alaska (GOLDTHWAIT *et al.* 1963, p. 72). Other advances have been dated to the Sub-Boreal in the Alps (GFELLER *et al.* 1961, p. 19), United States (MEIER 1963, p. 72) and Alaska (KARLSTROM 1964, p. 63). In a recent compilation by MERCER (1965) the two chief periods of advance are considered as having been between

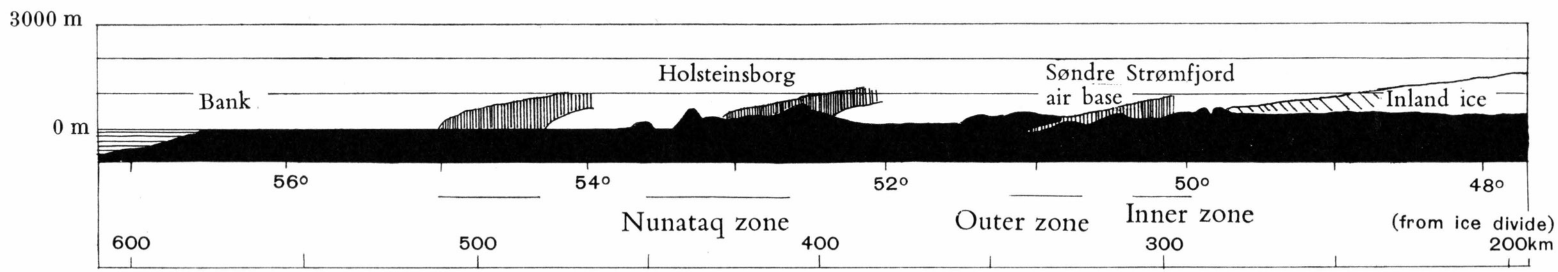

Fig. 58. Profile across West Greenland near 67° N showing, cross-hatched, the zones of marginal deposits from the Inland Ice. The substratum under the Inland Ice is after BAUER and HOLTZSCHERER (1954, pp. 22 and 25).

2000–2500 years B.P. and 4000–4800 B.P. respectively. Presumably the advances represented by the inner zone in West Greenland are to be correlated with these.

The outer zone does not seem to have any equivalent in northern Europe. It may be that the waning ice sheet had lowered its surface to such an extent that it was hardly affected by a slight lowering of the glaciation limit. However, in North America, the Cochrane advance of Hudson Bay occurred in this period and the ice cap between 8000 and 9000 years B.P. covered great areas around Hudson Bay (Karlstrom 1956, Falconer *et al.* 1965a, b). The altitude of this ice cover must therefore have been sufficiently high for it to have been responsive to fluctuations in the glaciation limits in the area. An advance from this period is also known from Alaska, the "Tanya advance" of Karlstrom (1965, p. 119).

Since nothing is known about the difference in age between the fjord stage 2 and the deposits of the nunatak zone (included the Avatdleq stage) it is difficult to correlate them with other areas.

It is possible that in Greenland readvances of the nunatak zone were minor events during the rapid retreat of the ice, and thus less significant than some of the younger ones. In this sense certain points of resemblance can be found in the situation before c. 10,000 B.C. in Norway (Andersen 1960, 1965) and with the southern parts of the North American ice sheet (Wayne and Zumberge 1965). The deposits in all three areas seem to represent a series of advances or halt phases closely following each other. At least parts of the nunatak zone deposits may represent the stages of Ra-Salpausselkä in Scandinavia and in Iceland, where moraines of this stage were described by Einarsson (1963, 1964). In this context, it is believable that the moraines of the last ice age in West Greenland must be found in the banks offshore, as already presumed by Helland (1876, p. 68). This last assumption is illustrated in fig. 58 by a moraine zone between Holsteinsborg and Store Hellefiskebanke.

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The illustrations were drawn in the drawing office of the Geological Survey of Greenland with the exception of the figs. 7 and 22, which were drawn by the author.

13. DANSK RESUME

1. (pp. 7–9). Indledningsvis gives en oversigt over såvel tidligere som igangværende glaciologiske og glacialgeologiske undersøgelser af området. Det undersøgte område omfatter Vestgrønland fra Julianehåb distrikt til Upernavik distrikt. Væsentlige dele af behandlingen er dog koncentreret til de centrale dele af dette område d. e. mellem Godthåb distrikt og Disko Bugt. Afhandlingens formål er her at sammenligne forløb og udbredelse af gletscherfluctuationerne i historisk tid med ældre tiders udbredelse af gletscherne.

2. (p. 10–18). En gennemgang af tidligere undersøgelse angiver, at det må antages at Vestgrønland har været omtrent totalt nediset under sidste istid. Det må formodes at enkelte højtliggende partier af de kystnære bjergmassiver omkring Sukkertoppen og Holsteinsborg på dette tidspunkt har raget frem af isen som nunatakker. En landhævning efter sidste istid har haft samme forløb i Vestgrønland som angivet for Østgrønland af Stuiver & Washburn (1962) og Lasca (in print). Det fremgår af landhævningens forløb (se fig. 2), at væsentlige dele af Indlandsisens rand må have været bortsmeltet fra det nuværende kystområde tidligt i Holocæn tid.

Undersøgelser af den klimatiske udvikling i området er tidligere foretaget på pollenanalytisk og marint faunistisk grundlag og viser overensstemmelser med det samtidige klimaforløb i Skandinavien. Mere detaillerede kvantitative undersøgelser over klimaudviklingen gennem det sidste hundrede år viser en generel temperaturstigning på ca. 2° C for året og ca. 1° C for sommerperioden. Temperaturstigningen har dog været afbrudt af kuldeperioder 1880/90 og 1913/16. Yderligere er af Vibe (1967) ved ældre observationer påvist kuldeperioder 1807/21 og 1860/66. Medens der for hele området spores samme forløb i temperaturændringer, kan noget sådant ikke påvises for nedbørsfluctuationer.

3. (pp. 19–27). Som bedst egnet udtryk for den recente nedisnings omfang angives glaciationsgrænserne og et kort over disses højder i Vestgrønland er angivet i fig. 6. Det fremgår af det sparsomme materiale, at de på forskellig vis definerede sne- og glaciationsgrænser alle udgør en zone, som med en bredde af ca. 200 m forløber fra 600–800 m o. h.

ved yderkysten til 1400–1700 m o.h. over Indlandsisens rand. Denne zone ses også at falde sammen med ligevægtsgrænsen for gletscherens omsætning, hvilket er udtrykt i gletschernes »aktivitetsindex«.

For Indlandsisens rand og mod nord i området findes et ringe aktivitetsindex, mod syd og ved lav beliggenhed af glaciationsgrænsen et stort. Det er derfor muligt at gletscherne i Sydgrønland vil reagere hurtigere på klimaændringer end hvad tilfældet er i Nordgrønland.

Højden af glaciationsgrænsen synes derfor at have betydning for gletschernes massebalance. Der findes for Vestkysten en grov korrelation mellem glaciationsgrænsens højde og sommertemperaturens og den årlige nedbørs størrelse. Temperaturens indflydelse på glaciationsgrænsens beliggenhed er langt større end nedbørens. Overensstemmende med denne forbindelse findes, at glaciationsgrænsen, vurderet fra tidligere observationer, i løbet af ca. 100 år har hævet sig 100–200 m.

4. (pp. 28–65). Fluctuationer af ca. 500 gletscherlober er gennemgået på grundlag af historisk materiale. For største delen af disse gletscherlober dækker de givne data kun de sidste ca. 30 år og loberne udviser her blot recession. Det behandlede materiale omfatter derfor kun 135 gletschere, hvis data dækker et længere tidsrum og hvor lokaliteterne har en passende geografisk spredning.

Data for de 135 gletscherlober med henvisning til observationens art eller kilder er givet i »List of glacier fluctuations«, pp. 162–187. De samme fluctuationer er grafisk opstillet i kurverne tavle 2.

Det fremgår af materialet, at fluctuationerne af gletscherfronterne har samme fase, men forskellige amplitude for gletscherlober med hhv. lokale firnområder (»lokalglaciationer«) og Indlandsisens firnområde. Yderligere, at temperaturændringerne er af primær betydning, og at forsinkelse i tid mellem temperaturminima og tilsvarende gletscherfremstød er på nogle få år eller årtier. Temperaturændringen i de første årtier af det 19. århundrede resulterede således i fremstød, afsættende moræner omkring 1840–1850. Perioden i 1860'erne initierede fremstødene mellem 1880 og 1900 (1890'ernes fremstød) og den korte fase 1913–1916 fremstød kulminerende i 1920'erne.

Hvad angår gletscherfluctuationernes amplitude, nåede fremstødene 1840–1850 og 1880–1900 oftest omtrent til maximal udbredelse for historisk tid, medens fremstødene i dette århundrede har været af mindre omfang. Et ældre fremstød, dateret af BESCHEL, angiver yderligere en maximal udbredelse af gletscherne omkring 1650 og 1750.

Det må fremhæves, at der i den sydligste og mest oceanisk influerede del af Vestgrønland spores en tendens til stor udbredelse af gletscherloberne allerede før 1880, d.v.s. måske o. 1750, medens de kontinentalt prægede gletscherlober fra Indlandsisen viser en maximal udbredelse omkring 1900. Yderligere ses omfanget af fremstødet omkring 1920 at stige i de nordligste dele af området (Umanak distrikt). Denne tendens bliver yderligere accentueret i Nordgrønland, hvor DAVIES har påvist en maximal udbredelse af flere gletscherlober omkring 1920–30.

Sammenholdt med gletscherfluctuationer fra andre områder i den atlantiske del af den nordlige hemisfære, synes de ovennævnte anomalier i gletscherfluctuationernes amplitude at passe ind i det billede af gletscherfluctuationer, SHUMSKII har givet. SHUMSKII anfører, at de højpolare gletschere i det nordligste Grønland, Axel Heiberg Land og i det nordlige dele af Baffin Island viser ringe reaktion på klimafluctuationer, medens en zone over Island-Sydvestgrønland og muligvis dele af Østgrønland, Jan Mayen, Norge og Spitzbergen viser en hurtig reaktion. I forbindelse med denne zonale fordeling af gletschernes reaktioner viser loberne ved højpolare gletschertyper en tendens til opretholdelse af eller udbredelse til maximum i historisk tid indtil de seneste årtier, medens de tempererede gletschere ofte angives at have maximum o. 1750 eller 1850. Den undersøgte region i Grønland ses således at danne overgangsled mellem de to zoner.

En sammenhæng mellem data for gletscherfluctuationerne og den kendte sekundære sænkning af Grønland, begyndt i det 17. århundrede og afløst ca. 1940 af hævning, kan ikke med sikkerhed findes på foreliggende grundlag.

Af de givne data ses, at sammensynkningen af Indlandsisens rand er noget mindre end for loberne i lokale firnområder, beliggende ved yderkysterne, hvor smeltningen har sat ind tidligere. Under antagelse af, at de undersøgte lober fra Indlandsisen repræsenterer et gyldigt gennemsnit af sammensynkningen af Indlandsisens rand i den sydlige halvdel af Grønland, og at en ringere afsmeltning af randområdet i Nordgrønland kompenseres af det ekstra tab ved israndens tilbagetrækning, findes et gennemsnitligt tab af Indlandsisens randområde på ca. 200 km^3 årligt. Resultatet svarer til en beregning af Indlandsisens tab, foretaget af BAUER på grundlag af estimerede værdier af Indlandsisens massebalance.

5. (pp. 66–78). De recente aflejringer ved Indlandsisens rand behandles kun kort. Det anses dog for berettiget fra de observerede israndsaflejringer i historisk tid, i grove træk at kunne rekonstruere Indlandsisens overfladeforhold også under tidligere stadier ud fra aflejringer, afsat af isranden.

To typer aflejringer ses i Vestgrønland at være af interesse ved den recente isrand: dødisterræn og blokmoræner.

Det fremgår af observationerne, at dødislandskaber i historisk tid initialt dannes ved shear moraines, men at disse shear moraines kan dannes ved enten transversale eller longitudinale forskydninger i gletscherisen. Dødisdannelse må forudsætte delvist eller totalt stop af den egentlige gletscherflydning. Dette betinges af en udtynding af gletscheren til et minimum af mægtighed, hvis størrelse afhænger af gletscheroverfladens hældning.

Blokgletschere ses at optræde i stort tal i den nordlige del af det undersøgte område, især på Nûgssuaq halvøen og i Umanak distrikt. Foruden den almindeligt antagne forudsætning for deres eksistens: udpræget jointing af oprindelsesbjergarten og et alpint terræn, antages her en oprindelse, hvis væsentligste elementer er sidemoræner- og talusdannelse. Under udtynding af den oprindelige gletscher afsnøres gennem marginal shear dødisområder, dækket med blokmoræne. Derefter omformes disse dødismoræner ved dalværts krybning til blokgletschere med den karakteristiske ogive dannelse på overfladen. Ved gradvis forsvinden af isen i blokmellemrummene ophører krybningen og det nuværende bloktilskud består blot i talusdannelse på toppen af blokmorænerne. Det antages derfor at udstrakte blokmoræneforekomster såvel med som uden blokgletscherdannelse ækvivalerer sidemoræner ved bestemmelse af ældre israndsstadier.

De øvrige glaciale aflejringer frembyder ligheder med, hvad der er beskrevet fra Sydvestnorge. Mere specielt for grønlandske forhold er den hyppige forekomst af interlobat moræneterræn og nunatakmoræner. Årsmoræner dannet i historisk tid er kun observeret et enkelt sted, og morænematerialet tillader ingen sondring mellem push- og depositional moraines.

Som helhed konkluderes, at samtlige israndsaflejringer afsat i historisk tid danner et bælte af moræner, der karakteriserer et enkelt stadium i lobernes udbredelse, medens de lokale faser (1650?, 1750?, 1850?, 1890 og 1920) næppe kan korreleres morfologisk over større strækninger uden en nøje datering. Dette gælder såvel Indlandsisens aflejringer som aflejringer fra lokale firnområder.

6. (pp. 79–82). Undersøgelse af aflejringer dannet i historisk tid må angive, at bæltet af israndsaflejringer af samme bredde og mægtighed som dem der er dannet i historisk tid, må angive "klimaforværringer" af samme størrelse (ækvivalente temperaturændringer) og/eller samme intensitet (d. e. dækkende flere hundrede år).

Præhistoriske israndsaflejringer forekommer især i stor mængde afsat af Indlandsisen. På grund af den hyppige forekomst af nunatakmoræner er foretrukket en preliminær inddeling og korrelation ved hjælp af de tidligere israndsaflejringers beliggenhed over den nuværende overflade af Indlandsisen (israndsaflejringernes »karakteristik«). Kontrol af denne korrelation kan derefter i enkelte tilfælde foretages ved undersøgelse af havets niveau under israndsaflejringernes afsætning. Af aflejringer fra

Indlandsisen skelnes mellem en »Indre zone«, en »Ydre zone« og en »»Nunatak zone«.

Præhistoriske aflejringer afsat af lokalglaciationer er kun kortlagt i begrænset omfang. Til korrelation anvendes her glaciationsgrænsebestemmelse ved morænens afsætning efter Penck og Brükner's fremgangsmåde i Alperne. Kun i enkelte tilfælde er en korrelation ved bestemmelse af det marine niveau under aflejringernes afsætning mulig.

7. (pp. 83–92). Aflejringerne fra den »Indre zone« gennemgås distriktsvis. Aflejringerne fra denne zone forekommer spredt og karakteriseres alle ved at være dannet ved et marine niveau omtrent som det nuværende. Forekomst af eskimoruiner på og indenfor disse aflejringers område taler for, at i det mindste en del af zonen må være afsat for mere end 3500 år siden. Muligheden for, at disse aflejringer er dannet under tidsmæssigt forskellige stadier i Indlandsisens udbredelse kan ikke udelukkes.

8. (pp. 93–124). En lokalitetsvis gennemgang af den »Ydre zone's« aflejringer foretages. Zonens aflejringer er tidligere af forfatteren beskrevet under navn af »fjord stadiet«. Zonens aflejringer er afsat ved havniveauer mellem 35 og 80 m over det nuværende. I Julianehåb distrikt er det dog muligt, at det marine niveau under disse aflejringers afsætning kun har været 15 m over det nuværende havniveau. Der kan undertiden skelnes mellem et nedre moræne system, dannet ved et marint niveau af 35–60 m og et øvre system, dannet 70–80 m o. h. Det må derfor antages fra forløbet af landets hævning (fig. 2) at det nedre system (fjord stage 1) er afsat for mellem 7500 og 8500 år siden, det øvre (fjord stage 2) for mellem 9000 og 9500 år siden.

9. (pp. 125–129). »Nunatak zonens« aflejringer er kun fundet i de centrale dele af Vestgrønland: Sukkertoppen og Holsteinsborg distrikter, hvor de i kystlandets fjelde er fundet mellem 600 og 1000 m o. h. Herfra kan israndsaflejringerne i Ikertôq fjordens indre område for det inderste og yngste stadiums vedkommende følges ned til fjorden. Det ses dannet ved et marint niveau 130 m o. h., hvilket peger mod en alder af 10000 år eller mere. Denne alder angiver således et minimum for det indre og yngste system af zonens aflejringer.

10. (pp. 130–134). Israndsaflejringerne fra de lokale gletschere angiver næsten alle en dannelse ved en glaciationsgrænse, næppe mange hundrede meter under den nuværende. Undtaget er dog Niaqornakasik i Julianehåb distrikt, dannet ved en glaciationsgrænse 400–700 m under den nuværende. De fleste aflejringer med kontakt til marin behandling

angiver desuden en dannelse ved marine niveauer nær det recente. Her må dog ligeledes undtages Niaqornakasikstadiet i Sydgrønland, der er skåret af terrassedannelser mellem 38 og 24 m o.h.

11. (pp. 135–141). Den oftest sene udvikling af lokalglaciationernes aflejringer antyder, at glaciationsgrænsen allerede ved kystlandets deglaciation var hævet omtrent til den nuværende højde. Således fandtes der ikke betingelser for, at extensive lokalglaciationer dækkede områder, fremsmeltet fra Indlandsisen.

Det er sandsynligt, at Julianehåb distrikt er blevet deglacieret betydeligt tidligere end det øvrige område fra den kontinentale nedisning, hvorfor mulighed for en mere udbredt følgende sekundær, partiel nedisning under lokalglaciationer er større her end i de mere nordligt beliggende områder.

Sammenligninger med områder udenfor Grønland gør det sandsynligt, at den »Indre zone's« aflejringer må henføres til flere stadier; formentligt dem, dannet indenfor de to perioder 2000–2500 og 4000–4800 før vor tid.

De to stadier i den »Ydre zone's« aflejringer: Fjord stage 1 og 2 synes ikke at have nogen europæiske ækvivalenter. Derimod synes de tidsmæssigt at kunne sammenstilles med Cochrane fremstødet i Nordamerika, der vides at have fundet sted mellem 8000 og 9000 år før vor tid.

»Nunatak zonen« aflejringer må derimod formodes at kunne korreleres med yngre Dryas tid og muligvis også ældre Dryas tids aflejringer i Nordeuropa og Nordamerika. En nøjere datering og differentiation af de enkelte faser i denne zones aflejringer er ikke gennemført.

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Abbreviations: CRREL; Cold Regions Research and Engineering Laboratory, U.S. Army Material Command, formerly SIPRE.
IASH; International Association of Scientific Hydrology, Gentbrügge, Belgium.
SIPRE; Snow, Ice and Permafrost Research Establishment, Corps of Engineers, U.S. Army.

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15. TABLES

15.1. Relationship between precipitation, summer temperature and glaciation limit

Table 1 a. *Annual precipitation (p) and summer temperature (s), i.e. mean of June, July, August and September temperatures, at 15 West Greenland stations (after* BLINKENBERG *1952). Altitude of glaciation limit (g) at same stations taken from the map in fig. 6.*

Name of station	Latitude	Longitude	(s)	(p)	Period of observation	(g)
			(X_1)	(X_2)		(Y)
Prins Christians Sund	60°03′	43°17′	5.9° C	265 cm	1943–1948	750 m
Nanortalik	60°07′	45°11′	6.6° C	88 cm	1925–1941	750 m
Simiutaq	60°40′	46°35′	5.2° C	78 cm	1942–1948	1000 m
Narssaq	60°54′	46°00′	6.8° C	71 cm	1944–1948	1200 m
Narssarssuaq	61°09′	45°22′	8.7° C	70 cm	1941–1948	1500 m
Ivigtut	61°12′	48°12′	8.3° C	135 cm	1925–1941	1200 m
Grønnedal	61°14′	48°07′	7.0° C	113 cm	1943–1948	1200 m
Marraq	63°26′	51°12′	5.3° C	82 cm	1943–1948	700 m
Godthåb	64°10′	51°44′	6.4° C	51 cm	1925–1941	700 m
Cruncher Island	66°04′	53°30′	5.4° C	95 cm	1942–1948	600 m
Sdr. Strømfjord	67°00′	50°43′	7.5° C	13 cm	1941–1948	1350 m
Egedesminde	68°42′	52°50′	4.2° C	23 cm	1943–1945	750 m
Godhavn	69°15′	53°32′	6.4° C	41 cm	1925–1935	700 m
Jakobshavn	69°13′	51°07′	6.1° C	24 cm	1925–1941	1000 m
Upernavik	72°47′	56°10′	4.5° C	20 cm	1925–1941	725 m

Table 1b. *Determination of relation between s(= X_1), p(= X_2) and g(= Y). The determination was kindly made by* E. SVEINSDOTTIR. *Y_{EXP} is the observed value and Y_{CALC} the calculated value of Y.*

Linear regression analysis with two independent variables

$$Y = A + B_1 \times X_1 + B_2 \times X_2$$

X_1: Mean summer temperature June–Sept. in degrees centigrade
X_2: Mean annual precipitation in cm
Y: Glaciation limit in metres

X_1	X_2	Y_{EXP}	Y_{CALC}	$(Y_{EXP}-Y_{CALC})/S$
5.9	265	750	711	0.2
6.6	88	750	988	-1.3
5.2	78	1000	750	1.3
6.8	71	1200	1038	0,9
8.7	70	1500	1374	0.7
8.3	135	1200	1247	-0.2
7.0	113	1200	1037	0.9
5.3	82	700	764	-0.3
6.4	51	700	985	-1.5
5.4	95	600	770	-0.9
7.5	13	1350	1212	0.7
4.2	23	750	621	0.7
6.4	41	700	994	-1.5
6.1	24	1000	956	0.2
4.5	20	725	677	0.3

	A	B_1	B_2
	-99.381	176.371	-0.869
Standard deviations	48.997	40.103	0.814
Correlation coefficents	r_{YX_1} 0.764	r_{YX_2} -0.072	$r_{X_1X_2}$ 0.151
	$R_{YX_1X_2}$	= 0.787	

Standard error of estimate, S = 190

15.2. List of glacier fluctuations

Reference to the data listed is given in pp. 149–160. The first column shows the archive number of the locality as given in plate 1, lowermost point of the trim line zone (M_0) and the surface inclination of the glacier lobe (g). The significance of M_0 is shown in fig. 8 and the surface inclination of the glacier lobes (g) is given in ⁰/₀, i.e. 100 × the tangent of the slope. Where the slope varies greatly, the values of the inclination nearest to the front are italicised.

The second column gives the year, or approximate year, of observation and the third column shows the approximate values of a_t (see fig. 8) in metres. In the fourth column the values of a_t are converted to percentage values of maximum retreat (maximum retreat = 100 $^0/_0$) as explained on pp. 30–31 and p. 56. For a few of the localities (Loc. 69, 76, 106 and 107) no values have been given because of the lack of appropriate data.

In the comments on glacier fronts in the fifth column, "Calving" always means calving in the sea, otherwise "Calving in lake" is used.

The last column gives the source of the information. Besides the literary sources given in the bibliography it refers to the following collections in Copenhagen:

1) Arktisk Institut
JETTE BANG (loc. 39), J. C. D. BLOCH (loc. 11), A. JESSEN (loc. 2), L. KOCH (loc. 85, 86), H. MOLTKE (loc. 25), M. PORSILD (loc. 39), anon. (loc. 51, 52, 53, 54, 55, 130).
2) Det Kongelige Bibliothek (The Royal Library)
S. KLEINSCHMIDT (loc. 28), J. KREUTZMANN (loc. 33, 40, 41, 49, 50, 51, 52, 53, 55, 56, 57, 59, 61, 65), P. MOTZFELDT (loc. 12), L. MØLLER (loc. 25), B. PETERS (loc. 20), H. RINK (loc. 37) and D. SCHMIDT (loc. 15).
3) Kryolitselskabet Øresund
R. BØGVAD (loc. 14, 15), K. J. V. STEENSTRUP (loc. 15), N. SØRENSEN (loc. 15), anon. (loc. 15).
4) Mineralogisk Museum
A. JESSEN (loc. 2, 8, 9), M. PORSILD (loc. 95), K. J. V. STEENSTRUP (loc. 1, 5, 6, 86, 103, 106, 107).
5) Nationalmuseets Etnografiske Samling
P. NØRLUND (loc. 12), K. STEPHENSEN (loc. 15).
6) Nordisk Films Kompagni
(loc. 75, 76).
7) Private ownership
R. BESCHEL (loc. 86), J. GEORGI (loc. 75, 76), K. GRIPP (loc. 21, 25, 28), G. HATT (loc. 8), E. KNUTH (loc. 26), E. KOCH (loc. 61), H. KRAGEGAARD (loc. 60), J. NIELSEN (loc. 1), AA. NISSEN (loc. 18, 26), A. STOCKFLETH (loc. 86), B. SØNDERGAARD (loc. 6), H. SØRENSEN (loc. 59) and C. L. VEBÆK (loc. 14).

Other information is derived from the Geodetic Institute's map sheets at 1:250,000 of the northern part of the area made in the 1930's, and the map sheets at 1:50,000 of the U.S. Army for parts of the Julianehåb district, made from aerial photographs in the early 1940's. The abbreviation "Aer. phot." refers to the photographs of the Photogrammetric Section, Geodetic Institute.

Table 2.

I. JULIANEHÅB-GODTHÅB DISTRICTS: LOCAL GLACIATIONS

Loc.no., M_0 and g	Year	a_t	$^0/_0$	Comments on glacier front	Source of information
Loc. 1a	1833	200	21	Calving, before 1833 advancing	Weidick (1959)
$M_0 = 0$ m					
$g = 20^0/_0$	1876	200	21	Calving	Steenstrup (diary)
	1894	200	21	–	Jessen (1896)
	1926	550	58	Front on land	Roussell (1941)
	1942?	900	95	Front calving in lake	Aer. phot.
	1958	950	100	–	Weidick (1963b)
	1961	?	?	"In state of retreat"	Kelly & Wallis (1961)
Loc. 1b	1876	100	7	Calving	Steenstrup (diary)
$M_0 = 0$ m	1881	100	7	–	Holm (1883)
$g = 60^0/_0$	1889	0	0	–	Hartz (1898)
	1894	800	57	–	Jessen (1896)
	1926	800	57	–	Roussell (1941)
	1942?	1400	100	–	Aer. phot.
	1949	1400	100	–	J. Nielsen (phot.)
	1958	1400	100	–	Weidick (1963b)
Loc. 2	1894	100?	22?	Front calving in lake	Jessen (phot.)
$M_0 = 495$ m	1951	300	67	–	Aer. phot.
$g = 20^0/_0$	1957	450	100	Front on land	Fristrup (1961)
Loc. 3	1881	250	17	Calving	Holm (1883)
$M_0 = 0$ m, $g = 10^0/_0$	1953	1500	100	–	Aer. phot.
Loc. 5	1876	0	0	Calving	Steenstrup (diary)
$M_0 = 0$ m	1926	0?	0?	–	Ødum (1927)
$g =$ c. $5^0/_0$	1942?	1500	100	–	Aer. phot.
	1947	1500	100	–	Aer. phot.
	1953	1500	100	–	Aer. phot.
	1957	1500	100	–	Weidick (1963b)
	1960	1500	100	–	Weidick (1963b)
Loc. 6	1876	0	0	Steep front to small lake	Steenstrup (phot.)
$M_0 < 50$ m	1899	100	68	Thinning front. Initial dead ice formation	Steenstrup (phot.)
$g =$ c. $4^0/_0$					
	1942	150	100	Dead ice formation	Aer. phot.
	1950	150	100	–	Søndergaard (phot.)
	1953	150	100	–	Aer. phot.
	1955	150	100	–	Weidick (1963b)
	1958	150	100	–	Weidick (1963b)
	1960	150	100	–	Weidick (1963b)
Loc. 7	1900	100	8	Steep front	Ussing (1912)
$M_0 = 700$ m	1952	1200	96	–	Weidick (1959)
$g = 20^0/_0$	1955	1200	96	–	Weidick (1959)
	1957	1250	100	Thinning	Weidick (1959)

(continued)

Table 2 (continued).

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
Loc. 21	1878	0	0	Calving lobe	Kornerup (1879)
M_0 = 0 m	1930	100	40	Thinning lobe just reaching the sea	Gripp (phot.)
g = 100%	1955	250	100	Front on land	Weidick (1959)
Loc. 25	c. 1890	200	25	Clear, steep front	Møller (phot., sketch)
M_0 = 10 m	1903	200	25	–	Moltke (sketch)
g = 100%	1930	800	100	–	Gripp (phot.)
	1953	800	100	–	Aer. phot.
	1955	800	100	–	Weidick (1959)

II. JULIANEHÅB-GODTHÅB DISTRICTS: INLAND ICE MARGIN

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
Loc. 8	1894	0	0	Calving	Jessen (1896, diary)
M_0 = 0 m	1932	0	0	–	Hatt (phot.)
g = 3%	1947	0	0	–	Aer. phot.
	1953	0	0	–	Aer. phot.
	1955	0	0	Calving, margins expanding	Weidick (1963b)
Loc. 9	1894	2700	60	Calving	Jessen (1896, phot.)
M_0 = 0 m	1947	4000	89	–	Aer. phot.
g = 4–30%	1953	4500	100	–	Aer. phot.
Loc. 10 (Ø)	1876	100	4	Front on land	Steenstrup (1881)
M_0 = 80 m	1944	1940	87	Front calving in lake	Map, aer. phot.
g = 15%	1953	2340	96	–	Aer. phot.
	1957	2440	100	–	Weidick (1963b)
Loc. 10 (V)	1876	100	12	Partly calving, piedmont-formed lobe	Steenstrup (1881)
M_0 = 0 m					
g = 13–25%	1944	700	81	Front on land, tongue-shaped lobe	Map.
	1953	800	93	–	Aer. phot.
	1957	850	99	–	Weidick (1963b)
	1958	860	100	–	Weidick (1963b)
Loc. 11	1890	300	35	Tongue-shaped lobe	Block (1893, phot.)
M_0 = 0 m	1948	800	94	Margin of Inland ice	Aer. phot.
g = 10–20%	1957	850	100	–	Weidick (1963b)
Loc. 12	1854	600	50	Calving	Motzfeldt (map.)
M_0 = 0 m	1890	600	50	–	Bloch (1893)
g = 5–10%	1926	600?	50?	–	Nørlund (phot.)
	1953	1200	100	–	Aer. phot.

(continued)

Table 2 (continued).

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
Loc. 14 A	1880	40	7	Piedmont-formed lobe on land	HOLST (1886)
M_0 = 0 m					
g = *14*–5%	1890	0	0	–	BLOCH (1893)
	1903	0	0	–	BRUUN (1918)
	1938	600	85	Tongue-shaped lobe on land	BØGVAD (phot.)
	1949	700	100	–	BØGVAD (phot.)
	1953	700	100	–	Aer. phot.
	1955	700	100	–	WEIDICK (1959)
Loc. 14 B	1890	0	0	Calving	BLOCH (1893)
M_0 = 0 m	1903	0	0	–	BRUUN (1918)
g = 10–20%	1949	600	92	–	BØGVAD (phot.)
	1951	630	98	–	VEBÆK (phot.)
	1955	650	100	–	WEIDICK (1959)
Loc. 15	1863	0	0	Calving	SCHMIDT (map)
M_0 = 0 m	1869	0	0	–	Phot.
g = 10%	1871	200	10	–	Phot.
	1899	400	20	–	STEENSTRUP (phot.)
	1912	500	25	–	STEPHENSEN (phot.)
	1923	750	38	–	Phot.
	1927	1100	55	–	N. SØRENSEN (phot.)
	1945	1800	90	–	BØGVAD (phot.)
	1950	1950	98	–	BØGVAD (phot.)
	1955	2000	100	–	WEIDICK (1959)
Loc. 18	1919	75	13	Calving	BENDIXEN (1921)
M_0 = 0 m	1921	75	13	–	NISSEN (phot.)
g = 50%	1955	575	100	Calving, front partly on land	WEIDICK (1959)
Loc. 20	1859	0	0	Calving	PETERS (map)
M_0 = 0 m	1878	0	0	–	KORNERUP (1879)
g = 50–100%	1948	c. 2000	100	–	Aer. phot.
Loc. 23 – S	1729	0	0	Front on land	PAARS (1936)
M_0 = 360 m	1776	0?	0?	–	THORHALLESEN, in WEIDICK (1959)
g = 8%	1888	1200	44	Front calving in lake	NANSEN (1890)
	1937	2400	87	–	Aer. phot.
	1948	2450	89	–	Aer. phot.
	1955	2750	100	–	Aer. phot.
Loc. 23 – N	1729	0	0	Front on land	PAARS (1936)
M_0 = 360 m	1776	0?	0?	–	THORHALLESEN (1914)
g = 8%	1888	1200	55	Front calving in lake	NANSEN (1890)
	1937	1700	77	–	Aer. phot.
	1948	2150	94	–	Aer. phot.
	1955	2200	100	–	WEIDICK (1959)

(continued)

Table 2 (continued).

Loc.no., M_0 and g	Year	a_t	$^0/_0$	Comments on glacier front	Source of Information
Loc. 26	1808	5500	32	Calving	Giesecke (1910)
$M_0 = 0$ m	1850/60	11000	65	–	Weidick (1959)
$g = 4–17^0/_0$	1885	11000	65	–	J. A. D. Jensen (1889)
	1903	11000	65	–	Bruun (1918)
	1921	15000	88	–	Nissen (phot.)
	1932	15000	88	–	Knuth (phot.)
	1937	15000	88	–	Aer. phot.
	1948	16500	97	–	Aer. phot.
	1955	17000	100	–	Aer. phot.
Loc. 28	1855	0	0	Calving	Kleinschmidt (map)
$M_0 = 0$ m	1903	0	0	–	Bruun (1918)
$g = 3^0/_0$	1916	0	0	–	Bendixen (1921)
	1930	0	0	–	Gripp (phot.)
	1937	0	0	–	Aer. phot.
	1942?	0	0	–	Aer. phot.
	1955	0	0	–	Weidick (1959)

III. SUKKERTOPPEN-EGEDESMINDE DISTRICTS: INLAND ICE MARGIN

Loc.no., M_0 and g	Year	a_t	$^0/_0$	Comments on glacier front	Source of Information
Loc. 31	1885	0?	0?	Calving	J. A. D. Jensen (1889)
$M_0 = 100$ m	1936	0	0	Alluvial plain before front	Aer. phot.
$g = 2–10^0/_0$	1942?	0	0	–	Aer. phot.
Loc. 32	1885	0	0	Piedmont-shaped lobe on land	J. A. D. Jensen (1889)
$M_0 = 20$ m					
$g = 50^0/_0$	1936	1800	90	Tongue-shaped lobe calving in lake	Aer. phot.
	1948	2000	100	Part of front calving in lake	Aer. phot.
Loc. 33	1863	0	0		Kreutzmann (map)
$M_0 = 680$ m	1936	160	80	Calving in lake	Aer. phot.
$g = 3–50^0/_0$	1948	180	90	–	Aer. phot.
	1952	200	100	–	Aer. phot.
Loc. 33 (B_3)	1884	0	0	Front calving in lake	J. A. D. Jensen (1889)
$M_0 =$ c. 100 m	1936	150	100	–	Aer. phot.
$g = 3–25^0/_0$	1952	150	100	–	Aer. phot.
	1959	150	100	–	Weidick (phot.)
Loc. 35	c. 1870	0	0		J. A. D. Jensen (1889)
$M_0 = 50$ m	1909	0	0		O. Nordenskiold (1914)
$g = 10^0/_0$	1942?	0	0	Clear ice to front	Aer. phot.
	1959	0	0	–	Aer. phot.

(continued)

Table 2 (continued).

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
Loc. 36	1879	100	33		J. A. D. JENSEN (1881)
M_0 = 100 m	1936	300	100	Thin moraine cover	Aer. phot.
g = 5%	1942?	300	100	–	Aer. phot.
Loc. 37	1851?	0?	0?	Calving	RINK (map)
M_0 = 0 m	1879	0?	0?	–	J. A. D. JENSEN (1881)
g = 8–12%	1911	150	14	–	NORDMANN (1961)
	1936	1040	95	–	Aer. phot.
	1942?	1060	96	–	Aer. phot.
	1961	1100	100	–	Aer. phot.
Loc. 38	1931/33	0	0		Map
M_0 = 100 m	1936	250	56	Thin moraine cover	Aer. phot.
g = 10%	1942?	450	100	–	Aer. phot.
Loc. 39	1851	0	0	Calving	RINK (1857)
M_0 = 0 m	1870	0	0	–	A. NORDENSKIÖLD (1871)
g = 1–5%	1883	20	33	–	A. NORDENSKIÖLD (1886)
	1924	30	50	Thinning, alluvial plain before parts of front	PORSILD (phot.)
	1931/33	40	67	–	Map
	1936	50	83	–	BANG (phot.)
	1948	60	100	–	Aer. phot.

IV. SUKKERTOPPEN DISTRICT: LOCAL GLACIATIONS

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
Loc. 40	c. 1860	0	0		KREUTZMANN (map)
M_0 = 0 m	1884	0	0		J. A. D. JENSEN (1889)
g = 20%	1925	150	14	Thinning front, dead ice formation	SCHEEL (1927)
	1937	1000	90		Aer. phot.
	1956	1110	100	Thinning	WEIDICK (phot.)
	1960	1110	100	–	WEIDICK (phot.)
Loc. 41	c. 1860	0	0		KREUTZMANN (map)
M_0 = 0 m	1885	0	0		J. A. D. JENSEN (1889)
g = 12%	1912/15	50	10		BOBÉ (1921)
	1920	50	10	Thinning	ROSING (1958)
	1936	490	96	Partly moraine covered	Aer. phot.
	1942?	490	96	–	Aer. phot.
	1956	500	98	–	WEIDICK (phot.)
	1960	510	100	Initial dead ice formation	WEIDICK (phot.)

(continued)

Table 2 (continued).

Loc.no., M_0 and g	Year	a_t	$^0/_0$	Comments on glacier front	Source of information
Loc. 42	1885	0?	0?		J. A. D. JENSEN (1889)
M_0 = 0 m	1936	1900	95	Thinning? clear ice to front	Aer. phot.
g = 10 $^0/_0$	1942?	1900	95	–	Aer. phot.
	1956	2000	100	Thinning, clear ice to front	WEIDICK (phot.)
	1960	2000	100	–	WEIDICK (phot.)
Loc. 43	1885	0	0	Calving	J. A. D. JENSEN (1889)
M_0 = 0 m	1936	1000	100	Thinning?	Aer. phot.
g = 14–17 $^0/_0$	1942?	1000	100	–	Aer. phot.
	1956	1000	100	–	WEIDICK (phot.)
	1960	1000	100	Thinning, clear ice to front	WEIDICK (phot.)
Loc. 44	1885	0?	0?		J. A. D. JENSEN (1889)
M_0 = 200 m	1936	300	100	Thinning, clear ice to front	Aer. phot.
g = 50 $^0/_0$	1960	300	100	–	WEIDICK (phot.)
Loc. 47 B	1885	0	0		J. A. D. JENSEN (1889)
M_0 = 0 m	1936	550	98	Thinning, clear ice to front	Aer. phot.
g = 30 $^0/_0$	1960	560	100	–	WEIDICK (phot.)
Loc. 47 E	1885	10	1		J. A. D. JENSEN (1889)
M_0 = 0 m	1909	150	16	Thinning	O. NORDENSKIÖLD (1914)
g = 30–40 $^0/_0$	1936	560	62	Thinning, clear ice to front	Aer. phot.
	1942?	860	94	–	Aer. phot.
	1960	910	100	–	WEIDICK (phot.)
Loc. 48 A	1909	50?	7?	Clear ice to front	O. NORDENSKIÖLD (1914)
M_0 = 20 m	1926/27	?	?		BELKNAP (1941)
g = 100 $^0/_0$	1936	550	79	Thinning, clear ice to front	Aer. phot.
	1942?	620	88	–	Aer. phot.
	1961	700	100	–	WEIDICK (phot.)
Loc. 48 B	1909	100	20		O. NORDENSKIÖLD (1914)
M_0 = 150 m					
g = ?	1936	200	40		Aer. phot.
	1961	500	100	Glacier transformed to rock glacier	WEIDICK (phot.)

(continued

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
Loc. 48 C	1909	320	62	Thinning,	O. NORDENSKIÖLD
M_0 = 125 m				clear ice to front	(1914)
g = *33*–100%	1936	410	78	–	Aer. phot.
	1942?	440	84	–	Aer. phot.
	1960	520	100	–	WEIDICK (phot.)
Loc. 49	c. 1860	0	0	Calving	KREUTZMANN (map)
M_0 = 0 m	1942?	350	100	–	Aer. phot.
g = 17%					
Loc. 50 N	c. 1860	2100	68	Calving	KREUTZMANN (map)
M_0 = 0 m	1936	2400	77	–	Aer. phot.
g = *14*–7%	1942?	3100	100	–	Aer. phot.
Loc. 51 S	c. 1860	250	9	Calving	KREUTZMANN (map)
M_0 = 0 m	1884	250	9	–	J. A. D. JENSEN (1889)
g = *20*–7%	1930	1950	68	–	Phot.
	1936	2350	82	–	Aer. phot.
	1942?	2700	95	–	Aer. phot.
	1960	2850	100	Front mainly resting	WEIDICK (phot.)
				on land	
Loc. 52 N	c. 1860	500?	33?	Calving	KREUTZMANN (map)
M_0 = 0 m	1930	1500	100	–	Phot.
g = *17*–6%	1936	1500	100	Calving, thinning	Aer. phot.
	1942?	1500	100	–	Aer. phot.
	1960	1500	100	–	WEIDICK (phot.)
Loc. 52 S	c. 1860	1000?	100?	Calving	KREUTZMANN (map)
M_0 = 0 m	1930	1000	100	–	Phot.
g = *17*–5%	1936	1000	100	Calving, thinning?	Aer. phot.
	1938	1000	100	Calving, thinning	ETIENNE (1940)
	1942?	1000	100	–	Aer. phot.
	1956	1000	100	–	HOLLAND (1958, 1961)
	1958	1000	100	–	BØCHER (1961)
	1960	1000	100	–	HOFF (1961)
Loc. 53 I–N	c. 1860	200	20		KREUTZMANN (map)
M_0 = 0 m	1930	1000	100	Thinning	Phot.
g = 17–25%	1936	1000	100	–	Aer. phot.
	1942?	1000	100	–	Aer. phot.
	1960	1000	100	–	WEIDICK (phot.)
Loc. 53 III–N	1885	0	0	Calving	J. A. D. JENSEN (1889)
M_0 = 0 m	1930	50	12	Front just reaching	Phot.
g = 50%				sea, thinning	
	1936	300	75	Front on land, thinning	Aer. phot.
	1942?	350	88	Thinning	Aer. phot.
	1960	400	100	–	WEIDICK (phot.)

(continued)

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	$^0/_0$	Comments on glacier front	Source of information
Loc. 53 IV–N	1930	900	90	Thinning	Phot.
M_0 = 200 m	1936	900	90	–	Aer. phot.
g = 16 $^0/_0$	1942?	900	90	–	Aer. phot.
	1960	1000	100	–	WEIDICK (phot.)
Loc. 53 IV–S	1884	20?	5?	Calving?	J. A. D. JENSEN (1889)
M_0 = 0 m	1902	20	5	Calving	MYLIUS-ERICHSEN & MOLTKE (1906)
g = 6 $^0/_0$	1936	320	80	Front on land, dead ice formation	Aer. phot.
	1942?	360	90	Dead ice formation	Aer. phot.
	1960	400	100	–	WEIDICK (phot.)
Loc. 53 V–S	1902	0	0	Calving	MYLIUS-ERICHSEN & MOLTKE (1906)
M_0 = 0 m					
g = 100 $^0/_0$	1936	0	0	–	Aer. phot.
	1960	0	0	Front just reaching sea, thinning	WEIDICK (phot.)
Loc. 54–1	1884	0	0	Calving	J. A. D. JENSEN (1889)
M_0 = 0 m	1930	20	7	–	Phot.
g = 50 $^0/_0$	1936	20	7	–	Aer. phot.
	1942?	90	33	–	Aer. phot.
	1958	240	89	Thinning, just reaching sea	BØCHER (1961)
	1960	270	100	Thinning, clear ice to front	WEIDICK (phot.)
Loc. 54–2	1884	20	8	Calving	J. A. D. JENSEN (1889)
M_0 = 0 m	1930	170	68	–	Phot.
g = 50 $^0/_0$	1942?	220	88	–	Aer. phot.
	1958	250	100	–	BØCHER (1961)
	1960	250	100	–	WEIDICK (phot.)
Loc. 55–1	c. 1860	0	0	Calving	KREUTZMANN (map)
M_0 = 0 m	1884	0	0	–	J. A. D. JENSEN (1889)
g = 20 $^0/_0$	1930	100	30	Front on land, thinning	Phot.
	1936	150	50		Aer. phot.
	1942?	300	100		Aer. phot.
	1960	300	100		WEIDICK (phot.)
Loc. 56–1	c. 1860	50?	9?	Calving	KREUTZMANN (map)
M_0 = 0 m	1884	50?	9?	–	J. A. D. JENSEN (1889)
g = 30–50 $^0/_0$	1936	500	91	–	Aer. phot.
	1942?	520	94	–	Aer. phot.
	1960	550	100	Half part on land	WEIDICK (phot.)
Loc. 56–3	1884	50?	2?	Calving	J. A. D. JENSEN (1889)
M_0 = 0 m	1942?	3000	100	–	Aer. phot.
g = 8 $^0/_0$					

(continued)

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	$^0/_0$	Comments on glacier front	Source of information
Loc. 57 V–N	1884	0	0		J. A. D. JENSEN (1889)
M_0 = 200 m	1936	300	100		Aer. phot.
g = 50 $^0/_0$	1942?	300	100	Clear ice to front	Aer. phot.
	1948	300	100	–	Aer. phot.
	1952	300	100	–	Aer. phot.
	1960	300	100	–	WEIDICK (phot.)
Loc. 57 VI–N	c. 1860	1000?	19?	Calving, piedmont lobe	KREUTZMANN (map)
M_0 = 0 m	1884	1800	35	–	J. A. D. JENSEN (1889)
g = *30–5* $^0/_0$	1936	1800	35	–	Aer. phot.
	1942?	2650	51	–	Aer. phot.
	1948	4150	80	–	Aer. phot.
	1960	5150	100	Tongue-shaped lobe on land	WEIDICK (phot.)
Loc. 57 I–S	1885	1800	69	Calving, piedmont lobe	J. A. D. JENSEN (1889)
M_0 = 0 m	1936	1800	69	–	Aer. phot.
g = 100 $^0/_0$	1942?	2650	100	–	Aer. phot.
	1952	2650	100	Calving tongue	Aer. phot.
	1960	2650	100	–	WEIDICK (phot.)
Loc. 58	1884	50	4		J. A. D. JENSEN (1889)
M_0 = 30–40 m	1936	1050	87		Aer. phot.
g = *25*–50 $^0/_0$	1942?	1200	100	Moraine cover on half of front	Aer. phot.
	1961	1200	100	–	WEIDICK (phot.)
Loc. 59	c. 1863	30?	5?	Calving	KREUTZMANN (map)
M_0 = 0 m	1885	30	5	–	J. A. D. JENSEN (1889)
g = *12*–5 $^0/_0$	1912	30	5	Calving, thinning	DE QUERVAIN (1925)
	1936	250	45	–	Aer. phot.
	1942?	550	100	–	Aer. phot.
	1948	550	100	–	H. SØRENSEN (phot.)
Loc. 60	1912	250	13	Calving, 1885–1912 expanding?, 1912 expanding	BENDIXEN (1921), DE QUERVAIN (1925)
M_0 = 0 m					
g = *11*–6 $^0/_0$	1936	1000	54	Calving, thinning	Aer. phot.
	1942?	1750	94	–	Aer. phot.
	1955	1850	99	–	KRAGEGAARD (phot.)
	1957	1850	99	–	KRAGEGAARD (phot.)
	1960	1860	100	–	WEIDICK (phot.)
Loc. 60a	1912	150?	75?	Calving, thinning	DE QUERVAIN (1925)
M_0 = 0 m	1936	180	90	–	Aer. phot.
g = 50 $^0/_0$	1942?	180	90	–	Aer. phot.
	1955	200	100	–	KRAGEGAARD (phot.)
	1960	200	100	–	WEIDICK (phot.)

(continued)

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
Loc. 61 I	1885	50?	11?		J. A. D. JENSEN (1889)
M_0 = 50 m	1893	100?	22?		KRABBE (1929)
g = 25%	1936	300	67	Thinning, clear ice to front	Aer. phot.
	1942	400	89	–	Aer. phot.
	1960	450	100	–	WEIDICK (phot.)
Loc. 61 II	1885	40	80	Calving?	J. A. D. JENSEN (1889)
M_0 = 0 m	1936	50	100	Thinning, clear ice to front	Aer. phot.
g = 25–50%	1942?	50	100	–	Aer. phot.
	1960	50	100	–	WEIDICK (phot.)
Loc. 61 III	1936	450	90	Thinning, clear ice to front	Aer. phot.
M_0 = 0 m					
g = 50%	1942?	450	90	–	Aer. phot.
	1960	500	100	–	WEIDICK (phot.)
Loc. 61 IVA	c. 1860	0	0	Calving?	KREUTZMANN (map)
M_0 = 60 m	1885	0	0		J. A. D. JENSEN (1889)
g = 30%	1942?	400	89	Thinning, clear ice to front	Aer. phot.
	1960	450	100	–	WEIDICK (phot.)
Loc. 61 V	c. 1860	20?	5?	Calving?	KREUTZMANN (map)
M_0 = 0 m	1885	20?	5?		J. A. D. JENSEN (1889)
g = 30–50%	1942?	320	80	Thinning, clear ice to front	Aer. phot.
	1951	370	93	–	E. KOCH (phot.)
	1960	400	100	–	WEIDICK (phot.)
Loc. 61 VI	1885	35	14	Calving?	J. A. D. JENSEN (1889)
M_0 = 0 m	1942?	235	94	Thinning, clear ice to front	Aer. phot.
g = 33%					
	1951	250	100	–	E. KOCH (phot.)
	1960	250	100	–	WEIDICK (phot.)
Loc. 62	1885	50	7	Damming a lake	J. A. D. JENSEN (1889)
M_0 = 0 m	1936	350	47	Thinning, clear ice to front	Aer. phot.
g = 20%	1942?	450	60	–	Aer. phot.
	1961	750	100	–	BESCHEL & WEIDICK (phot.)
Loc. 63	1885	90?	6?		J. A. D. JENSEN (1889)
M_0 = 50 m	1936	1590	99	Thinning, clear ice to front	Aer. phot.
g = 20%					
	1948	1600	100	–	Aer. phot.

(continued)

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	$^0/_0$	Comments on glacier front	Source of information
Loc. 64	1885	0	0	Piedmont-like lobe	J. A. D. JENSEN (1889)
M_0 = 150 m	1936	1800	95	Tongue-shaped lobe	Aer. phot.
g = 10 $^0/_0$	1948	1900	100	–	Aer. phot.
Loc. 65 A	c. 1860	10	5	Damming Sarfartôq river	KREUTZMANN (map)
M_0 = 650 m					
g = 10 $^0/_0$	1936	190	95	Thinning	Aer. phot.
	1948	200	100	–	Aer. phot.
Loc. 65 B	1909	15	10	Calving in lake	O. NORDENSKIÖLD (1914)
M_0 = 980 m					
g = 12 $^0/_0$	1936	25	17	–	Aer. phot.
	1948	150	100	–	Aer. phot.

V. DISKO BUGT: INLAND ICE MARGIN

Loc.no., M_0 and g	Year	a_t	$^0/_0$	Comments on glacier front	Source of information
Loc. 66 A	1902	1500	100	Calving in lake	ENGELL (1904)
M_0 = 77 m	1942?	600	40	–	Aer. phot.
g = 6 $^0/_0$	1948	600	40	–	Aer. phot.
Loc. 66 C	1902	500	33	Calving in lake	ENGELL (1904)
M_0 = 77 m	1942?	1500	100	Isolated from lake by alluvial plain	Aer. phot.
g = 25–5 $^0/_0$					
	1948	1500	100	–	Aer. phot.
Loc. 67	1902	0	0	Calving in lake, slightly thinning	ENGELL (1904)
M_0 = 200 m					
g = 5–7 $^0/_0$	1942?	0	0	–	Aer. phot.
	1948	0	0	–	Aer. phot.
Loc. 68	1880	1000?	100?	Calving, thinning slightly since 1850	HAMMER (1883)
M_0 = 0 m					
g = 3 $^0/_0$	1903	100	10	Calving, expanding	ENGELL (1910)
	1913	100	10	Calving	KOCH & WEGENER (1930)
	1931/33	100	10	–	Map
	1948	100	10	–	Aer. phot.
	1949	100	10	–	Aer. phot.
	1953	100	10	–	Aer. phot.
	1959	100	10	–	Aer. phot.
	1963	100	10	–	WEIDICK (phot.)

(continued)

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	$^0/_0$	Comments on glacier front	Source of information
Loc. 69	1851	100		Calving	RINK (1857)
$M_0 = 0$ m	1880	100		Calving, slightly thinning?	HAMMER (1883)
$g = 3^0/_0$	1903	100		Calving	ENGELL (1910)
	1913	100		–	KOCH & WEGENER (1930)
	1931/33	100		–	Map
	1948	100		–	Aer. phot.
	1949	100		–	Aer. phot.
	1953	100		–	Aer. phot.
	1959	100		–	Aer. phot.
	1963	100		Calving, margin slightly expanding	WEIDICK (phot.)
Loc. 70	1851	200	1	Calving	RINK (1857)
$M_0 = 0$ m	1870	0	0	–	A. NORDENSKIÖLD (1871)
$g =$ c. $2^0/_0$	1875	5000	19	–	HELLAND (1876)
	1879	7000	27	–	HAMMER (1883)
	1880	7500	29	–	HAMMER (1883)
	1883	8500	33	–	HAMMER (1889)
	1893	9300	36	–	DRYGALSKI (1897)
	1902	12500	48	–	ENGELL (1904)
	1913	17000	65	–	KOCH & WEGENER (1930)
	1929	18500	71	–	WEGENER (1930)
	1931	19700	76	–	Map
	1948	26000	100	–	Aer. phot.
	1953	26000	100	–	Aer. phot.
	1959	26000	100	–	Aer. phot.
	1963	24000	92	–	WEIDICK (phot.)
	1964	c. 24000	92	–	Aer. phot.
Loc. 70 (Tivss.)	1879	200	17	Calving in lake	HAMMER (1883)
$M_0 = 43$ m	1902	200	17	Thinning	ENGELL (1904)
$g = 3–6^0/_0$	1913	400?	33?		KOCH & WEGENER (1930)
	1929	1000	83	Thinning, calving in lake	WEGENER (1930)
	1931	1000	83	–	Map
	1948	1200	100	–	Aer. phot.
	1953	1200	100	–	Aer. phot.
	1959	1200	100	–	Aer. phot.
	1963	1200	100	–	WEIDICK (phot.)
	1964	1200	100	–	Aer. phot.

(continued)

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	$^0/_0$	Comments on glacier front	Source of information
Loc. 71	1851	100	12	Calving	RINK (1857)
M_0 = 0 m	1902	100	12	–	ENGELL (1904)
g = *8–4* $^0/_0$	1931/33	100	12	–	Map
	1949	800	100	–	Aer. phot.
	1953	800	100	–	Aer. phot.
	1959	800	100	–	Aer. phot.
	1964	800	100	–	Aer. phot.
Loc. 71 B	1902	25	5		ENGELL (1904)
M_0 = c. 50 m	1931/33	325	65		Map
g = *8–4* $^0/_0$	1949	350	70	Clear ice to front	Aer. phot.
	1953	400	80	–	Aer. phot.
	1959	500	100		Aer. phot.
	1964	500	100		Aer. phot.
Loc. 72	1851	380	100		RINK (1857)
M_0 = 10 m	1883	0	0		HAMMER (1889)
g = *12–6* $^0/_0$	1904	155	41		ENGELL (1910)
	1931/33	155	41	Readvance between 1904 and 1933?	Map
	1949	320	84	Clear ice to front	Aer. phot.
	1953	330	87	–	Aer. phot.
	1959	380	100		Aer. phot.
	1964	380	100		Aer. phot.
Loc. 73	1903	c. 700	87	Thinning?	ENGELL (1910)
M_0 = 0 m	1931/33	c. 700	87		Map
g = *5–12* $^0/_0$	1949	c. 700	87	Thinning, clear ice to front	Aer. phot.
	1953	c. 800	100	–	Aer. phot.
	1959	c. 800	100	Thinning	Aer. phot.
	1964	c. 800	100	–	Aer. phot.
Loc. 74A (Eqip s.)	1851	100	5	Calving	RINK (1857)
M_0 = 0 m	1883	100	5	–	HAMMER (1889)
g = *8–3* $^0/_0$	1903	100	5	–	ENGELL (1910)
	1912/13	500	23	Calving, receding before 1912, readvance begins 1912	DE QUERVAIN (1925)
	1923	100	5	Maximum extent c. 1920, calving	BAUER (1955c)
	1929	1100	50	Calving	WEGENER (1930)
	1932	1200	55	–	Map
	1948	2200	100	–	BAUER (1955c)
	1953	2100	95	–	BAUER (1955c)
	1959	2000	91	–	Aer. phot.
	1964	1800	82	–	Aer. phot.

(continued)

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
Loc.74B(Kang.s.)	1851	100?	7?	Calving	RINK (1857)
$M_0 = 0$ m	1883	100?	7?	–	HAMMER (1889)
g = 2%	1903	1200	86	–	ENGELL (1910)
	1931/33	1200	86	–	Map
	1948	1400	100	–	Aer. phot.
	1953	1400	100	–	Aer. phot.
	1959	800	57	–	Aer. phot.
	1964	300	21	–	Aer. phot.
Loc. 75	1851	20?	1?	Calving	RINK (1857)
$M_0 = 0$ m	1880	700?	35?	–	STEENSTRUP (1883a)
g = 8–2%	1903	1200	60	–	ENGELL (1910)
	1929	1300	65	–	GEORGI (phot.)
	1949	2000	100	–	Aer. phot.
	1953	2000	100	–	Aer. phot.
	1956	2000?	100?	–	Phot.
	1959	2000?	100?	–	Aer. phot.
	1961	2000?	100?	–	WEIDICK (phot.)
	1964	1300	65	–	Aer. phot.
Loc. 76	1851	100		Calving	RINK (1857)
$M_0 = 0$ m	1903	200		–	ENGELL (1910)
g = 7–2%	1929	200		–	GEORGI (phot.)
	1931/33	200		–	Map
	1949	200		–	Aer. phot.
	1953	200		–	Aer. phot.
	1956	200		–	Phot.
	1959	200		–	Aer. phot.
	1961	200		–	WEIDICK (phot.)
	1964	200		–	Aer. phot.

VI. DISKO ISLAND: LOCAL GLACIATIONS

Loc. 78A	1898	4500	78		STEENSTRUP (1901)
$M_0 = 100$ m	1932/33	4750	83		Map
g = c. 20%	1953	5750	100	Thinning, moraine	Aer. phot.
(before 7%)				covered	
Loc. 78B–C	1898	c. 4500	75		STEENSTRUP (1901)
$M_0 = 100$ m	1932/33	5000	83		Map
g = c. 20%	1953	c. 6000	100	Thinning, moraine	Aer. phot.
(before c. 7%)				covered	
Loc. 79A	1898	200	36		STEENSTRUP (1901)
$M_0 = 650$ m	1932/33	500	91		Map
g = c. 25%	1942?	550	100	Thinning	Aer. phot.
	1953	550	100	–	Aer. phot.
	1961	550	100	–	BESCHEL (phot.)

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
Loc. 79 B	1898	200	36	Possibly new glacier advancing over dead ice	STEENSTRUP (1901)
$M_0 = 400$					
$g = 20\%$					
	1932/33	500	91		Map
	1942?	550	100	Thinning, moraine covered	Aer. phot.
	1953	550	100	–	Aer. phot.
	1961	550	100	–	WEIDICK (phot.)
Loc. 79 C	1898	200	12	Thinning, clear ice to front	STEENSTRUP (1901)
$M_0 = 400$ m					
$g = 12\%$	1942?	1200	75	–	Aer. phot.
	1953	1400	87	–	Aer. phot.
	1961	1600	100	–	WEIDICK (phot.)
Loc. 83	1898	300	12		STEENSTRUP (1901)
$M_0 = 50$ m	1931/33	600?	23?		Map
$g = 5–10\%$	1942?	2000	77	Thinning	Aer. phot.
	1953	2600	100	Thinning, partly moraine covered	Aer. phot.
Loc. 84	1898	100	10	Thinning	STEENSTRUP (1901)
$M_0 = 300$ m	1942?	650	65	Thinning, clear ice to front	Aer. phot.
$g = 17\%$					
	1953	1000	100	–	Aer. phot.
	1962	1000	100	–	Aer. phot.
	1963	1000	100	–	WEIDICK (phot.)
Loc. 85 B	1848	0	0		RINK (1857)
$M_0 = 350$ m	1893	0	0	Thinning, clear ice to front	CHAMBERLIN (1894–96)
$g = 12\%$					
	1913	100	8	–	L. KOCH (phot.)
	1931/33	800	67		Map
	1942?	850	71	Thinning	Aer. phot.
	1953	1150	96	–	Aer. phot.
	1961	1200	100	Thinning, clear ice to front	WEIDICK (phot.)
	1962	1200	100	–	Aer. phot.
Loc. 85 A	1893	0	0		CHAMBERLIN (1894–96)
$M_0 = 400$ m	1942?	300	75	Thinning	Aer. phot.
$g = 25\%$	1953	350	88	–	Aer. phot.
	1961	400	100	Thinning, clear ice to front	WEIDICK (phot.)
	1962	400	100	–	Aer. phot.

(continued)

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	$^0/_0$	Comments on glacier front	Source of information
Loc. 86 – 1	1812	0?	0?		GIESECKE (1910)
M_0 = 400–450 m	1848	0?	0?		RINK (1857)
g = c. 20 $^0/_0$	1870	?	?	Advance since RINK's visit	A. NORDENSKIÖLD (1871)
	1893	400	21	Thinning, clear ice to front	CHAMBERLIN (1894–96)
	1897	420	22	–	PJETURSS (1898)
	1898	420	22	–	STEENSTRUP (1901, phot.)
	1912	540	28	–	DE QUERVAIN (1925)
	1923	575	30	Thinning	FRODA (1925)
	1931/33	900	46		Map
	1942?	1100	56	Thinning, clear ice to front	Aer. phot.
	1953	1200	62	Thinning. Lowermost 400 m partly separated as dead ice body	Aer. phot.
	1956	1900	97	Thinning. Lowermost 400 m separated totally from front. New front formed	WEIDICK (phot.)
	1960	1925	99	Thinning, clear ice to front	STOCKFLETH (phot.)
	1961	1950	100	–	WEIDICK (phot.)
	1962	1950	100	–	Aer. phot.
Loc. 86 – 2	1893	0	0	Thinning	CHAMBERLIN (1894–96)
M_0 = 450 m	1897	20	2	Clear ice to front	PJETURSS (1898)
g = 14–25 $^0/_0$	1898	20	2	–	STEENSTRUP (phot.)
	1912	100	12	Thinning	DE QUERVAIN (1925)
	1923	200	24	–	FRODA (1925)
	1931/33	200?	24?		Map
	1942?	600	71	Thinning, clear ice to front	Aer. phot.
	1953	700	82	Thinning	Aer. phot.
	1961	850	100	–	BESCHEL (phot.)
Loc. 86 – 3	1848	c. 1800?	98?		RINK (1857)
M_0 = 150 m	1893	50	27	Thinning	CHAMBERLIN (1894–96)
g = c. 14 $^0/_0$	1898	50	27	–	STEENSTRUP (phot.)
	1913	800	44	Thinning, clear ice to front	L. KOCH (phot.)
	1942?	1650	90	–	Aer. phot.
	1953	1800	98	–	Aer. phot.
	1961	1830	100	–	BESCHEL (phot.)

(continued)

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	$^0/_0$	Comments on glacier front	Source of information
Loc. 86 – 4	1897	50	6	Thinning	PJETURSS (1898)
M_0 = 350 m	1898	50	6	Thinning, clear ice to front	STEENSTRUP (1901)
g = 10–14 $^0/_0$	1912	70	9	Thinning	DE QUERVAIN (1925)
	1923	110	14	–	FRODA (1925)
	1931/33	110?	14?		Map
	1953	750	96	Thinning	Aer. phot.
	1961	780	100	Thinning, thin moraine cover	BESCHEL (phot.)
Loc. 88	1811	0	0		GIESECKE (1910)
M_0 = 0 m	1815	0	0		GIESECKE (1910)
g = ?	1849	0?	0?		RINK (1857)
	1879/80	0?	0?	Thinning, dead ice formation	STEENSTRUP (1883 c)
	1898	200?	16?		STEENSTRUP (1901)
	1931/33	1250	100	Glacier nearly disappeared	Map
	1949	1250	100	–	Aer. phot.
	1953	1250	100	–	Aer. phot.
Loc. 95	1813	0	0	Steep front?	GIESECKE (1910)
M_0 = 10 m	1880	0?	0?		STEENSTRUP (1901)
g = c. 25 $^0/_0$	1898	0?	0?		STEENSTRUP (1901)
	1902	100	6	Thinning, clear ice to front	PORSILD (phot.)
	1931/33	900	56		Map
	1942?	1500	94	Thinning, clear ice to front, prominent medial moraines developed	Aer. phot.
	1949	1600	100	–	Aer. phot.
	1953	1600	100	–	Aer. phot.
Loc. 96	1880	0	0		STEENSTRUP (1883 a)
M_0 = 50 m	1942?	1500	94	Thinning, dead ice formation	Aer. phot.
g = 30–50 $^0/_0$	1953	1600	100	–	Aer. phot.

(continued)

(Table 2 continued)

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
VII. NÛGSSUAQ PENINSULA: LOCAL GLACIATIONS					
Loc. 103	1849	0?	0?		Rink (1857)
$M_0 = 250$ m	1869	0?	0?	Before 1869 advancing. 1869 thinning	Whymper (Heer 1869)
$g = 20\%$	1898	0?	0?	Thinning	Steenstrup (1883a, phot.)
	1903	20	2	–	Engell (1910)
	1942?	500	59	Thinning, clear ice to front	Aer. phot.
	1948	800	94	–	Aer. phot.
	1953	850	100	–	Aer. phot.
	1961	850	100	–	Weidick (phot.)
Loc. 106	1811	0?		Dead ice formation	Giesecke (1910)
$M_0 = 200$	1850	0?		–	Rink (1852, 1857)
$g = 7\%$	1880	0?		Readvance before 1879	Steenstrup (1883a, phot.)
	1893	0?		Dead ice formation	Drygalski (1897)
	1939	250		–	Laursen (1944)
	1942?	250		–	Aer. phot.
	1948	250		–	Aer. phot.
	1953	250		–	Aer. phot.
	1956	250		–	Weidick (phot.)
Loc. 107	1850			Dead ice formation	Rink (1852, 1857)
$M_0 = 200$ m	1875			–	Helland (1876)
$g = 8–10\%$	1880			1879: Steep front with clear ice	Steenstrup (1883a)
	1893			Dead ice formation	Drygalski (1897)
	1929			–	Loewe (1935)
	1942?			–	Aer. phot.
	1948			–	Aer. phot.
	1953			–	Aer. phot.
	1956			–	Weidick (phot.

(continued)

(Table 2 continued)

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
Loc. 108	1811	0	0	Calving	GIESECKE (1910)
$M_0 = 0$ m	1850	0	0	–	RINK (1857)
g = 5–20%	1875	0	0	–	HELLAND (1876)
	1880	0	0	–	STEENSTRUP (1883a, phot.)
	1893	10	2	Calving, thinning	DRYGALSKI (1897)
	1915			Under advance c. 1915	LOEWE (1934)
	1929	470	70	Under advance, steep front	LOEWE (1935)
	1932	300	45	–	LOEWE (1934)
	1948	500	75	Steep front	Aer. phot.
	1953	550	82	Steep front, clear ice to front	Aer. phot.
	1956	670	100	–	WEIDICK (phot.)
Loc. 109	1850	250	10	Thinning	RINK (1857)
$M_0 = 0$ m	1875	500	20	Dead ice formation	HELLAND (1876)
g = 12–14%	1879	c. 1000	39	Expanding	STEENSTRUP (1883a)
	1880	1000	39	–	STEENSTRUP (1883a)
	1893	20	1	Calving	DRYGALSKI (1897)
	1929	1700	67	1915 under recession	LOEWE (1935)
	1932	1800	70	Thinning	LOEWE (1934)
	1948	2450	96		Aer. phot.
	1953	2550	100		Aer. phot.
	1956	2400	94	Steep front	WEIDICK (phot.)
Loc. 110	1850	0	0	Calving	RINK (1857)
$M_0 = 0$ m	1875	0	0	–	HELLAND (1876)
g = 10%	1880	0	0	–	STEENSTRUP (1883a)
	1893	0?	0?	–	DRYGALSKI (1897)
	1932	100	10	Front on land, thinning	LOEWE (1934)
	1948	850	85	Thinning	Aer. phot.
	1953	925	93	–	Aer. phot.
	1956	1000	100	–	WEIDICK (phot.)
Loc. 111	1880	0	0		STEENSTRUP (1883a, phot.)
$M_0 = 450$ m					
g = 25–33%	1948	400	95	Thinning, clear ice to front	Aer. phot.
	1953	420	100	–	Aer. phot.
	1956	420	100	–	WEIDICK (phot.)

(continued)

(Table 2 continued)

Loc.no., M_0 and g	Year	a_t	$^0/_0$	Comments on glacier front	Source of information
Loc. 112	1850	400	20	Thinning	RINK (1857)
$M_0 = 0$ m	1875	322	16	Expanding	HELLAND (1876)
g = *20*–10 $^0/_0$	1879	230	11	–	STEENSTRUP (1883 a, phot.)
	1880	210?	10?	–	STEENSTRUP (1883 a, phot.)
	1896	0?	0?	Front near or reaching the sea	BARTON (1897)
	1932	c. 1500	75	Recession c. 1915?, thinning 1932	LOEWE (1934)
	1948	c. 2000	100	Thinning	Aer. phot.
	1953	c. 2000	100	–	Aer. phot.
Loc. 113	1850	0	0	Thinning	RINK (1857)
$M_0 = 0$ m	1875	0	0	Calving	HELLAND (1876)
g = 11–12 $^0/_0$	1879	0	0	–	STEENSTRUP (1883 a)
	1880	0	0	Thinning	STEENSTRUP (1883 a)
	1932	c. 900	100		LOEWE (1934)
	1948	c. 900	100	Thinning	Aer. phot.
	1953	c. 900	100	–	Aer. phot.
	1956	c. 900	100	–	WEIDICK (phot.)
Loc. 114	1879	150	25	Thinning, clear ice near to front	STEENSTRUP (1883 a)
$M_0 = 350$ m					
g = *20*–50 $^0/_0$	1896	150	25	–	BARTON (1897)
	1948	600	100	–	Aer. phot.
	1953	600	100	–	Aer. phot.

VIII. UMANAK DISTRICT: LOCAL GLACIATIONS

Loc. 118	1850	0	0	Calving	OLDENDOW (1955)
$M_0 = 0$ m	1931	0	0	–	WEGENER *et al.* (1932)
g = 70 $^0/_0$	1933/34	0?	0?	–	Map
	1942?	0	0	Calving, thinning	Aer. phot.
	1949	0	0	–	Aer. phot.
	1953	0	0	–	Aer. phot.
Loc. 119 I	1850	0	0	Calving	RINK (1857)
$M_0 = 0$ m	1879	0	0	–	STEENSTRUP (1883 a)
g = 14–20 $^0/_0$	1933/34	c. 150	23	Thinning	Map
	1942?	c. 650	100	–	Aer. phot.
	1949	c. 650	100	Thinning, partly dead ice	Aer. phot.
	1953	c. 650	100	–	Aer. phot.

(continued)

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
Loc. 119 IV	1850	0	0	Calving	RINK (1857)
$M_0 = 0$ m	1879	0	0	Advance before 1879, calving	STEENSTRUP (1883a)
g = 10–20%	1933/34	300	25	Advance between 1920 and 1930, thinning	MØLLER (1959)
	1942?	850	71	Thinning, clear ice to front	Aer. phot.
	1949	1000	83	–	Aer. phot.
	1953	1100	92	Thinning, margin partly dead ice	Aer. phot.
	1957	1200	100	Thinning, moraine covered front	MØLLER (1959)
Loc. 119 VI	1850	0	0		RINK (1857)
$M_0 = 0$ m	1879	c. 630	32	Thinning, stagnant	STEENSTRUP (1833a)
g = 10–20%	1933/34	c. 600	30		Map
	1942	c.2000	100	Thinning, moraine covered	Aer. phot.
	1949	c. 2000	100	–	Aer. phot.
	1953	c. 2000	100	–	Aer. phot.

IX. UMANAK DISTRICT: INLAND ICE MARGIN

Loc. 125	1850	0?	0?	Calving	RINK (1857)
$M_0 = 0$ m	1879/80	350	78	–	STEENSTRUP (1883a,b)
g = 4–5%	1891/93	350	78	–	DRYGALSKI (1897)
	1896	350	78	–	BARTON (1897)
	1909	350	78	–	HEIM (1911)
	1929	350	78	–	LOEWE (1935)
	1942?	450	100	–	Aer. phot.
	1948	450	100	–	Aer. phot.
	1949	450	100	–	Aer. phot.
	1953	450	100	–	Aer. phot.
	1959	0	0	–	Aer. phot.
	1964	450	100	–	Aer. phot.
Loc. 126	1850	0	0	Calving	RINK (1857)
$M_0 = 0$ m	1891/93	0	0	–	DRYGALSKI (1897)
g = c. 10%	1929	0	0	–	LOEWE (1935)
	1931/33	500	62	–	Map
	1953	800	100	–	Aer. phot.
	1959	500	62	–	Aer. phot.
	1964	300	38	–	Aer. phot.

(continued)

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
Loc. 127	1850	0	0	Calving	RINK (1857)
$M_0 = 0$ m	1880	300	43	–	STEENSTRUP (1883 a)
g = 13–25 %	1891/93	300	43	–	DRYGALSKI (1897)
	1909	300	43	–	KRABBE (1929)
	1929	300	43	–	LOEWE (1935)
	1948	700	100	–	Aer. phot.
	1953	700	100	–	Aer. phot.
	1959	300	43	–	Aer. phot.
	1964	550	78	–	Aer. phot.
Loc. 128	1850	0	0	Calving	RINK (1857)
$M_0 = 0$ m	1880	900	56	–	STEENSTRUP (1883 a)
g = 6–16 %	1896	800?	50?	–	BARTON (1897)
	1931/33	1500	94	–	Map
	1942?	1600	100	Calving, thinning	Aer. phot.
	1949	1600	100	Calving	Aer. phot.
	1953	1600	100	Calving, partly expanding	Aer. phot.
	1959	1500	94	–	Aer. phot.
	1964	1500	94	Calving	Aer. phot.
Loc. 129	1850	0	0	Calving	RINK (1857)
$M_0 = 0$ m	1880	300	19	–	STEENSTRUP (1883 a)
g = 2–8 %	1891/93	300?	19?	Calving, partly expanding	DRYGALSKI (1897)
	1896	300?	19?	Calving	BARTON (1897)
	1915	1000	62	–	LOEWE (1935)
	1929	1000	62	Calving, thinning	LOEWE (1935)
	1931/34	1200	75	Calving	Map
	1942?	1300	81	–	Aer. phot.
	1948	1400	87	–	Aer. phot.
	1949	1500	94	–	Aer. phot.
	1953	1600	100	–	Aer. phot.
	1959	1200	75	–	Aer. phot.
	1964	600-800	44	–	Aer. phot.
Loc. 130	1850	0	0	Calving	RINK (1857)
(Main glacier)	1880	700	29	–	STEENSTRUP (1883 a)
$M_0 = 0$ m	1917	2000?	83?	–	LOEWE (1935)
g = 4–20 %	1929	2000?	83?	–	LOEWE (1935)
	1931/34	2000	83	–	Map
	1936	2000	83	Calving, thinning	Phot.
	1942?	2100	88	–	Aer. phot.
	1949	2200	92	Calving	Aer. phot.
	1953	2400	100	–	Aer. phot.
	1959	2100	88	–	Aer. phot.
	1964	2100	88	–	Aer. phot.

(continued)

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
Loc. 130	1850	0	0	Calving	RINK (1857)
(northern small	1880	400	33	–	STEENSTRUP (1883a)
glacier)	1933/34	900	75	–	Map
M_0 = 0 m	1942?	1100	92	–	Aer. phot.
g = 11–14%	1949	1100	92	–	Aer. phot.
	1953	1100	92	–	Aer. phot.
	1959	1150	96	–	Aer. phot.
	1964	1200	100	–	Aer. phot.
Loc. 131	1929	600	43	Thinning, clear ice to front	LOEWE (1935)
M_0 = 0 m					
g = 20%	1932	620	44	–	LOEWE (1934)
	1949	1100	79	–	Aer. phot.
	1953	1150	82	–	Aer. phot.
	1959	1400	100	–	Aer. phot.
Loc. 132	1880	0	0	Calving	STEENSTRUP (1883a)
M_0 = 0 m	1933/34	1200	39	–	Map
g = 5%	1949	2100?	68?	–	Aer. phot.
	1953	2300	74	Calving, northern part on land	Aer. phot.
	1959	2900	94	Calving, northern part in contact with island	Aer. phot.
	1964	3100	100	Calving	Aer. phot.
Loc. 133	1875	0	0	Calving	HELLAND (1876)
M_0 = 0 m	1933/34	1300	100	–	Map
g = 3%	1942?	1100	85	–	Aer. phot.
	1949	1100	85	–	Aer. phot.
	1953	1100	85	–	Aer. phot.
	1959	0–800	31	Calving, southern part advancing	Aer. phot.
	1964	0	0	–	Aer. phot.
Loc. 134	1893	0	0	Calving	DRYGALSKI (1897)
M_0 = 0 m	1932	3000	57	–	SORGE (1932)
g = 5%	1949	4470	85	–	Aer. phot.
	1950	4645	89	–	RANSLEY (1952)
	1953	4645	89	–	Aer. phot.
	1959	5245	100	–	Aer. phot.
	1964	5245	100	–	Aer. phot.

X. UPERNAVIK DISTRICT: LOCAL GLACIATIONS

Loc. 137B	1879/80	0	0		STEENSTRUP (1883b)
M_0 = 100 m	1934/37	1600	55		Map
g = 14–17%	1942?	2500	86	Thinning	Aer. phot.
	1949	2800	97	–	Aer. phot.
	1953	2900	100	–	Aer. phot.

(continued)

Table 2 (continued)

Loc.no., M_0 and g	Year	a_t	%	Comments on glacier front	Source of information
XI. UPERNAVIK DISTRICT: INLAND ICE MARGIN					
Loc. 142	1849	2100	10	Calving	Rink (1857)
$M_0 = 0$ m	1886	4000	19	–	Ryder (1889)
g = 4%	1931	7000	33	–	Carlson (1941)
	1937	9000	43	–	Map
	1942?	14000	67	–	Aer. phot.
	1946	18600	88	–	Aer. phot.
	1949	21000	100	–	Aer. phot.
	1953	21000	100	–	Aer. phot.
Loc. 147	1887	0	0	Calving	Ryder (1889)
$M_0 = 0$ m	1896	200?	20?	Calving, margin thinning	Tarr (1897)
g = 10%					
	1931	200?	20?	Calving	Carlson (1941)
	1942?	1000	100	–	Aer. phot.
	1949	1000	100	–	Aer. phot.
	1953	1000	100	–	Aer. phot.

15.3. Height of the trim line zone (Δh_H) above the present glacier surface (h)

All figures are given in metres.

Table 3.

Loc.no.	h =	0	100	200	300	400	500	600	700	800	900	1000	1100	1200
Local glaciations, Julianehåb district														
1a	Δh_H=	300	234	167	100	86	73	60	47	33	20			
1b	–	200	184	167	150	126	103	80	64	47	30	24	17	10
3	–	60	50	40	30	30	30	20	18	15	10			
4b	–		100	80	60	46	33	20	18	15	13	10		
6	–	50	50	45	40	43	45	48	50	25	20	17	13	10
Δh_H mean =		152	124	100	76	66	57	46	39	27	19	17	15	10
Standard deviation =		104	74	57	44	35	28	23	18	12	7	6	–	–
Local glaciations, Sukkertoppen district														
40	Δh_H=		40	35	30	25	20	20			20			
41	–	50	50	44	37	31	25	24	23	21	20	15		
43	–		150	100	50	70	50	50	30	40	50	20	10	15
46a	–		50	50	50	45	50	50	30	10				
47	–	100	50	25	10									
48	–	200	200	150	100	100	20							
49	–	250	200	50	40									
50	–	100	100	50	50	50	50	25						
51	–	100	100	50	50	40	50	30						
59a	–	50	50	50	50	50	40	30	10					
63a	–	100	80	60	40	33	25	18	10	5	0	0	0	
Δh_H mean =		119	97	60	46	49	37	31	21	19	23	12	5	–
Standard deviation =		66	57	34	21	22	13	12	9	13	18	9	–	–
Local glaciations, Blæsedalen, Disko island														
85 B	Δh_H=					100	50	100	75					
85 A	–						100	75	60					
86–1	–							75						
86–2	–						50	50						
86–3	–					100	50	50	50	30				
86–4	–						50	50						
Δh_H mean =						100	60	67	62	–				
Standard deviation =						–	20	19	10	–				

(continued)

(Table 3 continued)

Loc.no.	h =	0	100	200	300	400	500	600	700	800	900	1000	1100	1200
Inland Ice														
8	Δh_H=	0	0	0	0	0	0	0	0	0	0			
9 V	–	250	200	150	100	74	47	20						
9 Ø	–	250	200	150	100	74	47	20	10	0				
10	–	150	100	100	100	80	60							
12	–	100	70	70	70	70	70							
14 B	–	130		25		70								
14 A	–	100	94	88	82	76	70							
15	–	150	100	80	100	93	85	60	20		45			
18	–	40	37	33	30	27	23	20						
20	–	200	169	138	106	74	42	10						
23–N	–					160	70							
23–S	–							100	60	20				
26–27	–	300	300	190	150	140	96	77	83	83	68	42	42	60
28	–	0	0	0	0	0	10	10	10	10				
30	–							10	10	10	10			
31	–	10	8	6	5	5	5	0						
32	–	20	17	13	10									
33	–		50	20	20	20	20	20	20					
34 A	–	20	18	16	14	12	10							
34 B	–	50	42	34	26	18	10							
35	–	10	10	10	10	10	10							
36	–		15	10	10	5	5							
37	–	250	150	50	30	10								
39	–	10	0	0	0	0								
66	–	20	20	20										
67	–	0	0	0	0	0								
68	–	0	0	0	0	0								
69	–	0	0	0	0	0	0							
70	–	250	250	250	250	200	100							
71	–	100	50	50	50									
72	–		100	100	50									
73	–	100	75	50	50	50								
74 A	–	100	75	50	50	40	30	20						
74 B	–	50	20	20	20									
75	–	50	20	20	20	20								
76	–	50	40	30	30	30	30	30						
Δh_H mean =		92	70	54	48	47	38	28	27	21	31	–	–	
Standard deviation =		92	78	61	55	51	32	29	27	28	27	–	–	

15.4. Percentage of glacier lobes forming dead ice

(see pp. 67–68).

Table 4.

Inclination of glacier	Glacier lobes forming dead ice		Glacier lobes without dead ice formation		Total	% forming dead ice
	Number	Localities	Number	Localities		
2.0°–2.9° (3.5–5.1 %)	3	6, 36, 83	2	31, 39	5	60
3.0°–6.8° (5.2–12.0 %)	6	38, 41, 53 IV–S, 86–4, 106, 107	11	35, 42, 64, 65 A, 71 B, 72, 73, 79 C, 85 B, 109, 113	17	35
6.9°–10.3° (12.1–18.2 %)	3	119 I, 119 IV, 119 VI	10	10(V), 11, 14 A, 43, 53 IV–N, 84, 86–1, 86–2, 86–3, 137 B	13	23
10.4°–15.5° (18.3–27.8 %)	5	33 B_3, 40, 78 A, 78 B–C, 79 B	15	1 a, 2, 7, 51 S, 53 I–N, 55–1, 61 I, 62, 63, 79 A, 85 A, 95, 103, 108, 131	20	25
15.6°–21.8° (27.9–40.0 %)	1	96	12	47 B, 47 E, 48 C, 56–1, 57 VI–N, 58, 61 II, 61 IV–A, 61 V, 61 VI, 111, 114	13	8
>21.9° (>40.1 %)	0		5	25, 44, 48 A, 54–1, 61 III	5	0
Total	18		55		73	

15.5. Radiocarbon dated samples

The dates used in the compilation of fig. 2 include those of the following shell samples collected by the author:

Table 5a.

No.	Locality & occurrence	Species	Altitude	Age before 1950
K- 992	Claushavn, Disko Bugt. cf. p. 118.	*Mya truncata*	40±5 m	7110±140
K- 993	Eqaluit, Disko Bugt. cf. p. 114.	*Mya truncata*	52±5 m	7650±140
K- 994	Sarqaq, Disko Bugt. Uppermost part of marine clay	*Saxicava arctica*	63±7 m	8940±170
K-1033	Akugdleq, Holsteinsborg district. Upper marine clay	*Mya truncata*	40±5 m	6860±150
K-1034	Holsteinsborg town. Upper marine clay	*Mya truncata*	48±3 m	8840±170
K-1035	Holsteinsborg town. Stony beach ridge	*Mytilus edulis*	20±3 m	4590±110
K-1036	Kapisigdlit, Godthåb district. Top of clayey beach ridge	*Mya truncata*	40±5 m	7560±150
K-1037	Avatdleq, Holsteinsborg district. Upper marine shell gravel under beach ridges (cf. p. 137)	*Mya truncata*	95±5 m	8250±130

In addition, dates for limnic and terrestrial deposits were obtained from the following sources:

K-144, Igdlorssuit, Disko Bugt (Larsen and Meldgaard 1958, Tauber 1960a).
K-518, Sarqaq, Disko Bugt (Larsen and Meldgaard 1958, Tauber 1960b).
K-588, Itivnera kitdleq, Godthåb district (Larsen and Rosing *in* Tauber 1962a).
K-966, Godthåb town, Godthåb district (Kelly 1964).
K-987, Claushavn, Disko Bugt (Kelly 1964).
K-988, Nerutussoq, Frederikshåb district (Kelly 1964).

All samples have been dated in the Carbon-14 Dating Laboratory at the National Museum, Copenhagen, by H. Tauber.

No attempts were made to investigate thoroughly the fauna of the samples taken for dating, but the following specimens were noted. The samples are listed in the order of their height above sea level. G. Høpner Petersen, E. Nordmann, A. Rosenkrantz and E. Steenstrup have kindly helped in the identification of the species.

(continued)

(Table 5 continued)

Table 5b.

Species	Sample no. and its altitude above present sea level K–1035 (20 m)	K–992 (40 m)	K–1033 (40 m)	K–1036 (40 m)	K–1034 (48 m)	K–993 (52 m)	K–994 (63 m)	K–1037 (95 m)
Lamellibranchiata								
Nucula sulcata	.	.	.	.	.	.	×	.
Portlandia arctica	.	.	.	.	.	×	.	.
Pecten islandicus	×	.	×	×	×	.	.	×
Mytilus edulis	×	.	.	×	.	.	.	×
Cardium ciliatum	.	.	×	× ?	.	.	.	.
Serripes groenlandicum	.	×	×	×	.	.	.	.
Macoma calcarea	.	×	×	×	×	×	× ?	.
Saxicava artica	×	×	×	×	×	×	×	×
Mya truncata	.	×	×	×	×	×	×	×
Cirrepedia								
Balanus balanus	×	.	.	×	× ?	.	×	×
Balanus crenatus	×	.	.	.	× ?	.	.	.
Echinoidea								
Strongylocentrotus droebachiensis	.	.	.	×	.	.	.	.
Gastropoda								
Puncturella noachina	×	.	.	.	×	.	.	×
Natica clausa	.	.	×	.	.	.	.	.
Buccinum tenue	× ?	.	.	.	× ?	.	.	.
Variae								
Laminaria sp.	.	.	.	×	.	.	.	.

16. INDEX OF PLACE NAMES

Abbreviations used: Jhb.: Julianehåb district
Frhb.: Frederikshåb –
Ghb.: Godthåb –
Skt.: Sukkertoppen –
Hbg.: Holsteinsborg –
Egm.: Egedesminde –

Færdig fra trykkeriet den 13. maj 1968.

PLATES

Plate 1

Index map of localities with glaciers of which historical fluctuations have been investigated. Numbers in brackets indicate an area of overlap between the maps to the left and in the centre.

Plate 2

Historical fluctuations of the best documented glacier lobes. Numbers refer to the localities shown on the index map, plate 1. The sources of the date are given in section 15.2.

The abscissae are in years A.D. The ordinates are the longitudinal distance in metres (a_t), one scale unit being equivalent to 1000 m. See fig. 8 and text p. 29 for definition of a_t.

Instead of Blæsedal, read Blæsedalen and instead of Tâterat sermia, Tâterat sermiat.

Plate 3

Prehistoric ice margin deposits between Godthåb and Umanak districts. Map based on the Geodetic Institute's 1:250,000 map sheets of West Greenland. Names in inverted commas are not authorized. Other incorrect place names should be read as follows:

Nukapiaq (Sukkertoppen district), read Nukagpiaq
Tasiunaq (–), read Tasiussaq
Kangâmiut kangerdluarssat (–), read Kangâmiut kangerdluarssuat
Sermitsiap (–), read Sermítsiaq
Tâterat sermit (–), read Tâterat sermiat
Sigssarigssut (–), read Sigssarigsut
Mitdlugssalik (Holsteinsborg district), read Mitdlutigssalik
Majoriarssuatsiap qàqâ (Nûgssuaq peninsula), read Majoriarssuatsiaup qáqâ
Sermiarssuit sermikarsât (–), read Sermiarssuit sermikavsât
Qaerssuarssuk (–), read Qaerssuarssuk kitdleq
Agssakait sermia (–), read Agssakait sermiat

In legend; instead of presumbly, read presumably,

GRØNLANDS GEOLOGISKE UNDERSØGELSE
THE GEOLOGICAL SURVEY OF GREENLAND
MEDD. OM GRØNL. BD. 165 NR. 3 (ANKER WEIDICK)

GRØNLANDS GEOLOGISKE UNDERSØGELSE
THE GEOLOGICAL SURVEY OF GREENLAND


Agpat
SÂTUT
UMANAK
Storøen
NÛGSSUAQ
"Qarajaqs Isfjord"
Drygalskis Halvø
Hareøen
VAIGAT
QUTDLIGSSAT
SARQAQ
DISKO
GODHAVN
DISKO BUGT
JAKOBSHAVN
Jakobshavns Isfjord
CLAUSHAVN
CHRISTIANSHÅB
EGEDESMINDE
Arfersiorfik
AGTO
Nordre Strømfjord
KANGILINAQ
NIAQUA
QÂQAPALÂT
HOLSTEINSBORG
Søndre Strømfjord
TROLDFJELDENE
SUKKERTOPPEN ISKAPPE
Sermersût
SUKKERTOPPEN
FINNEFJELD
QÔRNOQ
Godthåbsfjord
KAPISIGDLIT
Sadlen
GODTHÅB
HJORTETAKKEN
NUNATARSSUK
TUNGMERALIK
Fiskenæsset
QAQAT AKULERIT
NUKAGPIARSSUAQ
UMANARSSUAQ
QAQARSSUAQ
FREDERIKSHÅBS ISBLINK
10 0 10 20 30 40 50
Ice margin deposits of pre-historic age
Marine deposits of wider extent. Partially or fully covered by glacio-fluviatile deposits
Eskers
Spill ways
Dead ice terrain
Present glaciers
Area glaciated during the fjord stage
Area, presumbly glaciated during the fjord stage