Geomorphological observations at Kangerdlugssuaq, East Greenland

Charles Kent Brooks

1979

Geomorphological observations at Kangerdlugssuaq, East Greenland

Charles Kent Brooks

Table of contents

Geomorphological observations at Kangerdlugssuaq, East Greenland

CHARLES KENT BROOKS
Institute of Petrology, University of Copenhagen

The Kangerdlugssuaq area is mainly comprised of two contrasting rock groups: on
the one hand the easily-eroded lavas and sediments of late Mesozoic to early Tertiary
age and on the other the highly resistant Precambrian gneisses. Intermediate between
these two types in terms of behaviour with respect to erosion are the Tertiary plutonic
complexes and the basaltic areas along the coast which have been intruded by intense
dyke swarms.

In the late Mesozoic the area was a peneplain, and low relief apparently persisted
throughout the volcanic episode as there is good evidence that the lava plateau
subsided during its formation. During this period ocean-floor spreading gave rise to
the embryonic Danmark Stræde. Shortly after the volcanic episode the
Kangerdlugssuaq area became the centre of a massive domal upwarping which has
been a dominant feature of the land-forms up to the present day. The original surface
of the dome has been reconstructed on the basis of topographic and geological
evidence to show that it was elliptical in form with a major axis of at least 300 km in
length and a height above present sea-level of about 6.5 km. However, subsequent
isostatic effects are not considered in deriving these figures. The updoming is esti-
mated to have occurred about 50 m.y. ago.

Several kilometres thickness of sediments and lavas were eroded off this dome at
an early stage exposing the gneissic core, which still stands in alpine peaks up to about
2.7 km altitude in the central part, and dumping ca. 50000 km³ of sediment on the
continental shelf. The erosion was effected by a radial, consequent drainage system,
relicts of which can still be found. Kangerdlugssuaq itself may owe its origin to a
tectonic line of weakness formed in response to doming, but there are also good
arguments for its being purely erosional. The erosion of the dome was probably
fluviatile but all trace of this stage has been obliterated by the subsequent glaciation.

In the period between the Eocene and the early Miocene, possibly around 35 m.y.
ago, the entire area underwent epeirogenic uplift raising the undeformed parts of the
original lava plateau to around 2.5 km above sea-level. At present this plateau is
undergoing dissection from the seaward side, but considerable areas are still pre-
served under thin, horizontal ice-caps.

A brief description of the various types of glaciers, an impermanent, ice-dammed
lake and the areas of ice-free land is given. In the Pleistocene, the Kangerdlugssuaq
glacier was considerably thicker than at the present time and extended far out over
the shelf, excavating a deep channel here. Finally some observations on the coastlines
are presented.

*C. K. Brooks, Institute of Petrology, Øster Voldgade 10,
DK-1350 Copenhagen K.*

This paper is built to a large extent on the work of Professor L. R. Wager. At the time of his sudden death in 1965 he was actively working on a paper on this topic, and an unfinished manuscript, the essence of which had already been published (Wager 1937), together with a heap of unsorted notes, photographs etc. were given to me by Professor W. A. Deer in 1977, who himself was responsible for much of the fieldwork. I had discussed this work with Professor Wager on a number of occasions, in connection with the Oxford University Expedition to East Greenland in 1965 of which I was one of the leaders and during which we visited the plateaus just south of Scoresby Sund. Wager's observations were based on work carried out on three expeditions: the British Arctic Air Route Expedition in 1930–1931 (Watkins 1932), during which the broad outlines of the geology of the coastal area of East Greenland between Angmagssalik and Kangerdlugssuaq were determined, the Scoresby Sound Committee's 2nd East Greenland Expedition in 1933 (Mikkelsen 1933), when the earlier work was extended to the coast north-east of Kangerdlugssuaq, and the British East Greenland Expedition of 1935–1936 (Courtauld 1936, Wager 1937). It was on this last expedition that journeys were made through the inland region and an impression of the interior of the country first gained.

I have had the good fortune to visit East Greenland frequently during the last 10 years and have been able to supplement Wager's observations. None of the journeys hitherto made in the area have been specifically designed for physiographical observations and air photographic cover of much of the area is inadequate, but sufficient information has been gathered to form a basis on which future work in the area can build.

Geology

The basement, which is extensively exposed between Angmagssalik and Kangerdlugssuaq and as inliers as far north as Nansen Fjord, consists of Archaean and Proterozoic rocks, overwhelmingly of acid gneisses (Bridgwater et al. 1978). It was roughly peneplaned during the Mesozoic and transgressed at the end of the Cretaceous with the deposition of largely estuarine- or shallow water-type sediments, now represented by the Kangerdlugssuaq Group (Wager 1947). At this time the basement surface was deeply weathered and uneven (islands apparently stood out of the depositional basin), but is described as having unaccentuated relief (Wager 1947: 16). In Gåseland, in the Scoresby Sund area, basalt-filled valleys up to 400 m deep may be seen dissecting the peneplain (W. S. Watt 1969, Watt & Watt 1971), but which is otherwise gently undulating (Watt & Watt 1973).

Basaltic volcanism commenced in the upper

Paleocene in the Kangerdlugssuaq area and rapidly built up a thick lava pile, possibly as much as 9 km thick in places (Soper et al. 1976). During this period no marked relief appears to have developed, as no significant unconformities have been found. The build-up of the plateau kept pace with sinking of the area as it is capped by marine sediments (see below).

Major intrusive activity in the Kangerdlugssuaq area occurred in two periods. The first of these was during or very shortly after the extrusive episode and is represented by the Skaergaard and other gabbroic intrusions. About 5 m.y. subsequently, many syenitic intrusions were emplaced in the area (Brooks & Gleadow, 1977) and this seems to have marked the end of large-scale magmatism here although intrusions in the Kialêq (earlier Kialineq) area some 150 km to the south have an age of about 35 m.y. (Brown et al., 1977; Brooks & Gleadow, unpublished).

This paper deals with the development of the area subsequent to the volcanic episode. The land at the close of this episode was one of low relief, lying approximately at sea level as shown by the presence of shallow water marine sediments (Ravn 1933, Wager 1935, Birkenmajer 1972, Soper & Costa 1976) overlying the basalts at Kap Dalton and Kap Brewster and hyaloclastites and pillow lavas in the upper part of the succession around Scoresby Sound (Fawcett et al. 1966).

The land surface prior to the present erosional cycle

Wager (1934, 1947) described in some detail the Tertiary tectonic movements which deformed the original monotonous landscape at the end of the volcanic episode. Of these, two are dominant: the coastal flexure and the inland dome. Their form can be deduced from a combination of geological and geomorphological evidence.

The ice-covered plateaus

These represent remnants of the once continuous landscape of low relief which existed at the close of volcanism and are found in the inland area between Kangerdlugssuaq and Scoresby Sund. They were first revealed by air photography carried out by the Seventh Thule Expedition and by Lauge Koch in 1933 and by subsequent air photography, although air photographs taken by the British Arctic Air Route Expedition in 1930 showed what appeared to be the distant edge of an elevated plateau in the area of Watkins Bjerge. Similarly Wager (1933) described an ice cap covering what seemed to be a little-modified elevated plateau cut from
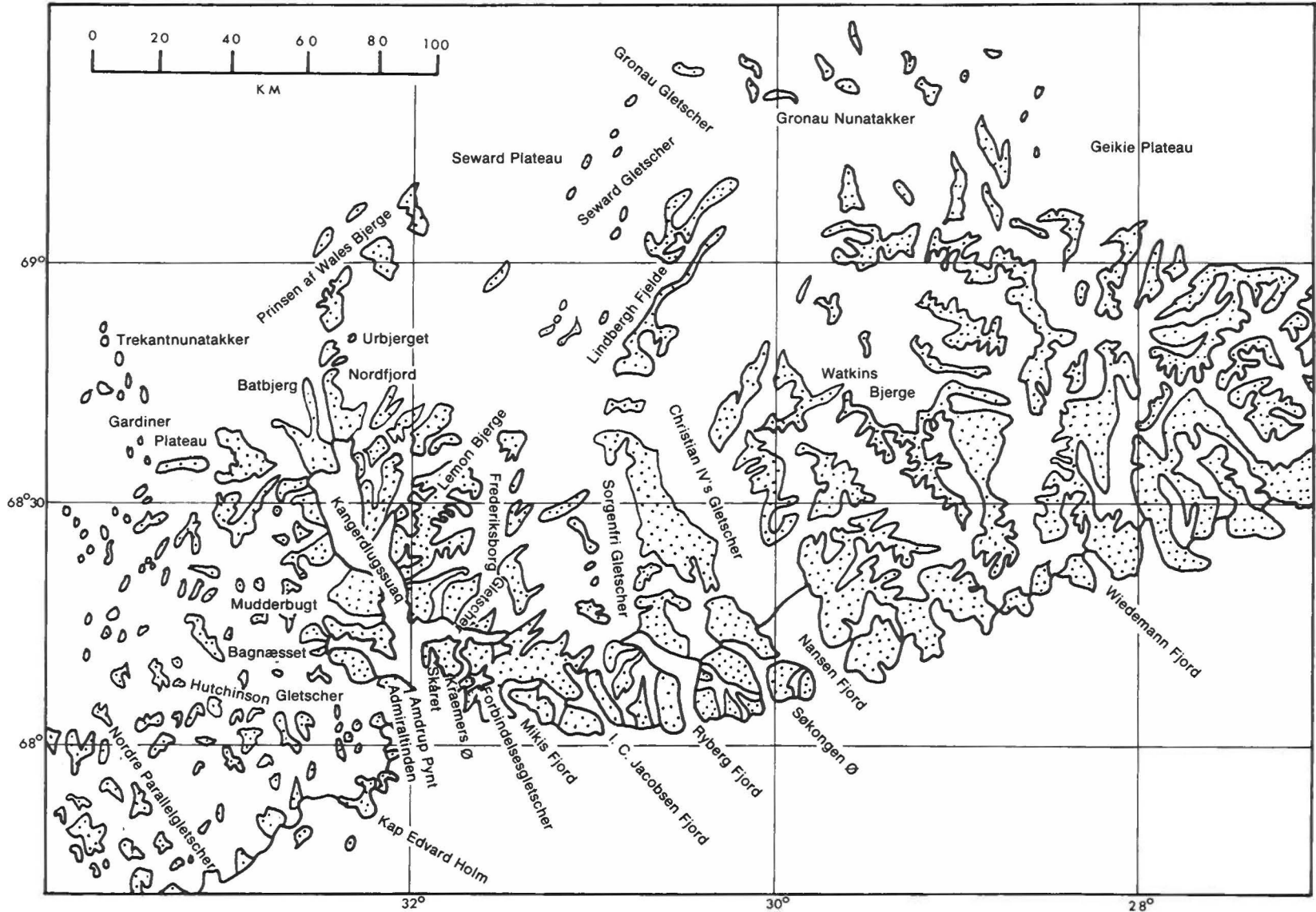
Fig. 1. General map of the area showing the most important localities discussed in the text.

Fig. 2. Oblique air photograph showing the ice covered plateau at around 69°30', i.e. well outside the area affected by doming. Note the close parallelism between the altitude of the flows and the flat tops of the nunataks which are believed to be very close to the original post-basaltic land surface. The plateau is undergoing dissection which is believed to have been initiated by uplift in the Oligocene. Note the characteristic 45° slope profile in these unmetamorphosed basalts and compare with that in the area injected by the dyke swarm to the south (Fig. 12). (Reproduced with permission of the Geodetic Institute, Copenhagen, A. 128/78. COPYRIGHT).

the basement gneisses in the Mont Forel region of the Angmagssalik district. The ice-covered plateaus were observed from the summit of Gunnbjørn Fjeld in 1935, and a small one, Seward Plateau, was visited during the spring journey of 1936 (Wager 1937), weather conditions having prevented the projected journey to traverse the Geikie Plateau. The Oxford Expedition of 1965 visited remnants of this ice-covered plateau in the area just south of Scoresby Sund (Fawcett et al. 1966) where typical features are very well seen (Fig. 2).

Seward Plateau to the north of the head of Kangerdlugssuaq lies between Prinsen af Wales Bjerge and Gronau Nunatakker. It is covered by a small ice cap, separated by a slight hollow from the main inland ice and composed of almost level uncrevassed ice spilling over on the north-west and west to the Kangerdlugssuaq basin, on the south to the upper Frederiksborg Gletscher, and on the east and south-east to

Gronau Gletscher and Seward Gletscher. The centre of the ice dome capping the plateau rises but slightly and the rock outcrops round the edge appear at almost constant level on the three sides which were visited on the inland journey in 1936 by W. A. Deer and E. C. Fountaine and consist of almost flat-lying basalts. Debris does not appear from under the ice cap. Available evidence suggests Geikie Plateau to be essentially similar, although it has not yet been visited. The Oxford party in the area south of Scoresby Sund confirmed these conclusions, although they only visited outlying remnants of the main plateau, whose aspect in this area is shown in Fig. 2.

These observations are interpreted as showing that these ice-covered plateaus consist of a relatively thin ice-sheet resting on an almost horizontal surface, which, because it appears to be conformable with the underlying basalt flows, is likely to be close to the original sur-

Fig. 3. Cirques and "matterhorns" developed in the basalts where the original surface of the plateau has been destroyed within the area affected by doming. Vertical air photograph just west of the Lindbergh Fjelde. (Reproduced with permission of the Geodetic Institute, Copenhagen, A. 128/78. COPYRIGHT).

face as left at the cessation of volcanic activity in the Lower Tertiary. The paucity of zeolites in the basalts of these areas supports this interpretation. Evidence to be introduced later suggests that the whole of this area was elevated not before the early Miocene although Wager (1947: 47) believed it to have been contemporaneous with the formation of the coastal dyke swarm in the earliest Eocene. It would be difficult to find an agent which could produce such a level and uniform surface at an altitude of about 2500 m over a very wide area, and, if it could be developed, that it should be parallel to the dip of the lavas over an area extending from the head of Kangerdlugssuaq to the hinterland of Kap Brewster (ca. 400 km). It seems that the hypothesis that this surface represents remnants of the original surface formed by

constructional volcanic activity is reasonable and will be adopted here. The ice on these plateaus which has little movement and carries no englacial debris, appears to protect them from the action of normal weathering processes and has itself little denuding power, otherwise the plateaus would not have been preserved to the extent which we now see them.

As far as can be determined, the ice-covered plateaus appear to dip outwards from Kangerdlugssuaq at a low angle, ca. 2° (Fig. 3). This is determined by generalized measurements of dip taken from a distance by parties travelling along the glaciers. Such a dip is similar to that of the flanks of Hawaiian shield volcanoes (Wentworth & Macdonald 1953) and it might be questioned if this is a remnant of a similar type of volcano. However, the

uplift of the basement which has taken place in the central area (Fig. 4) indicates that this is merely an outlying effect of the domal uplift to be described below.

The inland dome

Already mentioned briefly by Wager (1947: 38) and by Brooks (1973), who tried to link its formation to underlying movements in the mantle, the dome is a region of anticlinal uplift centered on about the middle of Kangerdlugssuaq. In this central area the basalts and sediments have been completely stripped away exposing the gneiss core which still rises to its original level in some of the mountain tops. In the area to the northeast of Kangerdlugssuaq the structure is seen from even a cursory glance at a geological map and parts of the domal flanks are still to be seen. Thus, the summit levels (gipfelflur) of the Lindbergh Fjelde and Watkins Bjerge form a plane dipping gently towards north-east and apparently continuous with the ice-covered plateaus already described. This surface is, however, very much dissected. Further to the south and west the basalt peaks on proceeding inwards towards the domal centre do not approximate to any surface, as erosion has gone even further. The basalts in this area are mere remnants carved into "matterhorns" by advanced cirque formation (Fig. 3).

In the vast area to the south of Kangerdlugssuaq no journeys have been made and the area is virtually unknown except for the immediate areas along the west of the fjord which are largely underlain by Tertiary plutonic complexes. The presence of Tertiary basalts to the north-east of Kangerdlugssuaq facilitates the reconstruction of the original land surface as will be described below. However, these rocks are apparently largely absent to the south and this possibility does not exist; I have therefore refrained from considering this area, although it is clear that regional uplift has been just as great here. The area was formerly also, at least in part, covered with Tertiary basalt as shown by remnants on Kap Edvard Holm and Kap Gustav Holm (Wager 1934) and as xenoliths in the plutonic rocks at Kialêq and elsewhere. Basalt has also been confirmed inland from Nûgâlik (Bridgwater et al. 1978) but its extent is at present unknown. The tops of the basement mountains reach a culmination at around 2500 m altitude about 40 km inland (near the heads of Nordre Parallel Gletscher and subsequently decline to 2000 to 2200 m towards the inland ice. At Gardiner Plateau the basement/Tertiary contact is at about 1200 m (Frisch & Keusen 1977), but this may be down-faulted. Therefore, unless this pattern is strongly affected by the isostatic effects of the inland ice, it appears probable that the anticlinal structure so clearly seen northeast of Kangerdlugssuaq continues towards the south.

Reconstruction of the dome

Fig. 4 summarizes the observations used in the construction of the original land forms. It is assumed that the ice-covered plateaus lying generally at an altitude of around 2500 m represent an original datum, while dips of lavas recorded on Lindbergh Fjelde and Watkins Bjerge represent the original dips on the flanks of the dome as these lavas were originally parallel to the early Tertiary land surface. It is further assumed that the tops of these mountains approximate to the original surface for the reasons already stated thus allowing the 2500 m contour on the ice-covered plateaus to be extended in this region with 3000 and 4000 m contours.

In the area around Kangerdlugssuaq different methods have to be employed due to the total obliteration of the initial surface of the dome. These are less reliable, as they make use of estimates of the original thickness of sediments and lavas, which, because they have subsequently been partly removed, are based on extrapolation. Indeed, even in the areas where large thicknesses are to be found, considerable differences of opinion appear to exist (Brooks et al. 1976, Soper et al. 1976). In addition, isostatic uplift in response to erosion and loading by the inland ice are

2·4 Height of gipfelflur in basement mountains (km).
∖₅ General dip of Tertiary lavas and sediments.
② Estimated thickness of Tertiary sequence (km).
⋌² Precambrian/Tertiary unconformity, attitude in km
     (gneiss shaded).

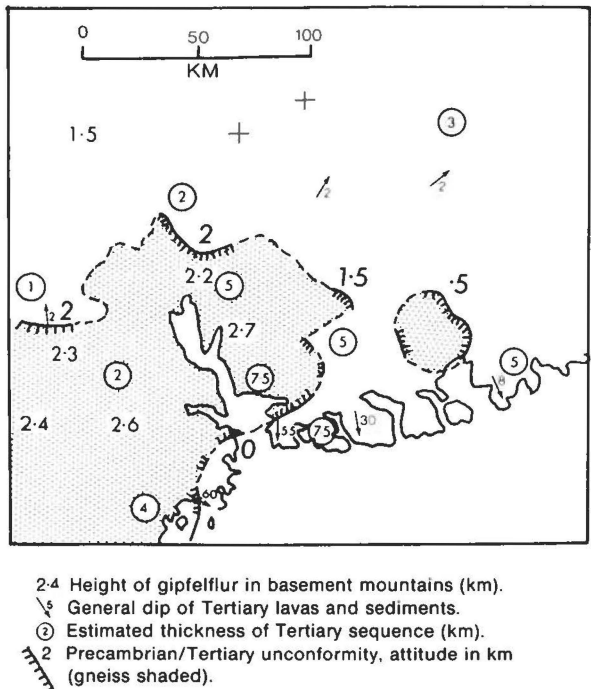Fig. 4. Summary of data used in the construction of Fig. 5. Individual observations, shown here in a highly generalized form, are taken from topographic maps at a scale of 1:250,000 by the Geodetic Institute, Copenhagen, and the U.S.A.F. Operational Navigation Chart, St. Louis, Missouri, at a scale of 1:1,000,000, and the published works of Wager (1934, 1947) and unpublished notes of L. R. Wager and the author.
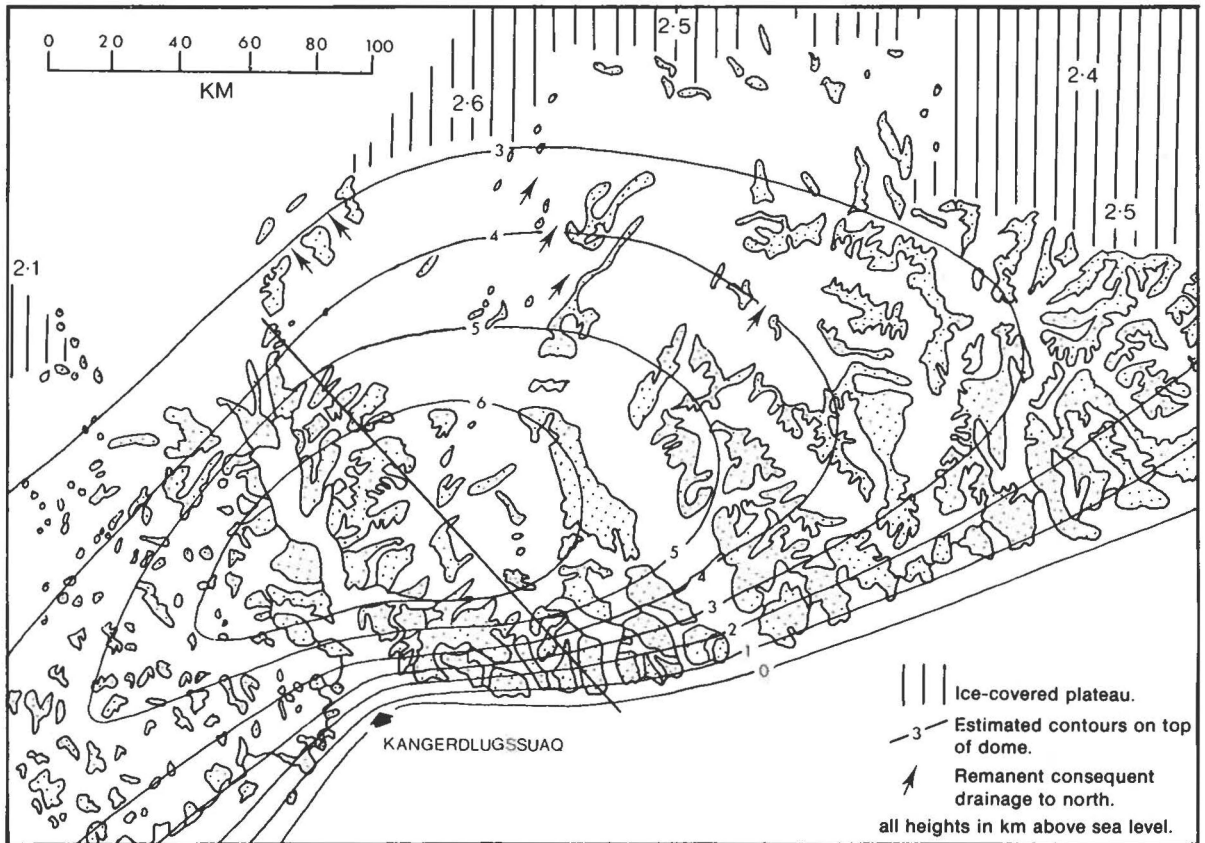
Fig. 5. Reconstruction of the Kangerdlugssuaq dome based on the data summarized in Fig. 4. The effects of faulting have been totally neglected and the form of the dome is uncertain to the southwest of Kangerdlugssuaq where basalts are absent. The diagonal line shows the position of the section in Fig. 6.

likely to have modified present day dips and altitudes and the net effect of these is likely to lead to an overestimation of the height of the dome. Nevertheless, the methods and data used are believed to be adequate to demonstrate the general pre-erosional form of the landscape.

In the central area it is assumed that the summits of the basement mountains (gipfelflur) approximate to the original peneplain on which the Kangerdlugssuaq Group was deposited. These reach a culmination of 2700 m in the Lemon Bjerge Group which thus lies approximately at the centre of the uplift. This assumption does not appear to be far from the truth, and in any case gives a minimum estimate for the extent of up-warping of the basement. The thickness of lavas and sediments originally overlying this unconformity was variable. It is at a minimum in the inland areas where almost the entire column is exposed at the edges of the ice-covered plateau and varies from about 1 km at Tre-kantnunatakker to 4 km in the Watkins Bjerge scarp, and a maximum of perhaps as much as 9 km in the coastal areas (Soper et al. 1976), although 7500 m has

been assumed to allow for possible faulting (Nielsen 1975). The intermediate thicknesses as extrapolated are shown in Fig. 4. Using these data, i.e. height of base-ment peaks, estimated thickness of Tertiary cover and the altitude of the Precambrian/Tertiary unconformity, the contours have been completed as shown in Fig. 5. The effects of the coastal flexure, as discussed below, also appear in this figure which is, of necessity, highly schematic. Fig. 6 shows a section through the structure from Urbjerget to Kap I. C. Jacobsen.

Two additional complicating factors should briefly be noted. These are firstly that the Mesozoic peneplain was not horizontal, as shown by the fact that consider-able thicknesses of sediment and basalt were deposited in the coastal regions while the inland areas were still dry land (Wager 1947). A similar situation obtained in the Scoresby Sund area where a slope af about 2.5° is recorded for the pre-basaltic land surface (e.g. Watt 1970). Secondly, because the basalts are wedging out inland their dip on the inland flank of the dome will be a little less than the dip of the reconstructed surface. Fi-nally it should perhaps be noted that although most of
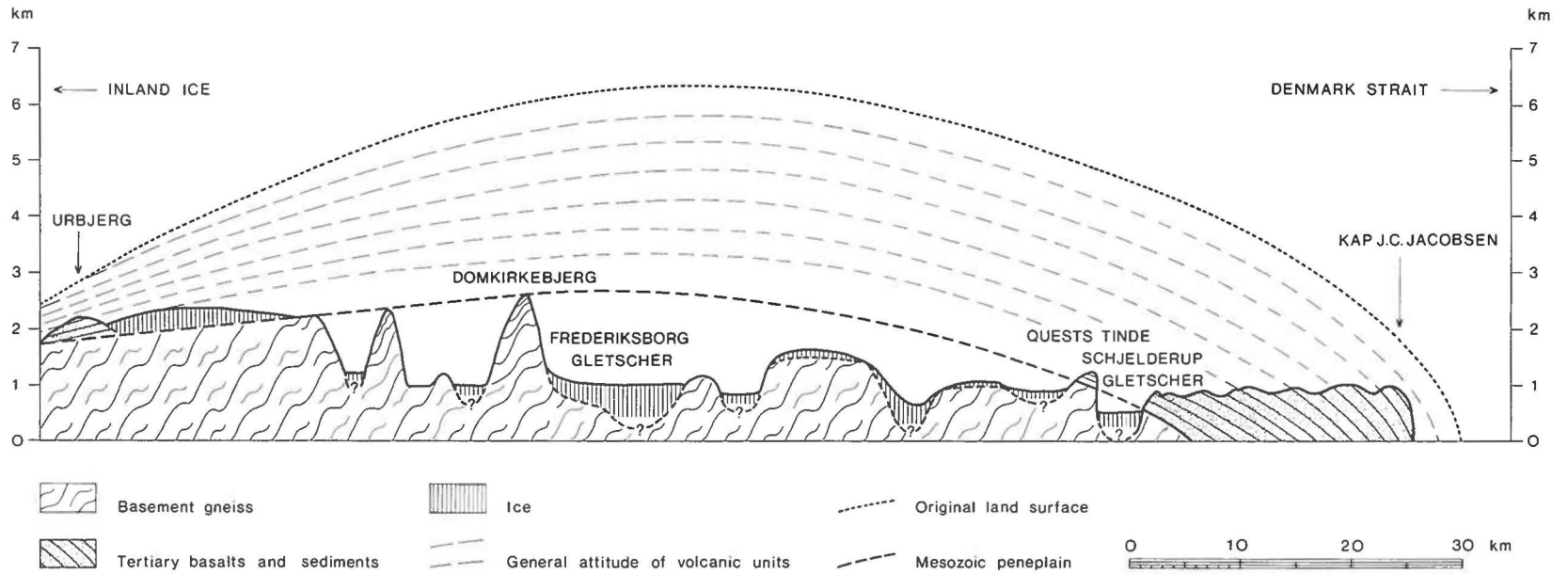
Fig. 6. Section of the dome from Urbjerget to Kap I. C. Jacobsen (see Fig. 5). The exhumed Mesozoic peneplain in the area of Nordfjord Plateau continuing into the gipfelflur in the area between Courtauld Gletscher and Frederiksborg Gletscher is apparent. The low mountains between Frederiksborg Gletscher and Schelderup Gletscher are believed to have been reduced in height and rounded at a time when they were overwhelmed by thicker ice cover.

the vertical movement is to be ascribed to tectonic forces, simple isostatic adjustment to erosion or ice loading must also be considered. Although the amount of uplift at the centre of the Kangerdlugssuaq dome is in excess of 6000 m relative to the present sea level, this is not to say that this height was ever attained in practice. It is likely that a significant part of this uplift is a continuing process in response to unloading as considerable thicknesses of rock were stripped away. This is the well-known situation of the mountains becoming higher as erosion advances due to isostatic uplift of the peaks, as the average land surface is lowered (see e.g. Gilluly et al. 1968: 514).

Origin of the domal uplift

Because both the dome and the intrusion of syenitic magmas were centred on the middle parts of Kangerdlugssuaq it is natural to suggest that the uplift was caused by a disturbance in the mantle which led to the development of these magmas. Similar domal uplifts are known from other areas, notably Africa, both within the rift-valley system and in isolated areas such as Tibesti (e.g. Gass 1972). These uplifts are often associated with intense volcanism, and geophysical work shows the development of an anomalously light upper mantle in these regions (Baker & Wollenberg 1971). It was this comparison which led Brooks (1973) to suggest that doming was related not only to the magmatic activity but possibly also to the initiation of ocean-floor spreading in the area. Thus rising material from deep levels within the mantle is pictured as wedging apart the pre-existing continent with splitting occurring along tensional lines of weakness developed at the top of the dome. It now appears from detailed stratigraphic work that doming occurred at some time after the extrusion of the basalts and thus probably after the continental fission had occurred (Soper et al. 1976). However, it still seems likely that the dome is related to the magmatism and I suggest that the most likely period of its formation was around the time of intrusion of the alkaline magmas, i.e. 50 m.y. ago during the early Eocene. This suggestion is supported by the fact that it has now been eroded deeply whereas the plateaus, which were probably regionally uplifted in the late Oligocene (see below), are to a large extent preserved. Thus its time of formation is limited to pre-late Oligocene, post-early Eocene and the only other major geological event in this period to which it might be related is the intrusion of numerous syenitic plutons in the Kangerdlugssuaq area.

The coastal flexure

This feature and its associated dyke swarm originally described by Wager & Deer (1938) and Wager (1974) is a major geological feature traceable along the coast for at least 400 km and recently shown to be present for at least this distance again offshore to the south (Larsen 1978). Nielsen (1975) has questioned the original interpretation of the mechanism of flexuring and suggested that it consists of a number of faults (reminiscent of those along the rift-valley margins in Africa) which downstep towards the Danmark Stræde. Other workers have reported coastal faulting farther to the north along its continuation, where the dike swarm is no longer seen (Fawcett et al. 1966, Birkenmajer 1972, M. Watt 1975) and similar faulting occurred throughout the Mesozoic in northern East Greenland (Surlyk 1977).

The effect of the coastal flexure has been to greatly modify the ice-covered plateaus and the inland dome, both of which plunge down steeply below sea level in the coastal areas as shown in Fig. 5. This figure has been constructed such that the contours are parallel to the dip of the lavas which is consistent with Wager & Deer's model of flexuring, but probably not with that of Nielsen (1975). As the exact number of faults and the magnitude of their throws is not known the Wager & Deer model has been retained for the purposes of this discussion.

Wager (1947) has noted additional faults and these have similarly been ignored in the present reconstruction. They would in a detailed analysis result in some modification but this would not destroy the overall result and, in fact, would have little effect on the map scale used here.

The importance of the coastal flexure, roughly following the coast for more than 800 km and controlling the location of the present coastline, can scarcely be doubted and this will be taken up in a subsequent section.

Timing of the coastal flexure and regional uplift

On the evidence presented by earlier writers (e.g. Wager & Deer 1938, Wager 1947) there can be no doubt that the coastal flexure dates from the early basic igneous episode. It deforms the basalts and gabbroic intrusions, dated at 54.6 ± 1.7 m.y., but not the syenitic ones, dated at 50.0 ± 0.4 m.y. (radiometric ages noted previously), and must therefore date from within this 5 m.y. period. However, Wager (1974) naturally assumed that the regional uplift, i.e. the time of uplift of the inland basalt plateau, must be correlated with the formation of the flexure, although evidence in support of this has not been forthcoming. On the contrary, investigations of post-basaltic sediments at Kap Brewster and Kap Dalton to the north indicate that the relief was subdued subsequent to the cessation of basaltic volcanism (Birkenmajer 1972, Soper & Costa 1976). It is not until the deposition of the later, possibly Miocene (Hassan 1953), sediments that there is evidence for uplift in the immediate area, in the form of the high energy sedi-

ments of the Kap Brewster Formation, possibly associated with major faulting in the above mentioned coastal fault swarm (Birkenmajer 1972). It was concluded that this uplift occurred in the period lower Oligocene to Miocene (i.e. ca. 40 to 25 m.y. ago).

That the regional epeirogenic uplift was significantly later than the doming is supported by the lower degree of denudation of the plateau in comparison to the dome. Discordancies in fission track ages for zircons and apatites from the Tertiary plutons of around 15 m.y. indicate also that an important event took place in the Oligocene (Brooks & Gleadow 1977 and in preparation).

It is surprising that uplift was not synchronous with volcanism. The classic development of oceans pictures upwarping of continental crust and rift formation. Later as the ocean begins to form, the continents retreat from each other and the rift margins gradually subside reaching sea level after about 100 m.y. (Vogt 1970). That there is no evidence of early uplift in this area thus seems to be anomalous. The only explanation that comes to mind (speculatively) is the fact that the dyke swarm also seems to be unusual, both in intensity and thickness, and Larsen (1978) suggested that this is related to an unusual type of spreading in this area at the time of anomaly 24 (50–60 m.y.) with massive attenuation of continental crust instead of oceanic crust formation. Such crustal attenuation combined with loading by intense basaltic dyke intrusion could account for the subsidence of the western Danmark Stræde area in the Paleocene.

It is perhaps worth noting that the Faeroe Islands, which on some pre-drift reconstructions (e.g. Bott & Watts 1971) are placed very close to Kangerdlugssuaq, have a very similar history (Waagstein 1977). Basalt extrusion took place about 54 m.y. ago after which the plateau was buckled into a series of domes and troughs. Epeirogenic uplift took place about the mid-Tertiary to give the present dissected plateau.

Present topography and the initial land surface

As noted previously, the rocks of the area fall into two main groups with respect to their very different powers of resistance to erosion. On the one hand are the easily eroded lavas, tuffs and sediments and on the other the gneisses of the basement which are extremely resistant. The former group consists of rocks which in general contain many planes of weakness (closely-spaced jointing, bedding, etc.) which are rapidly attacked by frost shattering, while the gneisses are massive and appear to resist this type of weathering. It is furthermore probable that the relative behaviour of these rock-types is similar for other weathering agents, a point to be considered when we remember that in the early Tertiary the local

climate was temperate to warm (Seward & Edwards 1943, Pedersen 1976).

My conclusion that the gneisses are more resistent to erosion than the basalts is based on observations that: a) rounded roches moutonnées, as for example on Kraemers Ø (see below), which date from the period of maximum glaciation are only preserved in areas of gneiss; and b) screes are negligible in amount in the gneiss country but extensive in the basalt country and this is surely an index of the rate of erosion. S. Funder (personal communication) is not in agreement with my conclusion and believes that in the Scoresby Sund area the two rock types erode differently but not necessarily at different speeds.

The plutonic rocks which cover a relatively restricted area in the immediate neighbourhood of Kangerdlugssuaq, show an intermediate behaviour, while the basalts near the outer coast, where they have been intensely invaded by dykes, erode in a different manner again.

The strong contrasts in resistance to erosion of the various rock groups can scarcely be evaluated quantitatively, but have clearly been an important factor in the development of the present land forms, and a more detailed description of the different erosional land forms will be found below.

Erosion of the inland dome and present drainage patterns

While the effects of the bedrock can usually be broadly understood other factors are not so clear. Thus, because both ice and water erode by virtue of movement under gravity, both produce broadly similar drainage patterns. In considering what mechanism has stripped away between 3 and 5 km of sediments and lavas from the inland dome exposing the underlying gneisses, it is not easy to decide if this was accomplished under conditions similar to those existing today, or whether normal fluviatile erosion should be invoked. However, as it is likely that most of this erosion took place during the early Tertiary prior to the regional uplift of the plateaus, as discussed above, and as the climate is believed to have been warmer at this time, the latter is more likely.

Originally, the impressive 2 km high southwest face of the Watkins Bjerge (Fig. 7) was taken to be a fault scarp (Wager in Watkins 1932). However, subsequent expeditions disproved this hypothesis and it is now known to be an erosional escarpment, where the upwarped lavas on the flanks of the dome have been truncated by erosional processes exposing the gneisses in the core of the dome. An indefinite plateau, at about 2200 m altitude running from around Sortekap to Nordfjord Plateau apparently represents the exhumed peneplain on which the sediments and lavas were deposited (Figs 7 and 8).

Nearer the heart of the dome, the more resistant rocks have been sculptured into shapely mountains of

Fig. 7. Oblique air photograph of the head of Kangerdlugssuaq looking eastwards. Nordfjord Plateau peneplain to the left and the alpine landscape to the south of Nordfjord to the right. The calving foot of the Kangerdlugssuaq Gletscher, which is about 5 km across, is in the foreground with a line of fresh moraine extending about 100 m up the lower slopes of Batbjerg. The erosional scarp of the Watkins Bjerge, rising to almost 4000 m and about 130 km distant can be seen in the upper right. Reproduced from a contact print. (With permission of the Geodetic Institute, Copenhagen, A. 128/78. COPYRIGHT).

alpine form such as Domkirkebjerget, Lemon Bjerg and Mítivagkat (Figs 7 and 8). On these mountains to the west of Frederiksborg Gletscher no obvious remains of the old pre-Tertiary peneplain are to be seen, but the summit levels clearly approximate to an extrapolation of this peneplain (Fig. 6) which is extensively preserved on the eastern side of the Frederiksborg Gletscher. The summits of all these basement mountains and probably those to the south of Kangerdlugssuaq also are believed to form a gipfelflur from which the amount of deformation over the dome was constructed.

If the removal of the central part of the dome is the result of prolonged fluviatile erosion it would be expected that consequent valleys would form running down the dip of the dome. A glance at the map shows that several clear examples can be found (Fig. 5). Thus the bend in the Christian IV's Gletscher located ca. 50 km inland occurs at just the point where an original watershed would have been located. Several good examples of drainage from the northern flanks of the dome are found in the Watkins Bjerge, Lindbergh Fjelde and Prinsen af Wales Bjerge and these are indicated in Fig. 5.

The origin of Kangerdlugssuaq is one of the more interesting problems and is related to the origin of fjords in general, a much discussed problem (see for example Holtedahl 1975: 3–6). Wager believed that it originated by headward cutting of a major drainage channel (down the groove in the dome formed by the inflection in the dyke-swarms trend at this point) with a consequent reversal of drainage from a valley to the north. On the

other hand, Brooks (1973) suggested that in order to explain such a large cleft cutting the highest part of the dome a tectonic control was likely. He suggested that Kangerdlugssuaq represents a side rift, or "failed arm" (Burke & Dewey 1973), of the mid-oceanic rift system whose main trend appears to have controlled the coastline of the Danmark Stræde. This interpretation was supported by the presence of a line of alkaline intrusions and dyke swarms parallel to the supposed rift. However, apart from two examples cutting the alkaline Gardiner Plateau intrusion at the head of the fjord (Frisch & Keusen 1977) and a possible fault along the line of Nordfjord (witnessed by a lower level of the pre-Tertiary erosion surface to the west of Nordfjord Gletscher in comparison to that on the east) it has not been possible to locate any definite rift-faulting in the area. It is therefore not a the present time possible to decide unequivocally between these alternatives, but it seems likely that an example of Wager's mechanism of headward erosion of southward flowing drainage and capture of a northward-flowing system is the Frederiksborg Gletscher. However, it does not to me seem possible to decide, if the agent involved was water or ice due to the very considerable modifications to these valleys during the Quaternary. Wager favoured glacial erosion in line with his suggestion for a very early formation for the inland ice (Wager 1933).

Perhaps the main argument in favour of Wager's hypothesis for the origin of the fjord is the staggered relationship between outer Kangerdlugssuaq and the inner part (Fig. 5). This he thought might reflect the fact that the original valleys were not directly opposite to

Fig. 8. Oblique air photograph looking northwards along the eastern side of Kangerdlugssuaq. In the foreground is the dissected alpine landscape between Watkins Fjord and Courtauld Fjord, whose gipfelflur merges into the exhumed Mesozoic peneplain seen in the distance on Nordfjord Plateau. This is overlain in the far distance by basalts at Urbjerget. Note the ice-polished lower fjord walls indicating the former height of the Kangerdlugssuaq Gletscher. Reproduced from a contact print. (With permission of the Geodetic Institute, Copenhagen, A. 128/78. COPYRIGHT).

each other and pointed out that the offset occurs about the highest point of the original dome. This feature is not easily explained by the alternative hypothesis.

Kangerdlugssuaq is about 75 km in length, from a line between Kap Hammer and Kap Deichmann and probably continues as an ice-filled cleft for a considerable distance under the inland ice. The outer part maintains an average width of about 6 km with mountains rising to 2 km and has depths of about 1 km. It has a well defined threshold (Royal Danish Hydrographic Office chart, 1965) and, whatever its origin, has clearly been extensively modified by ice. There is no evidence of reexcavation of old valleys as was suggested for northern East Greenland by Backlund (1930) and it has probably developed entirely since the early Tertiary.

An interesting feature to the west of Kangerdlugssuaq is the apparent control of the drainage patterns by the major plutons. Thus Hutchinson Gletscher divides around the Kap Edvard Holm gabbros. The Kangerdlugssuaq intrusion appears to have had its own radial drainage pattern, at least prior to regional ice-flooding, and the Gardiner intrusion (Frisch & Keusen 1977) occupies a particularly high nunatak which appears to act as a bastion against the powerful Kangerdlugssuaq Gletscher. This suggests to me that each of these intrusions originally expressed itself at the surface of the Kangerdlugssuaq dome as a topographic high, probably a volcanic edifice which influenced the consequent drainage off the dome. However, other interpretations (to my mind less likely) are possible. Wager (in Watkins 1932) noted a similar situation in the Kialêq district to the south.

Another notable exception to the overall pattern of radial drainage is to be seen in the coast-parallel valleys extending through Forbindelsesgletscher, the east–west reach of Mikis Fjord and through to I. C. Jacobsen Fjord, the parallel system through Watkins Fjord to the inner part of I. C. Jacobsen Fjord, and the depression through Ryberg Fjord, behind Søkongen Ø to Nansen Fjord. These valleys probably follow the fault lines described by Nielsen (1975) which were connected with the rift formation under continental break-up. Similar coast-parallel valleys are strikingly developed over a large area in the Angmagssalik district as noted by Wager (1932), who ascribed their origin to the same cause. They are also to be seen to the south of Scoresby Sund in an area where coast-parallel faulting is abundant (Birkenmajer 1972, M. Watt 1975).

Estimated from Fig. 5, the erosion of the dome has since the early Tertiary removed about 50000 km³ of rock material which has presumably been deposited on the shelf, and it is probably no coincidence that the shelf is exceptionally wide here (Iversen 1936, and Fig. 9). Such a volume of rock material would cover the entire shelf here to a depth of about 1 km (assuming a rough volume equivalence between the original rock and the derived sediment), but more probably occurs as a prograded mantle on the continental scope. A similar widening of the shelf is to be seen out from Sermilik in the Angmagssalik district and outside the mouth of Scoresby Sund, suggesting that these fjords were also the source of large volumes of clastic material. Sermilik also has a prominent off-shore trough carved by Quaternary glaciers as described for Kangerdlugssuaq below.

Preservation of the ice-covered plateau

In contrast to the impressive denudation shown in the inland dome, considerable areas of the original basalt surface have been preserved in the Seward Plateau and Geikie Plateau and on the plateau behind Gronau Nunatakker and also large areas to the north of these. This is believed to be strong evidence for a much later regional uplift. Ahlmann (1941) described widespread plateaus in northern East Greenland, particularly along the edge of the inland ice. Similarly, Wager (1933) noted the plateau areas of gneisses near the inland ice in the Mont Forel region of the Angmagssalik district. Extensive gipfelflur are also known to be present in South Greenland (Wegmann 1939). Raised peneplains are thus widespread in Greenland as in other North Atlantic areas.

In the case of the ice-covered plateaus to the north of Kangerdlugssuaq, we have a case where the peneplaned gneiss has been covered by basalt in the Tertiary, the original surface of which appears to have been preserved, in spite of its easily eroded nature. It therefore appears that another factor in addition to relatively recent uplift might be operative in preserving so well these plateaus. Wager (1933) believed that their small ice-caps, formed of almost debris-free ice with little movement, protected the surface of these high plateaus from the normal agents of weathering.

Wager was a supporter of the view that Greenland became glaciated at a very early stage, but his evidence for this was at the best slender and is contrary to most modern opinions. It is now believed that Greenland became glaciated at the beginning of the Pleistocene (Weidick 1976: 434–435), although there is evidence for Miocene glaciation of Antarctica and glaciations of Miocene-Pliocene age have been recognized in Alaska (Denton & Armstrong 1969). Wager (1933) suggested a Miocene age based on geomorphological arguments but later, with a better knowledge of the coastal dyke swarm and flexure, which is clearly Eocene, he seems to have revised his date upwards. I have suggested that the uplift of the high plateaus took place during the Miocene and this would explain their relatively good state of preservation, as they were originally low-lying and after uplift became armoured by the ice caps which prevented their destruction.
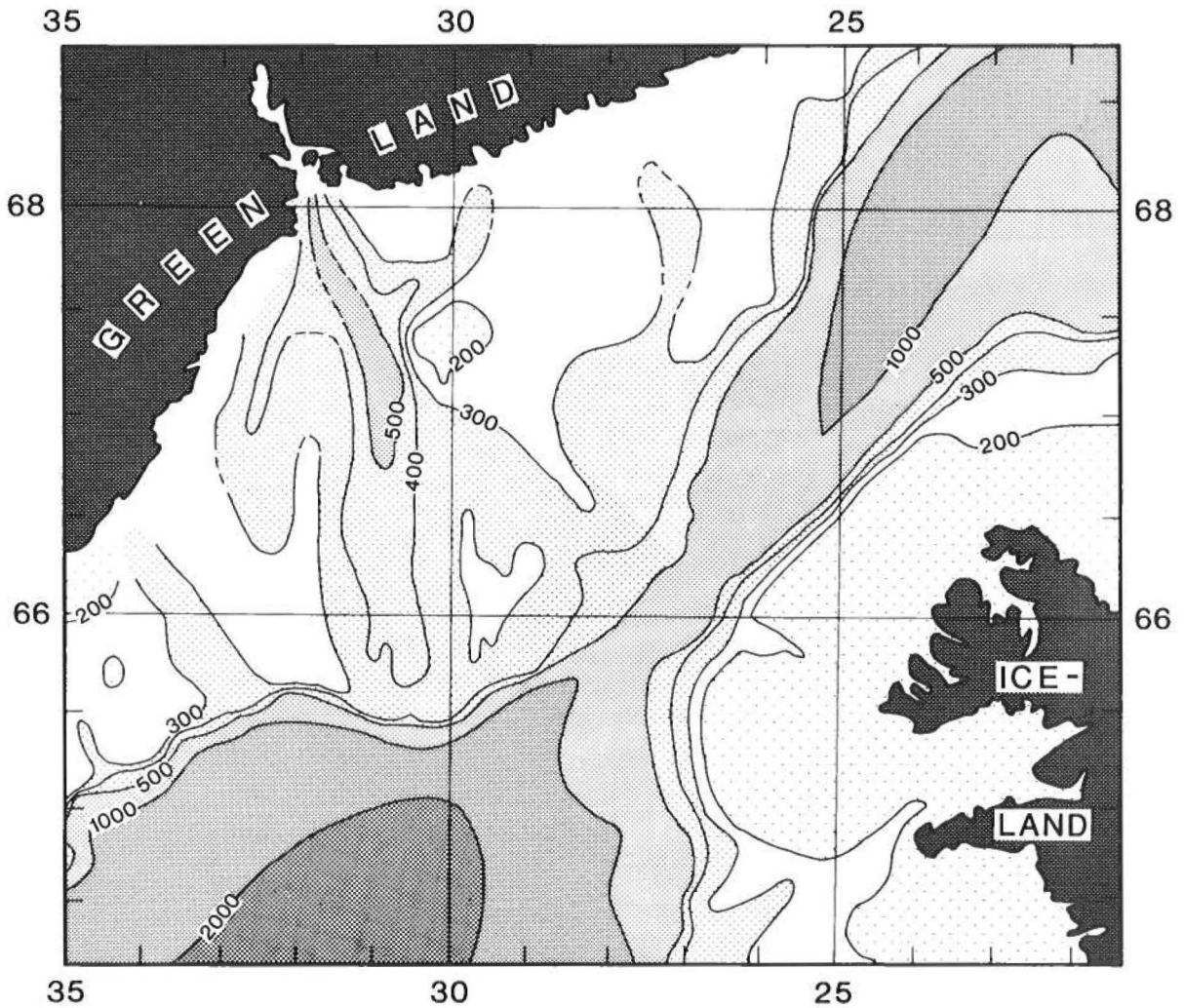
Fig. 9. Bathymetric map of the continental shelf offshore from Kangerdlugssuaq (reproduced from Iversen 1936). The area of the conspicuous fan is about 60,000 km² and is believed to consist of prograded clastic deposits derived from the erosion of the Kangerdlugssuaq dome, calculated to be about 50,000 km³ in volume. Note the deep trough carved by the former extension of the Kangerdlugssuaq Gletscher. Banks on the outer shelf are probably ice margin deposits.

Present day features

Glaciers

The area contains glaciers of every morphological type, varying from the continental ice-sheet on the inland side to valley glaciers, often of large dimensions, and small cirque glaciers in the coastal mountains. In addition there are regions where ice, either from the inland ice or generated locally, floods an area leaving only a few nunataks exposed. Finally, there are the small ice caps which have already been mentioned in connection with the high plateau. No glaciological studies have so far been carried out here and the following is a brief sum-

mary of information which may be largely gleaned from the maps.

Of the many large glaciers which reach the coast between Kangerdlugssuaq and Scoresby Sund, none drain from the inland ice proper, although they receive ice from the local ice-caps and many seem to have a contribution from the inland ice judging by the satellite photographs described by Matthews (1975 a). The Kangerdlugssuaq Gletscher (Fig. 7) and Nordfjord Gletscher are major outlets from the inland ice and drain a very large area; the Kangerdlugssuaq basin is visible on contour maps of the inland ice for more than 200 km inland. Although their productivity is unknown,
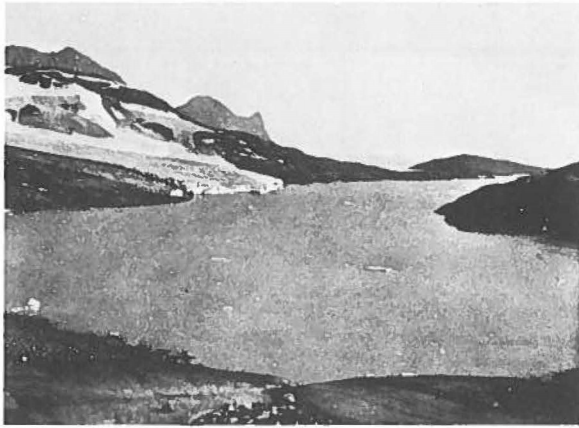
Fig. 10. Two photographs of the snout of Forbindelsesgletscher, a local glacier near Skærgården, showing retreat. The picture to the left is reproduced from Mikkelsen (1933: Fig. 17) and was taken in 1932. To the right a picture taken from a helicopter in 1970 from a slightly higher position. The glacier snout has thinned considerably and no longer discharges into the fjord. Note also the decrease in ice cover on the hillside beyond and the ice smoothed form of the low land on the right (Skærgårdshalvø).

it is probably high, as Kangerdlugssuaq is generally completely filled with discharged ice varying in size up to large bergs.

South of Kangerdlugssuaq glaciers front the coast over extensive distances and much of this ice may originate from the inland ice.

Glaciers of the area have typical gradients in their lower regions of 1 in 20 to 1 in 40 and there is no significant difference between the typical valley glaciers such as Sorgenfri Gletscher, Christian IV's Gletscher, Kronborg Gletscher and Borggraven on the one hand, and outlets of the inland ice such as Kangerdlugssuaq Gletscher and Nordfjord Gletscher on the other. The former type is limited over a considerable distance by high mountains while the Kangerdlugssuaq Gletscher in contrast is limited in this way for only about 20 km before fanning out into a major depression in the inland ice. These gradients do not seem to be appreciably different from those of the inland ice, where it reaches the coast unimpeded by extensive nunataks as to the south of Angmagssalik.

Christian IV's Gletscher is roughly comparable in length and width to glaciers such as the Beardmore in Antarctica, although it does not directly drain the ice sheet as these do. It probably differs considerably from typical Alpine and Himalayan glaciers and it does not seem likely that this ice-filled depression is an earlier river valley slightly modified by ice erosion but has developed over a prolonged period of glaciation, as discussed by Wager (1933) for the large valley glaciers of the Angmagssalik district.

All these glaciers show signs of retreat, even the Kangerdlugssuaq Gletscher (Fig. 7). Relatively fresh moraine is found on the hillsides above the glacier up to about 100 m above the present ice surface. There is unfortunately no clue as to the possible age of the re-

treat from this position, but similar unvegetated moraines elsewhere in Greenland are assumed to date from the period 1880–1920 (Ten Brink 1975) although the photograph in Fig. 7 was taken in 1932, and the ice was even then at about the same level as now. Wager (1937) noted a considerable retreat of the Kangerdlugssuaq Gletscher snout in the period 1930–1936 and retreat of the local Forbindelsesgletscher is shown in Fig. 10. However, recent years have seen a notable increase in the size of permanent snow patches. Large avalanche fans of a semipermanent nature completely obscure syenite exposures mapped by Wager along the northern foot of Admiraltinden (Kempe et al. 1970), and I have personally witnessed the disappearence of outcrops under such fans between 1970 and 1977.

Ice-dammed lakes

A lake marked on the 1:250,000 map (Geodetic Institute, Copenhagen) in the area between the Sorgenfri Gletscher and Christian IV's Gletscher about 20 km inland from the bend of Rybjerg Fjord was shown by Matthews (1975 b) to be intermittent. He described catastrophic drainage from this lake (hlaup) during which 0.5 km³ of water were discharged in a short period.

Ice-free areas

Areas which are not too steep for ice or snow to rest on and are ice-free at the present time are not extensive in the area. Such areas occur either on south facing slopes or on areas exposed to föhn winds (violent katabatic winds from inland, locally called "piteraq"). The outer coast is invariably highly glaciated, presumably due to

snow and not melting because of prevalent fogs during the melting season.

Particularly good examples of the effect of aspect are to be found in Amdrup Fjord, Mikis Fjord, I. C. Jacobsen Fjord and Ryberg Fjord. Here the south-facing slopes are snow-free and well vegetated, while their cirques are either ice-free or carry only small ice fields. In contrast, the opposite, north-facing slopes are almost entirely snow- and ice-covered with numerous small cirque glaciers reaching the sea.

Ice-free areas controlled by föhn winds occur in the inner parts of Kangerdlugssuaq, at Mudderbugt and around the lake at 700 m below Gardiner Plateau, an area known from experience to be particularly subject to such winds.

Other fairly extensive ice-free areas occur between Sorgenfri Gletscher and Christian IV's Gletscher (where the ice-dammed lake referred to above is situated) and between Christian IV's Gletscher and Rosenborg Gletscher. These areas are poorly known and no reason apart from their southerly aspect is known for their ice-free nature.

Bagnæsset, Amdrup Pynt and Kraemers Ø are some of the ice-free areas in the district which show clearly by their form that they have been totally submerged by ice-streams. Areas of this type are to be seen in Fig. 10 (Skærgården area) and to the east of Kærven (Fig. 13). Similar areas occur near the inland ice at Gardiner Plateau and northwards from Batbjerg. These areas are rounded and smoothed and almost totally scraped free of moraine. The highly resistant gneiss of the basement is completely unmodified by recent weathering except where small cirque glaciers have deposited moraines on this ice-smoothed surface. These moraines are comparable in size to those round similar Scandinavian and Alpine glaciers. As noted above such rounded, whale-back forms are not in evidence where the bedrock is basalt, presumably due to the much lower resistence of these rocks to weathering so that these land-forms have already been largely obliterated.

The characteristic land-forms of the syenitic rocks are vertical walls along the glaciers and head walls of cirques. Characteristic tower-like forms usually develop (Fig. 11) but the tops of these mountains often preserve remnants of an earlier more rounded topography, which is especially well seen along the northern side of Søndre Syenitgletscher and the southern side of Nordre Syenitgletscher. It is not clear whether this is a pre-glacial land surface of mature undulating relief, or whether it was formed during an earlier more extensive glaciation which completely overwhelmed these mountains which are over 1000 m high. Elsewhere, the mountain tops, e.g. in the area between Watkins Fjord and Courtauld Fjord (Figs 7 and 8), have a jagged, alpine aspect. Frost action on the coarse-grained plutonic rocks often produces a deep gravel cover whereas angular blocks form on the gneisses.

Basalts generally weather to stepped pyramid-like

forms with slopes of 45° (Fig. 2) except near the outer coast where they have been invaded and metamorphosed by the intense dyke swarm. Here they show a very characteristic slope profile of about 60° as seen in Fig. 12 which shows that in this area the rocks behave as a massive unit to erosion rather than a pile of individual flows as elsewhere (Wager 1934: 31). The situation is however slightly more complex than this simple analysis suggests, the basalts of this area by coincidence being of different composition and morphological type to those farther to the north and west (Brooks et al. 1976).

Extent of the former glaciation

The observations presented in the previous section indicate that the mountains of, for example, Kraemers Ø which are up to 1000 m high were to a large extent submerged by the Kangerdlugssuaq Gletscher at a time when it extended at least 60 km beyond its present snout. This conclusion is corroborated by the presence of a marked bench along the sides of Kangerdlugssuaq at a
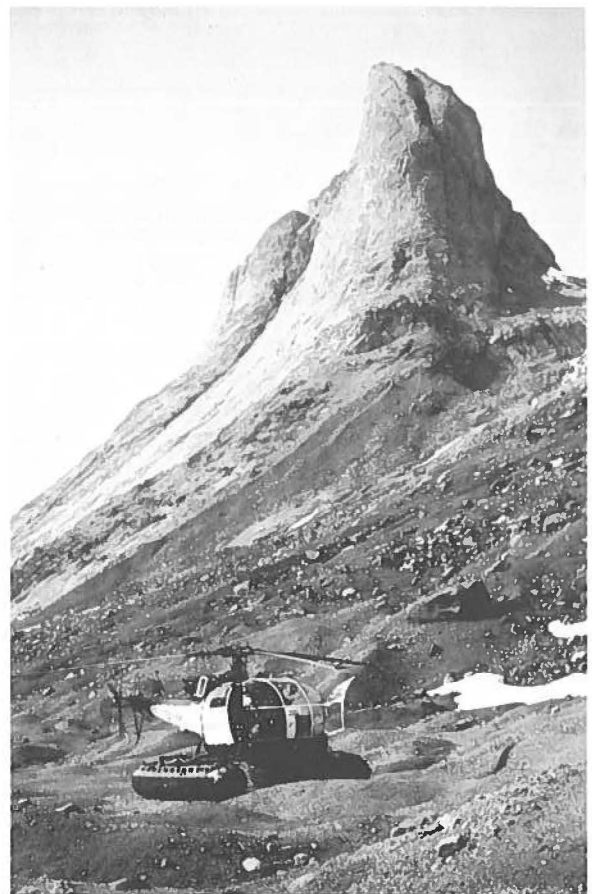

Fig. 11. Typical weathering forms in the plutonic rocks, in this case the pulaskites of the Kangerdlugssuaq intrusion (Kempe et al. 1970), Citadellet, inner part of Amdrup Fjord.

Fig. 12. Basalts in the area of the dyke swarm along the outer coast at the entrance to Mikis Fjord showing characteristic slope profiles.

height of some 800 m (Figs. 8 and 13) which separates the ice-smoothed lower slopes from the higher alpine-type topography. This bench clearly represents a prolonged period of time when the glacier surface in this area was about 800 m higher that the present sea level. Wager (unpublished notes) found blocks of gneiss on the higher parts of the Skaergaard gabbros at an altitude of about 1200 m. It is not at present clear whether these were formed prior to the period represented by the bench as these peaks are of the Alpine type, but it is perhaps most simple to conclude that they were deposited during brief periods when the glacier was even higher than the 800 m bench and the alpine topography developed in the crumbly gabbros much more rapidly than in the resistant gneisses. It therefore seems that the Kangerdlugssuaq Gletscher formerly (probably in Weichselian times) stood for considerable periods of time at about 800 m above present sea level and at times reached as high as 1200 m in the fjord region. At this time the glacier tongue may have extended far out onto the continental shelf. This glacier probably retreated at the end of the Pleistocene fairly rapidly as no lower benches are visible.

It is not the purpose of this paper to discuss the offshore morphology, but it seems appropriate to note that a trough in excess of 700 m deep extends to about 120 km off Kap Hammer and is still visible as much as 270 km offshore as a depression below 400 m. (The bathymetric chart of the shelf by Iversen 1936 is probably still the best map of the offshore regions in this area and a relevant section is reproduced in Fig. 9). This trough was undoubtedly cut by the extended glacier. The shelf here has a width of about 300 km and a depth at its outer edge of 500 m, so it is clear that a glacier of the dimensions in question would be in contact with the bottom even if considerable expansion of the tongue occurred beyond the confines of the fjord, especially when the eustatic lowering of sea level at this time is taken into account. A similar conclusion was reached by Sommerhoff (1973) for the shelf area off south-west Greenland, and the bathymetric features seen in Fig. 9 probably reflect similar glacial erosion and construction features to those described from the south.

At the time of this maximum extension of the ice, the coastal belt to the east of Kangerdlugssuaq probably resembled that to be seen today to the south of the fjord

18

Fig. 13. Kærven (1134 m) from a point in the air above the snout of Kælvegletscheren. The bench at about 800 m caused by the former glacial extension can be clearly seen as a discontinuity in the slope on the left of the peak. The unnamed twin peaks in the right centre are 1390 m high.

with only occasional mountain tops breaking the ice and long lines of ice cliffs fronting the sea. At the same time the ice in the hinterland was much deeper than at the present time with most of the nunataks submerged. The higher mountains were however above the ice level. It is possible that the subdued mountains to the west of Christian IV's Gletscher and north of the head of Watkins Fjord, whose heights are below that expected for the summit level in the section (Fig. 6), owe their form to this period when they were overtopped by the ice sheet.

Morphological features of the coastlines

The importance of the coastal flexure and dyke swarm in controlling the trend of the coastline in this area was noted previously. In view of the relatively recent origin for all the ocean basins surrounding Greenland and the evidence for coast-parallel structures in other areas (e.g. Surlyk 1977) it is probable that most of Greenland's present-day coastlines are dominated by little-modified rift margin type faulting. An exception to this is in northern East Greenland where rifting began in the Palaeozoic and continued throughout the Mesozoic. In the area discussed here, second order modifications have been effected by glacial and marine erosion and constructional features are almost non-existant.

The coast between Kangerdlugssuaq and Scoresby Sund (Blosseville Kyst) is characterized by steep head-lands divided by fjords. These fjords are all glacially-formed, possibly sometimes along lines of faulting (e.g. the E-W portions of Mikis Fjord and Ryberg Fjord, as noted previously). Wager (unpublished notes) discussed the typical steep cliffs of the outer coast and was doubtful if wave action in the area would be sufficient to account for them. However, in view of the fact that this coast faces the open Atlantic, which most years at the present time is ice-free for several months in the autumn, I do not feel this is any problem. These cliffs vary in slope according to the type of rock, basalts where unmetamorphosed outside the regions of intense dyking rise at 45°, while the metamorphosed basalts and gneisses (when not ice-smoothed) give steeper headlands (as discussed above).

Beaches are of negligible extent along the coast. Deltas occur at the heads of some fjords (e.g. the mouth of

Sødalen, Mikis Fjord) and curved shingle spits are found in Mikis Fjord, at Kap Dalton and at Kap Ravn. However, these only occur in areas of easily eroded basalt and even here they are generally absent, only fallen blocks or scree being present. A curved spit in the inner part of Mikis Fjord is the beautifully semi-circular end moraine of a glacier whose snout is now about 500 m up the hillside (illustrated by Mikkelsen 1933: Fig. 29).

Raised beaches occur but have not been systematically studied. They are even less extensive than the present beaches and generally consist of small deltaic cones. No datable material has been found in them. Wager (unpublished notes) reports raised beaches of 10–15 m from Skærgården, d'Aunay Bugt, Barclay Bugt and in bays north of Kap Ewart and Kap Dalton. I have measured heights (above present high water line to the top of the steep delta front) of 10–12 m and 15 m on Skærgården, at 26–28 m at Skåret on Kraemers Ø and at 38 m at the mouth of Sødalen. In Vandfaldsdalen raised beaches and small deltaic fancs occur at 10–12 m, 18 m, 28–30 m, 33–35 m and 50 m, although both here and in Sødalen some of these may be deposits in ice-dammed lakes. The lowest terrace in Vandfaldsdalen is that on which the graves, fox traps and meat caches described by Degerbøl (1936: 39) are found. Beaches were also observed but not measured at Mudderbugt, Ryberg Fjord, Wiedemann Fjord and Nansen Fjord. At Nansen Fjord Wager (unpublished notes) recorded heights of 95 and 195 feet (i.e. ca. 29 and 59 m). At Tugtilik and Nigertussoq, 225 km south of Kangerdlugssuaq, Storey et al. (1968) recorded raised beaches at 10, 15, 20 and 43–44 m above mean sea level. In the Kap Dan area of the Angmagssalik district Wager (in Watkins 1932) recorded a strandflat but nothing like this occurs in the district described here.

Conclusion

Ahlmann (1941: 49) commented that anybody who is familar with Norwegian topography will find, when he gets a general view of any part of northern East Greenland that it is in many respects very like a Norwegian landscape. Similar land forms are met with in other areas of the North Atlantic (Scotland, Spitsbergen and Labrador) in as much as raised summit levels (gipfelflur) are a dominant feature of the topography. It has always presented a considerable problem to date the formation of these peneplains and their times of uplift (see discussion in Ahlmann 1941). Similar landforms are found in the Kangerdlugssuaq district and, if it is assumed that all these areas have a similar origin, then a fortunate combination of circumstances in the Kangerdlugssuaq district allows a clearer insight into the problem than is perhaps obtainable elsewhere. Here, early Tertiary tectonism and volcanism which is accurately

dated, combined with a degree of denudation which allows the original landforms to be reconstructed, permits a detailed investigation of the chronological aspects of the various elements of the present day topography. The description presented in this paper is merely a reconnaissance of the possibilities and can no doubt be greatly amplified by future workers.

Undoubtedly, the most significant findings in this paper, which will be explored in more detail in a subsequent paper dealing with fission track dating, are the deductions concerning the timing of major epeirogenic movements in the area. These movements have given rise to the highest mountains in the Arctic, which are nevertheless the eroded stumps of much larger structures. The Kangerdlugssuaq dome can reasonably be assumed to have formed in connection with mantle upwelling, which has also caused the alkaline volcanism. An important task of the future is to determine the cause of the regional uplift, which according to this paper took place appreciably after the magmatism, and most probably in Oligocene time. There is no ready explanation as to why these movements should have been delayed in this manner, but it is a phenomenon well-known from other regions (see for example Gilluly et al. 1969).

References

Ahlmann, H. W. 1941. Studies in northeast Greenland, 1939–1940. Part I. The main morphological features of northeast Greenland. – Geogr. Annlr. 23: 148–182.

Backlund, H. 1930. Contributions to the geology of northeast Greenland. – Meddr Grønland 74 (11): 207–296.

Baker, B. H. & Wollenberg, J. 1971. Structure and evolution of the Kenya rift valley. – Nature, Lond. 229: 538–542.

Birkenmajer, K. 1972. Report on investigations of Tertiary sediments at Kap Brewster, Scoresby Sund, East Greenland. – Rapp. Grønlands geol. Unders. 48: 85–91.

Bott, M. H. P. & Watts, A. B. 1971. Deep structure of the continental margin adjacent to the British Isles. – Rept. Inst. Geol. Sci. 70/14: 89–109.

Bridgwater, D., Davies, F. B., Gill. R. C. O., Gorman, B. E., Myers, J. S., Pedersen, S. & Taylor, P. 1978. Precambrian and Tertiary geology between Kangerdlugssuaq and Angmagssalik, East Greenland. – Rapp. Grønlands geol. Unders. 83, 17 pp.

Brooks, C. K. 1973. Rifting and doming in southern East Greenland. – Nature, Lond. Phys. Sci. 244: 23–25.

Brooks, C. K. & Gleadow, A. J. W. 1977. A fission-track age for the Skaergaard intrusion and the age of the East Greenland basalts. – Geology 5: 539–540.

Brooks, C. K., Nielsen, T. F. D. & Petersen, T. S. 1976. The Blosseville Coast basalts of East Greenland: their occurrence, composition and temporal variations. – Contr. Mineral. Petrol. 58: 279–292.

Brown, P. E., Van Breeman, O., Noble, R. H. & Macintyre, R. M. 1977. Mid-Tertiary igneous activity in East Greenland – the Kialineq complex. – Contr. Mineral. Petrol. 64: 109–122.

Burke, K. & Dewey, J. F. 1973. Plume-generated triple junctions: key indicators in applying plate tectonics to old rocks. – J. Geol. 81: 406–433.

Courtauld, A. 1936. A journey in Rasmussen Land. – Geogrl. J. 88: 193–215.

Degerbøl, M. 1936. The former Eskimo habitation in the Kangerdlugssuaq district, East Greenland. – Meddr Grønland 104 (10): 54 pp.

Fawcett, J. J., Rucklidge, J. C. & Brooks, C. K. 1966. Geological expedition to the Tertiary basalt region of Scoresby Sund, East Greenland. – Nature, Lond. 212: 603–604.

Frisch, W. & Keusen, H.-R. 1977. Gardiner intrusion, an ultramafic complex at Kangerdlugssuaq, East Greenland. – Bull. Grønlands geol. Unders. 122: 62 pp.

Denton, G. H. & Armstrong, R. L. 1969. Miocene/Pliocene glaciations in southeastern Alaska. – Am. J. Sci. 269: 1121–1142.

Gass, I. G. 1972. The role of lithothermal systems in magmatic and tectonic processes. – Earth Sci., Leeds 8: 261–273.

Gilluly, J., Waters, A. C. & Woodford, A. O. 1968. Principles of Geology (3rd edition). – Freeman, San Francisco: 687 pp.

Hassan, M. Y. 1953. Tertiary faunas from Kap Brewster, East Greenland. – Meddr Grønland 111 (5): 42 pp.

Holtedahl, H. 1975. The geology of the Hardangerfjord, West Norway. Norges geol. Unders., Bull. 36: 87 pp.

Iversen, T. 1936. Sydøstgrønland Jan Mayen fiskeriundersøkelser. – Rep. Norw. Fishery mar. invest. (1): 171 pp.

Kempe, D. R. C., Deer, W. A. & Wager, L. R. 1970. Geological investigations in East Greenland. Part VIII. The petrology of the Kangerdlugssuaq alkaline intrusion, East Greenland. – Meddr Grønland 190 (2): 49 pp.

Larsen, H. C. 1978. Offshore continuation of the East Greenland dyke swarm and its bearing on North Atlantic ocean formation. Nature, Lond. 274: 220–223.

Matthews, D. W. 1975a. An observation on glaciation in the Watkins Mountains, East Greenland. – Polar. Rec. 17: 398–399.

Matthews, D. W. 1975b. Discovery of an ice-dammed lake in East Greenland. – Polar Rec. 17: 369–397.

Mikkelsen, E. 1933. The Scoresby Sound Committee's 2nd East Greenland Expedition in 1932 to King Christian IX's Land. Report on the expedition. – Meddr Grønland 104 (1): 71 pp.

Nielsen, T. F. D. 1975. Possible mechanism of continental breakup in the North Atlantic. – Nature, Lond. 253: 182–184.

Pedersen, K. R. 1976. Fossil floras of Greenland. – In: Escher, A. & Watt. W. S. (eds.), Geology of Greenland. – The Geological Survey of Greenland, Copenhagen: 519–535.

Ravn, J. P. J. 1933. New investigations of the Tertiary at Cape Dalton East Greenland. – Meddr Grønland 105 (1): 15 pp.

Seward, A. C. & Edwards, W. N. 1941. Fossil plants from East Greenland. – Ann. Mag. nat. Hist., Ser. II, 8: 169–176.

Sommerhoff, G. 1973. Formenschatz und morphologische Gliederung des südostgrönlandischen Schelfgebietes und Kontinentalabhanges. – "Meteor" Forsch. Ergebnisse, Reihe C 15: 1–54.

Soper, N. J. & Costa, L. I. 1976. Palynological evidence for the age of Tertiary basalts and post-basaltic sediments at Kap Dalton, central East Greenland. – Rapp. Grønlands geol. Unders. 80: 123–127.

Soper, N. J., Higgins, A. C., Downie, C., Matthews, D. W. & Brown, P. E. 1976. Late Cretaceous – early Tertiary stratigraphy of the Kangerdlugssuaq area, east Greenland, and the age of opening of the north-east Atlantic. – J. geol. Soc. Lond. 132: 85–104.

Surlyk, F. 1977. Mesozoic faulting in East Greenland. – Geologie Mijnb. 56: 311–327.

Ten Brink, N. W. 1975. Holocene history of the Greenland ice sheet based on radiocarbon-dated moraines in West Greenland. – Meddr Grønland 201 (4): 44 pp.

Vogt, P. R. 1970. Magnetized basement outcrops on the south-east Greenland continental shelf. – Nature, Lond. 226: 743–744.

Waagstein, R. 1977. The geology of the Faeroe plateau. – Unpubl. dissertation, University of Copenhagen: 115 pp.

Watkins, H. G. 1932. The British Arctic Air Route Expedition. – Geogrl. J. 79: 353–367 & 466–501.

Wager, L. R. 1933. The form and age of the Greenland ice cap. – Geol. Mag. 70: 145–156.

Wager, L. R. 1934. Geological investigations in East Greenland. Part I. General geology from Angmagssalik to Kap Dalton. – Meddr Grønland 105 (2): 46 pp.

Wager, L. R. 1935. Geological investigation in East Greenland. Part II. Geology of Kap Dalton. – Meddr Grønland 105 (3): 32 pp.

Wager, L. R. 1937. The Kangerdlugssuaq region of East Greenland. – Geogr. Jour. 90: 393–425.

Wager, L. R. 1947. Geological investigations in East Greenland. Part IV. The stratigraphy and tectonics of Knud Rasmussens Land and the Kangerdlugssuaq Region. – Meddr Grønland 134 (5): 64 pp.

Wager, L. R. & Deer, W. A. 1938. A dyke swarm and crustal flexure in East Greenland. – Geol. Mag. 75: 39–46.

Watt, M. 1975. Photo-reconnaissance of the Blosseville Kyst between Steward Ø and Søkongens Bugt, central East Greenland. – Rapp. Grønlands geol. Unders. 75: 90–95.

Watt, W. S. 1969. The Tertiary basalts of Scoresby Sund, a preliminary report. – Rapp. Grønlands geol. Unders. 21: 36–40.

Watt, W. S. 1970. Preliminary report of the mapping of the basalt area, South Scoresby Sund. – Rapp. Grønlands geol. Unders. 30: 30–37.

Watt, W. S. & Watt, M. 1971. Preliminary report of the mapping of the basalts of parts of Milne Land and Gåseland. – Rapp. Grønlands geol. Unders. 37: 42–50.

Watt, W. S. & Watt, M. 1973. Preliminary report on the mapping of the basalts of the inner part of Scoresby Sund. – Rapp. Grønlands geol. Unders. 58: 37–42.

Wegmann, C. E. 1939. Einleitung zur Vortragsreihe über die Geologie von Grönland. – Mitt. naturf. Ges. Schaffhausen 16.

Weidick, A. 1976. Glaciation and the Quaternary of Greenland. – In: Escher, A. & Watt, W. S. (eds.), Geology of Greenland. – The Geological Survey of Greenland, Copenhagen: 430–458.

Wentworth, C. K. & Macdonald, G. A. 1953. Structures and forms of basaltic rocks in Hawaii. – Bull. U.S. geol. Surv. 994: 98 pp.

Greenland Geoscience

1979.

Silver- and bismuth-rich galena concentrates have been produced for more than 70 years as a byproduct in the dressing of the crude cryolite from Ivigtut, South Greenland.

Concentrates from the years 1937 to 1962 contained from 0.44 % Ag and 0.74 % Bi to 0.94 % Ag and 1.93 % Bi. Conspicuous increases in the content of these elements appeared twice within this time interval, namely in 1955 and in 1960. Thus it seems that crude cryolite from specific areas within the mine carried galena high in silver and bismuth. This promoted a detailed study of the common Ivigtut galena and associated sulphides.

An outline of the geological setting of the deposit is given. The deposit is divided into two main bodies – the cryolite body and the quartz body. Both ate subdivided into units characterized by their content of siderite and fluorite. Galena samples from these units and from rock types surrounding the deposit have been studied.

Galena from units characterized by siderite follows the compositional pattern found in the galena concentrated, whereas the sparse galena mineralizations from unirs characterized by fluorite contain much smaller amounts og silver and bismuth, less than 0.2 %. However, within the fluorite-bearing units, two peculiar parageneses reveal high contents of silver and bismuth wxpressed by the presence of particular minerals such as marildite-aikinite and gustavite.cosalite respectively.

Further trace element studies on selected galena samples emphasize Sn and Te as chemically characteristic og the galena and of the sulphide-carbonate phase of the deposit.

The temperature of formation of the main part of the deposit is placed at 550–400°C, and between 300 and 200°C certain parts of the fluorite cryolite and the fluorite zone.

Greenland Bioscience

1979.

Annual hydrographic observations, measurements of primary production, and sampling of zooplankton were undertaken in Southwest Greenland waters in the 1950s and –60s. In the coastal area and at the entrance to Godthåbsfjord winter cooling normally extends to the bottom, resulting in a vertical mixing of the water and an effective replenishment of nutrients at the surface. The subsequent production rate is, therefore, high with an average annual gross production calculated to about 160 g C – m^2. In the inner fjord regions the stratification is normally much more stable with persisting warm bottom water, and the production is, therefore, lower here than in the coastal area. The seasonal variation in the relations between daylight, primary production, phosphate, and quantity of zooplankton is, presumably, representative of the coastal waters at SW Greenland. A maximum in primary production in spring is normally followed by another maximum in late summer. The number of animals in the microplankton samples from the upper 30 m (the productive layer) is at its maximum simultaneously with the second maximum of the primary production, while the maximum of the macroplankton biomass (taken by stramin net) extends until late autumn in the coastal and outer fjord regions.

A maximum of the macroplankton biomass during winter in the deep water layers in the inner Godthåbsfjord, caused by inflow of warm bottom water, stable stratification and cooled outflowing surface water acting as a barrier to the ascent of the animals, is assumed to be normal to the open, non-threshold, W Greenland fjords.

Seasonal vertical migration of the zooplankton is indicated by Hensen net hauls from different depths. There is a concentration of zooplankton in the upper water layers in April–September and a deeper concentration from autumn to spring.

Annual cycles of various animal groups are described for holoplankton and meroplankton, separately. Holoplankters are normally dominant, copepods being the most numerous group. Meroplankters, especially bottom invertebrate larvae, are relatively numerous in the microplankton in spring and summer with *Balanus* nauplii dominant in spring and lamellibranch larvae in the following months. In a special section on fish eggs and larvae it is shown *l.a.* that cod eggs and larvae are normally concentrated in the upper 50 m, where they are much exposed to temperature variations, while eggs and larvae of American plaice occur also in deeper water. This may partly explain why the cod stock is more vulnerable to low temperatures.

It is shown that the epipelagic plankton fauna in the survey area in terms of growth and mode of development is more similar to the arctic than to the boreal fauna. It could therefore be termed subarctic, which also corresponds to the environmental conditions in the area.

Instructions to authors

Manuscripts will be forwarded to referees for evaluation. Authors will be notified as quickly as possible about acceptance, rejection, or desired alterations. The final decision rests with the editor. Authors receive two page proofs. Prompt return to the editor is requested.

Alterations against the ms. will be charged to the author(s). Twenty five offprints are supplied free. Order form, quoting price, for additional copies accompanies 2nd proof. Manuscripts (including illustrations) are not returned to the author(s) after printing unless especially requested.

Manuscript

General. – Manuscripts corresponding to less than 16 printed pages (of 6100 type units) incl. illustrations, are not accepted. Two copies of the ms. (original and one good quality copy), each complete with illustrations should be sent to the Secretary.

All Greenland placenames in text and illustrations must be those authorized. Therefore sketch-maps with all the required names should be forwarded to the Secretary for checking before the ms. is submitted.

Language. – Manuscripts should be in English (preferred language), French, or German. When appropriate the language of the ms. must be revised before submission.

Title. – Titles should be kept as short as possible and with emphasis on words useful for indexing and information retrieval.

Abstract. – An English abstract should accompany the ms. It should be short, outline main features, and stress novel information and conclusions.

Typescript. – Page 1 should contain: (1) title, (2) name(s) of author(s), (3) abstract, and (4) author's full postal address(es). Large mss. should be accompanied by a Table of Contents, typed on separate sheet(s). The text should start on p. 2. Consult a recent issue of the series for general lay-out.

Double space throughout and leave a 4 cm left margin. Footnotes should be avoided. Desired position of illustrations and tables should be indicated with pencil in left margin.

Underlining should only be used in generic and species names. The use of italics in other connections is indicated by wavy line in pencil under appropriate words. The editor undertakes all other type selection.

Use three or fewer grades of headings, but do not underline. Avoid long headings.

References. – Reference to figures and tables in the text should have this form: Fig 1; Figs 2–4, Table 3. Bibliographic references in the text are given as: Shergold (1975: 16) and (Jago & Daily 1974b).

In the list of references the following usage is adopted:

Journal: Tarling, D. H. 1967. The Palaeomagnetic properties of some Tertiary lavas from East Greenland. – Earth planet. Sci. Lett. 3: 81–88.

Book: Boucot, A. J. 1975. Evolution and extinction rate controls. – Elsevier, Amsterdam: 427 pp.

Chapter (part): Wolfe, J. A. & Hopkins, D. M. 1967. Climatic changes recorded by Tertiary landfloras in northwestern North America. – In: Hatai, K. (ed.), Tertiary correlations and climatic changes in the Pacific. – 11th Pacific Sci. Congr. Tokyo 1966, Symp.: 67–76.

Title of journals should be abbreviated according to the last (4th) edition of the World List of Scientific Periodicals (1960) and supplementary lists issued by BUCOP (British Union-Catalogue of Periodicals). If in doubt, give the title in full.

Greenland Geoscience should be registered under *Meddelelser om Grønland.* Example (with authorized abbreviations): *Meddr Grønland, Geosci.* 1, 1979.

Illustrations

General. – Submit two copies of each graph, map, photograph, etc., all marked with number and author's name. Normally all illustrations will be placed within the text; this also applies to composite figures.

All figures (incl. line drawings) must be submitted as glossy photographic prints suitable for direct reproduction, i.e. having the format of the final figure. Do not submit original artwork. Where appropriate the scale should be indicated in the caption or in the illustration.

The size of the smallest letters in illustrations should not be less than 1.5 mm. Intricate tables are often more easily reproduced from line drawings than by type-setting.

Colour plates may be included at the author's expense, but the editor should be consulted before such illustrations are submitted.

Size. – The width of figures must be that of a column (77 mm) or of a page (160 mm). Remember to allow space for captions below full page figures. Maximum height of figures (incl. captions) is 217 mm. Horizontal figures are preferred.

If at all possible, fold-out figures and tables should be avoided.

Caption. – Caption (two copies) to figures should be typed on separate sheets.

Greenland Geoscience 1