MEDDELELSER OM GRØNLAND
UDGIVNE AF
KOMMISSIONEN FOR VIDENSKABELIGE UNDERSØGELSER I GRØNLAND
BD. 188 · NR. 1

GEOMORPHOLOGY OF INGLEFIELD LAND, NORTH GREENLAND

BY

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WITH 48 FIGURES AND 11 TABLES
IN THE TEXT

KØBENHAVN
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BIANCO LUNOS BOGTRYKKERI A/S
1969

Abstract

The coastal cliffs, structural terraces, and bedrock valleys of Inglefield Land resulted from the uplift and fluvial dissection of a late Tertiary peneplain. The straits and channels between Ellesmere Island and North Greenland may be due to the submergence of this dissected peneplain by the weight of the existing Greenland ice cap.

Inglefield Land was completely glaciated and the Carey Øer, 70 miles offshore from Thule, were covered by the Greenland Indlandsis, proving that during the Wisconsin the straits and channels between Ellesmere Island and Greenland were nearly or completely filled with glacial ice. C-14 measurements indicate that the deglaciation of the coastal areas near Rensselaer Bugt occurred before 8200 ± 300 B.P., and near Dallas Bugt before 6180 ± 200 B.P.

Elevated deltas and beaches show that the marine limit is approximately 285 feet above sea level near Force Bugt and 210 feet at Dallas Bugt, 70 miles to the northeast. The marine terraces reported by Bøggild at 1050–1800 feet above sea level between 77–81 degrees do not exist, and the post-Wisconsin epeirogenic movement postulated by Daly to account for them did not take place.

Permafrost, solifluction, block fields, nivation cirques, earth hummocks, non-sorted circles and polygons, and ice-wedge and sand-wedge furrows are common.

Although small dunes, loess, ventifacts, and wind-eroded earth hummocks are present, the work of wind is of minor importance.

Surface efflorescences, travertine on the bottoms of fragments, solution-faceted and solution-shaped fragments, deposition of travertine from ponds and streams, and meteorologic data all prove that the climate has been arid for thousands of years.

CONTENTS

INTRODUCTION

Inglefield Land is located along the northwest coast of Greenland from approximately 78°10′ N. to 79°10′ N. and from 66° W. to 73° W. It is bounded on the northeast by the Humboldt Gletscher, on the east and south by the Greenland Indlandsis, and on the west and northwest by Smith Sund and Kane Bassin. It extends for approximately 110 miles in a southwest-northeast direction and is as much as 40 miles wide (fig. 1).

Pre-Cambrian igneous and metamorphic rocks are widespread, as are late Pre-Cambrian and early Paleozoic sedimentary rocks (Cowie, 1961, p. 7, 17; Troelsen, 1950, Geol. map). Sills and dikes of dolerite cut the Pre-Cambrian rocks in the southeastern part of Inglefield Land. Glacial, glacio-fluvial, and periglacial deposits are widespread and common, although considerable areas of Pre-Cambrian rocks crop out. The highest point, a few miles from the edge of the Indlandsis, is approximately 3800 feet above sea level, but more than half the area is below 2000 feet. High cliffs cut in sedimentary rocks are found along most of the coastline but drowned valleys, promontories, and small islands characterize part of it. Extensive plateaus underlain by sedimentary rocks extend inland from the coastal cliffs. Seven glacier-fed streams run, in places in bedrock gorges, from the ice cap to Smith Sund and Kane Bassin. Inglefield Land is entirely ice free and in general devoid of vegetation except for small areas where there is abundant surface water. Although there are many perennial snow patches, Inglefield Land is below the climatic snowline.

The field work was carried out during the summers of 1953, 1963, and 1965. It was supported logistically by the U.S. Army (Transportation Corps, CRREL, and Research Support Group) and financially by the National Science Foundation, by Tufts University, and by the Stanford Research Institution, which in 1953 was under contract with the U.S. Army. During the 1953 field season Inglefield Land was reached from Thule by travelling in weasels over the Greenland Indlandsis. During the 1963 and 1965 field seasons personnel, food, fuel, and equipment were transported by helicopter from Thule to Inglefield Land, and 9 caches were dropped at strategic localities. The field work was carried out by moving on foot from cache to cache. In this way a reconnaissance

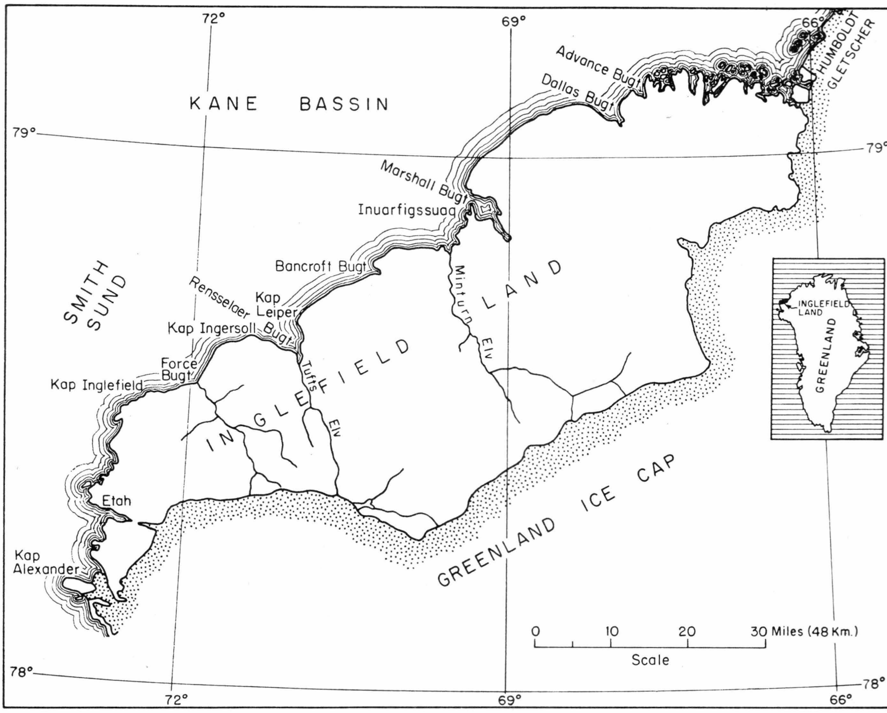

Fig.1. Map of Inglefield Land (after U.S. Air Force).

investigation was made of the coast from a point approximately 4 miles southwest of Force Bugt northeast almost to Advance Bugt, as well as of the Tufts Elv valley from the coast back to the Indlandsis (fig. 1). In the course of this work the writer and his assistants covered between 400–500 miles on foot.

Peter Roeder, Richard Meserve, and Christopher Miller, who were then students at Tufts University; Stephen Clark, then a student at Harvard University; William Meserve, a recent graduate of Tufts University; and Sergeant Carney Hampton, U.S. Army, ably assisted the writer in the field. They were excellent geologic assistants and pleasant companions. Without their dedication and their willingness to carry heavy loads for long distances, little reconnaissance field work could have been accomplished. Warm thanks must be given to Messrs. Alfred Clebsch, Jr., Joseph Hartshorn, and Richard W. Lemke, all of the U.S. Geological Survey, who, while working on their own geologic problems during the 1953 field season, nevertheless found time to visit some of the writer's outcrops and to make valuable suggestions as to their significance. Mr. A. D. Hastings, U.S. Army Quartermaster Research and Engineering Center, Natick, Massachusetts, compiled the meteorologic data and kindly made them available to the writer. Dr. Charles E. Stearns, Tufts University, critically read the manuscript and made many valuable suggestions for its improvement. Mention must also be made of the helicopter pilots of the U.S. Army who transported personnel and equipment from Thule to Inglefield Land during the 1963 and 1965 field seasons. Colonels Winston Butscher, Robert Giesen, and Neil B. Prentice of the U.S. Army extended to the writer and his field assistants many courtesies which will not be forgotten.

PRE-CAMBRIAN PALEOPLAIN

KOCH (1933, p. 14, 34), BENTHAM (1936, p. 428), WORDIE (1938, p. 399), MALAURIE (1955, p. 212–214), and COWIE (1961, fig. 2, p. 16, 40–41) have all commented on the remarkably flat erosion surface which cuts the Pre-Cambrian basement complex and which, where not exhumed or destroyed, is buried by the sedimentary rocks of the Pre-Cambrian Thule Group (COWIE, 1961, p. 17).

This paleoplain has been recognized in East Greenland (HALLER and KULP, 1962, p. 30; WENK, 1961, p. 22). It is found on the Bache Peninsula (WORDIE, 1938, p. 399; BENTHAM, 1936, p. 430–431) and is probably equivalent to that cut on the basement complex of the Canadian Shield, which was later buried by Proterozoic sedimentary rocks (AMBROSE, 1964, p. 825–830). This paleoplain has, therefore, great extent.

The writer was never able to reach an outcrop which contained the buried paleoplain but COWIE (1961, p. 13) found such an occurrence a few miles west of Etah. Here a coarse conglomerate overlies a red salic gneiss and, as might be expected, no weathering profile on the gneiss was reported.

Pre-glacial fluvial erosion has either modified or destroyed the paleoplain in most of Inglefield Land. It is probably preserved only where buried and in a narrow zone adjacent to the sedimentray rocks of the Thule Group where it has recently been exhumed.

COWIE's (1961, p. 40–41) analysis of the slope of the buried paleoplain and that of the present topography in southeastern Inglefield Land led him to conclude that here the paleoplain has been destroyed, presumably by pre-glacial fluvial erosion, as in places the present topography is hundreds of feet below the former position of the paleoplain (fig. 2).

Moreover, the valleys of the major rivers near the coast are cut in the Pre-Cambrian basement. In places this topography is hundreds of feet below the former position of the paleoplain; therefore the paleoplain has been destroyed in these areas.

In his discussion of the paleoplain, COWIE (1961, p. 42) states, "It is considered that the remarkably flat peneplain was produced by submarine erosion and is not subaerial in origin". He does not, however, present or discuss the data on which he bases this conclusion.

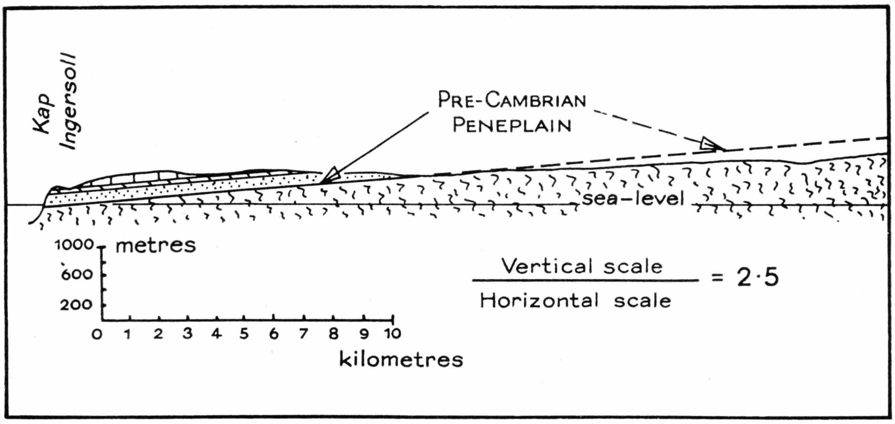

Fig. 2. Cross section of Inglefield Land, showing the area extending inland from Kap Ingersoll (by J. W. Cowie).

Ambrose (1964), who has made an interesting study of the voluminous literature on the exhumed pre-Paleozoic paleoplains of Canada, stresses the fact that when these surfaces were formed they had a drainage pattern that was adjusted to structure. They are, therefore, the result not of marine denudation but of subaerial erosion. The widespread distribution of the structurally controlled drainage pattern on the Canadian paleoplains suggests that the Inglefield Land paleoplain may also be due to fluvial erosion.

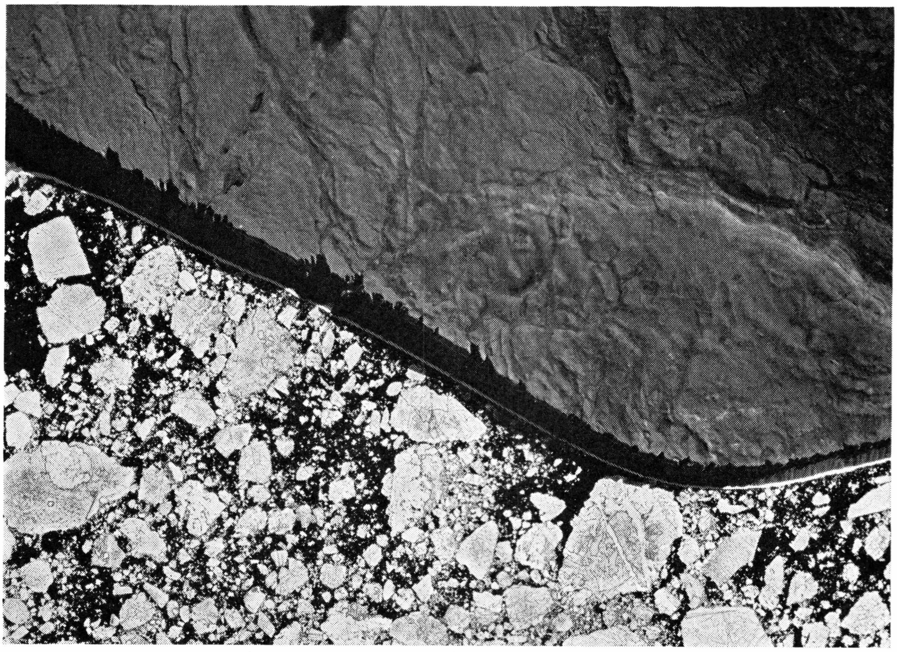

Fig. 3. An aerial photograph showing the coastal cliffs between the Minturn Elv and Bancroft Bugt. They are hundreds of feet high, cut in sedimentary rocks, and banked with talus. Vertical photograph by Geodætisk Institut, Copenhagen. Appr. scale 1:80.000.

TERTIARY PENEPLAIN

Steep cliffs are found along most of the coast of Inglefield Land (fig. 3). In places they are more than 500 feet high, and in general they cut sedimentary rocks. A plateau is found inland from the coastal cliffs. It rises gradually to the southeast, and near the Indlandsis it is more than 3000 feet high in many places. Near the coast the surface of the plateau cuts sedimentary rocks; farther inland it cuts the basement complex. The surface is essentially smooth where underlain by sedimentary rocks but somewhat less so where underlain by the basement complex. The major streams, most of which are fed by glacial meltwater, run southeast-northwest. Near the coast, valleys several hundred feet deep cut through the sedimentary rocks down into the basement complex. Structural terraces cut in the sedimentary rocks are common; so too are mesas which have been detached from the main body of the plateau (fig. 4). The plateau is an erosion surface. It is also an uplifted dissected peneplain because of its low relief and its elevated position with relation to sea level (Cowie, 1961, p. 44). At the coast the plateau peneplain is

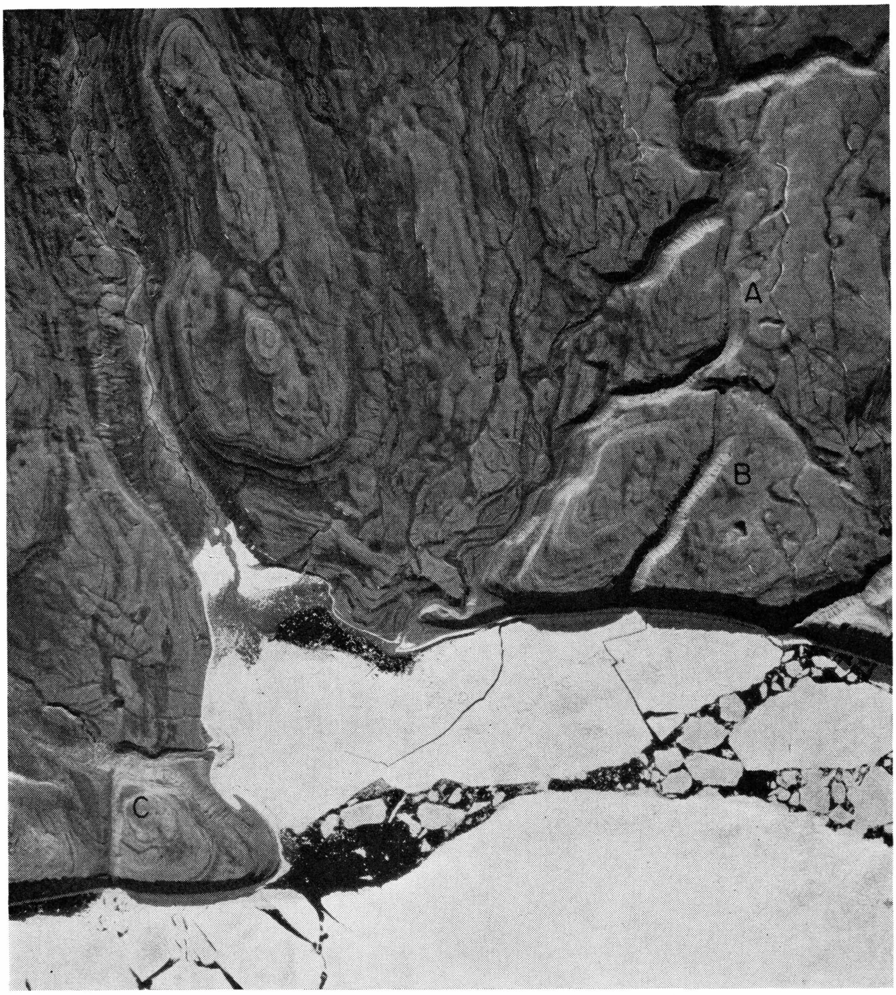

Fig. 4. An aerial photograph showing the area around Bancroft Bugt. Structural terraces, mesas, elevated beaches, marginal channels (A, B, C), and a dissected valley train are shown. Vertical photograph by Geodætisk Institut, Copenhagen. Appr. scale 1:65.000.

topographically higher than the older Pre-Cambrian paleoplain (Bentham, 1936, p. 428). The plateau peneplain, however, does not rise inland as rapidly as does the older surface; thus the two surfaces cross, and still farther inland the older surface has been destroyed (fig. 2). East of Advance Bugt, where the topography is lower, the sedimentary rocks have been removed, the basement complex crops out, and the Pre-Cambrian paleoplain has presumably been destroyed (Troelsen, 1950, Geol. map).

Little is known about the age of the plateau peneplain, but as most of the high level erosion surfaces in other parts of the world are Tertiary (ASHLEY, 1931, p. 537) it seems likely that this surface is also Tertiary. In the Thule area, DAVIES (1963, p. 11–12) has described a surface between 1000–1300 feet above sea level which he thinks is a Pliocene-Early Pleistocene erosion surface. Perhaps the plateau peneplain in Inglefield Land correlates with DAVIES' surface in the Thule area.

After the Inglefield peneplain was formed, it was uplifted several hundred feet and a period of fluvial erosion took place during which the valleys, cliffs, and structural terraces were formed. The plateau is everywhere veneered with Wisconsin glacial deposits, and Wisconsin glacio-fluvial deposits are found in the valleys. The peneplain, the uplift, and the erosion episode are therefore pre-Wisconsin. It is certain that there was not enough time to cut the valleys in the interglacial epochs. This is proved by the following: (1) glacial deposits are very much easier to erode than bedrock; (2) the original volumes of the glacial deposits in the valleys were small in comparison to the sizes of the bedrock valleys; (3) the streams have been unable in post-glacial time to remove all of the glacial material deposited in the valleys; (4) the duration of each interglacial epoch was roughly equivalent to the duration of the post-Wisconsin epoch; (5) the bedrock valleys could not be enlarged during the interglacial epochs until the glacial deposits in them were first removed. Therefore, the peneplain and the other major features of the landscape are pre-Pleistocene.

DROWNED COASTLINE

Rensselaer Bugt, Marshall Bugt, and Dallas Bugt, as well as other bays, are drowned valleys. Promontories, peninsulas, and islands similar to those which characterize drowned coasts are numerous between Dallas Bugt and the Humboldt Gletscher (fig. 1). Promontories and islands are also found south of Kap Inglefield. Although glacial erosion can accentuate initial coastal irregularities, it is unlikely that this coastline could have been developed without submergence.

The coastal cliffs, which in places are hundreds of feet high, are not due to glaciation, faulting, or marine action (fig. 3). They are too high to have resulted primarily from glacial erosion. The fact that they change alignment on the north side of Marshall Bugt as much as 90° within a few miles makes it unlikely that they are either fault scarps or fault line scarps. Moreover, faults close to the cliffs and more or less parallel to them have not been reported (Cowie, 1961, fig. 1, p. 35–36). Because the coastal cliffs are continuous with cliffs whose alignment along valleys proves them to be fluvial in origin, it is likely that they have had a similar origin. Following their formation, the submergence that was responsible for the promontories, islands, and drowned valleys brought the ocean to the foot of the cliffs. The subaerial topography which once extended out into Kane Bassin to the west of the cliffs and which was formed at the same time as the cliffs, was at this time drowned.

The amount of submergence necessary to form the coastal features is not known. An analysis of the topography around Rensselaer Bugt suggests that it may be a few hundred feet. It must be considerably less than 1000 feet, however, as hydrographic data indicate that Kane Bassin on the average is not as deep as this (U.S. Navy Dept. Hydrographic Office, 1963, pub. no. 6606).

It is not known for certain whether a delay in the isostatic rebound due to deglaciation, together with the submergence resulting from the weight of the existing ice cap, can account for the submergence, or whether tectonic movements are also necessary.

The Indlandsis is approximately 2000 feet thick 40 miles east of Kap Alexander (fig. 1) and approximately 4000 feet thick 140 miles east of Kap Alexander, and it increases in thickness to more than 6000 feet still farther eastward.[1]) The elastic and isostatic uplift consequent upon the complete deglaciation of all of the Indlandsis, together with the isostatic uplift which is still to take place because of existing deglaciation, would be many hundreds of feet[2]) (FLINT, 1957, p. 244). The eustatic rise of sea level consequent on this deglaciation and on simultaneous deglaciation in other parts of the world would be approximately 200 feet (THIEL, 1962, p. 175). It seems likely, therefore, that under these conditions most if not all of Kane Bassin would be dry land on the completion of isostatic rebound.

[1]) Personal communication from Dr. J. J. HOLTZSCHERER.

[2]) Personal communication from Dr. ROBERT MCCONNELL.

ORIGIN OF KANE BASSIN

Smith Sund, Kane Bassin, Kennedy Kanal, Hall Bassin, and Robeson Kanal (fig. 5) could have been formed by the submergence of subaerial topography or by the down-faulting of these areas (Koch, 1928a, p. 505–506). In any event, glacial and marine activity have modified them somewhat. No data known to the writer indicate that these basins and channels are down-faulted areas.

Cooke (1929, p. 91–120; 1931, p. 169–180) concluded that following the uplift of the Canadian shield at the close of the Cretaceous and probably during the Eocene, a long period of erosion took place which by the Pliocene had reduced the shield to a nearly perfect peneplain. Following peneplanation, the shield was uplifted and streams cut deep valleys in places. Glaciation was the next event; during the Pleistocene the valleys were enlarged and the shield was depressed about 1200 feet. Since glaciation, the shield has risen on the average about 600 feet (Cooke, 1930, p. 86–87). The coastline of Labrador—where, among other areas, this geologic history took place, according to Cooke—is much like that of western Greenland. Both coastlines are characterized by features indicating recent submergence. If these two areas have had a similar erosional geologic history in the Cenozoic (and they have had if the Plateau peneplain in Inglefield Land is Pliocene in age), then the channels and basins between Ellesmere Island and North Greenland may have been formed by the submergence of an uplifted dissected peneplain. Wilson (1963, p. 86–100) has postulated that a fault runs through Robeson Kanal, Hall Bassin, Kennedy Kanal, Kane Bassin, and Smith Sund, and on south into Baffin Bugt. If real, this fault may have guided erosion and helped in the formation of these channels and basins.

An analysis of the meagre hydrographic data now available for Smith Sund, Kane Bassin, Kennedy Kanal, Hall Bassin, and Robeson Kanal (U.S. Navy Dept., Hydrographic Office, 1963, pub. no. 6606) suggests that Kane Bassin is the shallowest of these water bodies and that they get progressively deeper going north to the Arctic Ocean and south to Baffin Bugt. Perhaps before submergence there were two river systems, one flowing south from a divide in Kane Bassin to Baffin Bugt, the other flowing north. The fact that the divide between these two hypothetical

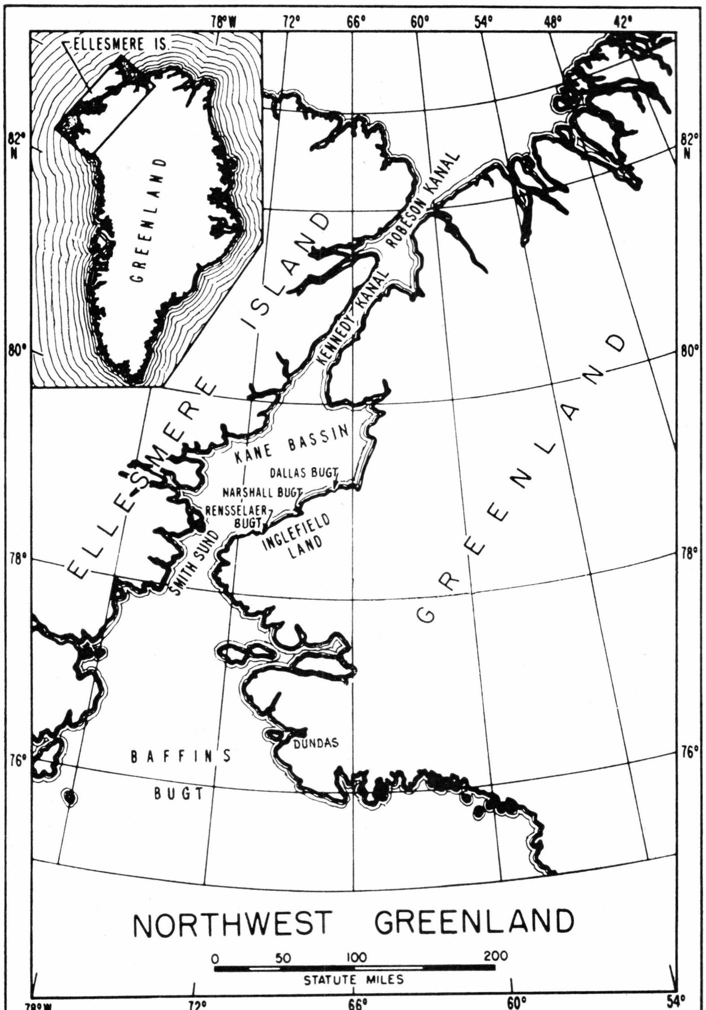

Fig. 5. Map showing the channels, basins, and sounds between Greenland and Ellesmere Island.

drainage systems formed, on submergence, a larger water body (Kane Bassin) than areas farther downstream (Smith Sund and Robeson Kanal) has no ready explanation.

Whether or not these channels and basins were formed in this way, the drowned valleys, promontories, and islands along the Inglefield coast indicate that Kane Bassin was smaller and shallower before the submergence necessary for their formation.

ELEVATED BEACHES

Several kinds of evidence have been used to prove that sea level in post-glacial time has been higher than at present in Greenland and other northern areas (Flint, 1948, p. 164–167). They are: (1) uplifted driftwood (Koch, 1928a, p. 517; Fristrup, 1952, p. 91; Washburn, 1965, table 5); (2) uplifted whalebone (Krinsley, 1963, p. 58–59); (3) uplifted marine shells (Krinsley, 1963, p. 57; Laursen, 1950, p. 1–143); (4) uplifted marine clay (Koch, 1928a, fig. 12; Laursen, 1950, 1954); (5) uplifted marine till (Krinsley, 1963, p. 57); (6) uplifted sea ice-pushed features (Koch, 1928a, p. 517; Nichols, 1953a); (7) uplifted deltas (Krinsley, 1961, p. 749; Pessl, 1962, p. 73–76); (8) uplifted wave-cut cliffs, benches, terraces (Bretz, 1935, table I; Flint, 1948, table I); (9) uplifted wave-washed surfaces; (10) uplifted beaches (Bretz, 1935, p. 204–222; Washburn, 1947, p. 64–68; Flint, 1948, p. 162–192); (11) fluvial terraces (Compton, 1964, p. 279–280); (12) lower limit of perched boulders (Blackadar, 1956, p. 22); (13) uplifted etched bands (Flint, 1948, p. 164, 190).

Elevated beaches, deltas, marine shells, and wave-washed surfaces in Inglefield Land prove that sea level in post-glacial time has been between 200–300 feet higher than at present. These features will now be discussed.

More than a score of elevated beaches extending from approximately 4 miles southwest of Force Bugt (ca. lat. 78°31′+, long. 72°15′) to Dallas Bugt (ca. lat. 79°03′, long. 67°55′) were investigated (table 1; fig. 6). The largest is a mile or so in length and a few hundred yards wide. The highest, which is 4 miles southwest of Force Bugt, is 285 feet above sea level (table 1; fig. 7).

To form well-developed beaches there must be adequate source material, the submarine and subaerial coastal topography must be neither too steep nor too flat, the fetch and exposure must be favorable, the sea ice must go out for a few months nearly every year, and during the open water stage there must be waves. Because these conditions are not found everywhere along the coast, the elevated beaches are not continuous (fig. 8). These conditions are found almost without exception in the drowned valleys where deltaic sediments are common. They are not generally

Fig. 6. Raised beaches up to 247 feet above sea level on southwest side of Rensselaer Bugt. Ridges and swales are well developed, frost furrows are common, and a solifluction sheet buries the beaches at the upper left hand side of photo. Stream has cut meander scars, cusps, and terraces. Oblique photo by U.S. Army.

Table 1. *Elevated beaches of Inglefield Land, Greenland (altitudes determined by altimeter)*

Location	Max. Altitude	Area Dimensions	Features	Remarks
4 miles southwest of Force Bugt and 2 miles east of Kap Inglefield at ca. lat. 78°30′+, long. 72°15′.	285′	Extensive.	Best ridge and swale topography of any beach seen in Inglefield Land.	Solifluction lobe and bedrock immediately above highest ridge. River immediately west of beach. Elevated wave-washed surfaces immediately northeast. Good beaches between this one and Force Bugt.
Southwest side of Rensselaer Bugt at ca. lat. 78°37′, long. 71°07′.	247′	Extensive; 1 mile long.	Ice wedge furrows and frost mounds present. In part buried by solifluction sheets. Excellent ridge and swale topography	Most extensive and best developed elevated beach in Inglefield Land. Derived from till and deltaic deposits.

(continued)

Table 1 (continued)

Location	Max. Altitude	Area Dimensions	Features	Remarks
East side of Rensselaer Bugt at ca. lat. 78°38′, long. 71°02′.	275′			Cuspate beaches; swale ponds; in pre-glacial valley; good beach ridges.
East side of Rensselaer Bugt; at mouth of Tufts Elv.	200′+		Beach ridge.	On delta.
Northeast side of Bancroft Bugt at ca. lat. 78°45′, long. 70°15′.	270′	Extensive.		Good beach ridges and swales. Easy to pick marine limit. Excellent ridges and swales on south side of Bancroft Bugt.
Approx. 2 miles northwest of mouth of Minturn Elv at ca. lat. 78°50′, long. 69°45′.	250′	Extensive.		Considerable thickness of beach material. Wave-washed surface above highest ridge. Well-developed ridges and swales. Other beaches at 160′, 110′ etc. Delta with fore-set beds dipping 20° and containing shells is close by.
Several beaches between mouth of Minturn Elv and Marshall Bugt.	240′ 145′ 130′			Not highest marine limit. Land did not go higher.
In valley immediately west of Marshall Bugt and south of Inuarfigssuaq.	240′	Small.	Beach ridge.	On delta.
Southwest side of Dallas Bugt at ca. lat. 79°03′, long. 67°55′.	210′		Faint ice-pushed ridges. Irregular topography – beach sands and gravels were deposited on and around ice.	Beach ridges developed on delta.

found along the talus-veneered bedrock cliffs between the drowned valleys, where the offshore profile is presumably steep, except where small streams that have incised themselves into the plateau have built small deltas out into Kane Bassin.

Fig. 7. Well-developed raised beaches up to 285 feet above sea level four miles southwest of Force Bugt. Solifluction sheets bury highest beaches. Oblique photo by CHRISTOPHER G. MILLER.

Many of the beaches are characterized by excellently developed ridges and swales (fig. 7), the ridges being commonly composed of coarser material than the swales. Small ponds are located in the swales of beaches on the southwest side of Rensselaer Bugt. No shells were found on any of the beaches, presumably because they are composed of well-rounded gravels. As might be expected, these beaches are characterized by features developed only under polar conditions (NICHOLS, 1961, p. 694–708). In places they are buried by solifluction lobes and sheets. Frost furrows with and without marginal levees are found on almost all of the raised beaches, and nonsorted circles and polygons are found on some of them. Faint ice-pushed ridges (NICHOLS, 1953a) occur, as well as beach gravels with a knob and kettle topography formed because the gravels were initially deposited by wave action on or around ice.

In places, well-developed ridge and swale topography extends continuously from the marine limit down to the modern strandline, indicating

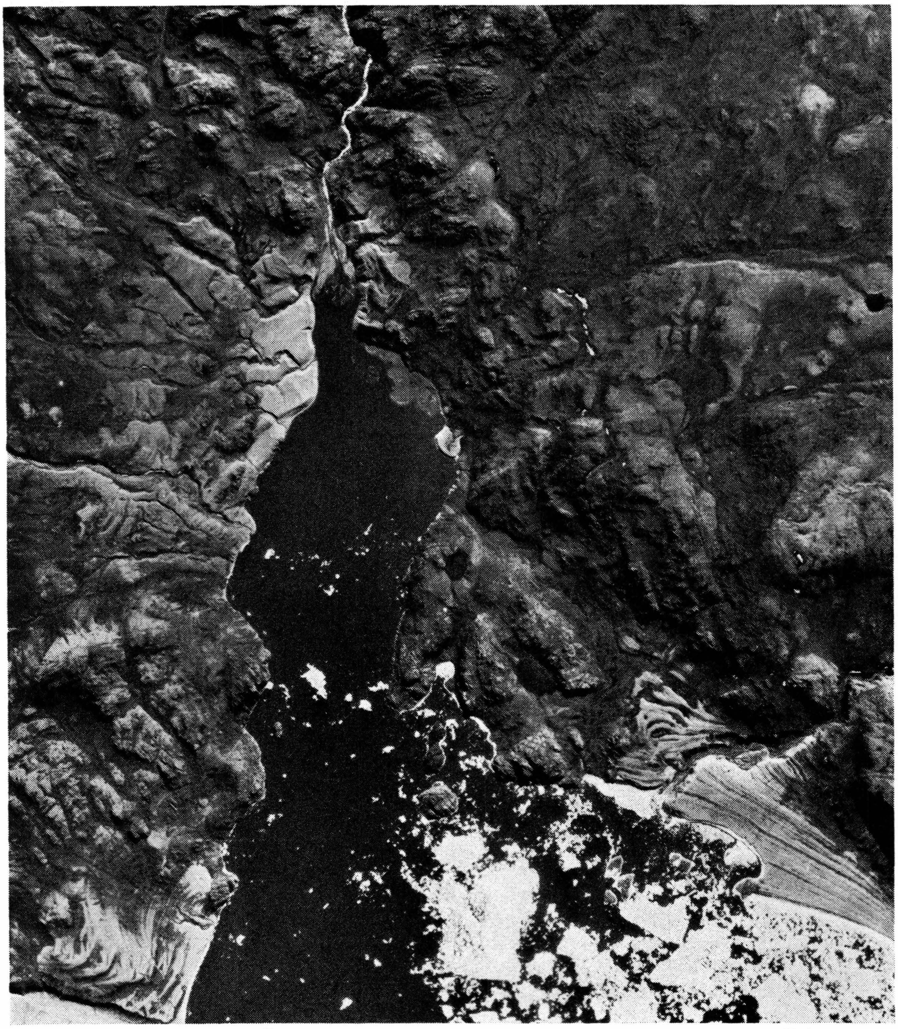

Fig. 8. Raised beaches (B) and uplifted dissected delta (D) at Rensselaer Bugt, Inglefield Land, Northwest Greenland. There is no evidence of a significant stillstand as sea level dropped with respect to the land from the marine limit down to the modern strandline, for the beaches extend in a gentle, continuous slope from the marine limit down to the modern strandline. Vertical photo by Geodætisk Institut, Copenhagen. Appr. scale 1:45.000.

that the open water conditions which now prevail during the summer months have existed since the highest beaches were formed.

The elevated beach ridges on the east side of Rensselaer Bugt about 3 miles south of Kap Leiper and immediately south of the nearby river are roughly parallel to the modern strandline. On approaching the valley

Fig. 9. Raised beaches on southwest side of Rensselaer Bugt. The irregularity of the beaches at upper left of photograph is due to the irregular bedrock topography which they veneer. Ponds are found in the swales and marine cliffs along the modern strandline. Vertical photo by Geodætisk Institut, Copenhagen. Appr. scale 1:28.000.

from the south, however, the ridges slowly change direction so that in the valley they are essentially perpendicular to the strandline and parallel to the axis of the valley. The valley is therefore older than the beaches.

The elevated beaches on the southwest side of Rensselaer Bugt, as well as those found elsewhere, extend continuously from the marine limit down to the modern strandline without significant change in slope. This indicates that the land was continuously elevated relative to the sea during the formation of the beaches and that no significant stillstand between land and sea took place at that time (fig. 9). Moreover, in East Greenland NOE-NYGAARD (1932, p. 8, 17) found a series of continuous raised beaches extending from sea level up to 174 feet which indicates no significant marine stillstand for this area. Nor did BRETZ (1935, p. 220) find evidence for any significant stillstand in East Greenland. He concluded that the uplift had been essentially continuous.

Till and deltaic deposits are important sources for the beach deposits. The importance of the source material in the development of the beaches is well shown by the beach on the southwest side of Rensselaer Bugt. Here the upper part of the beach, which is poorly developed and derived from till, is characterized by an abundance of boulders, an absence of good ridge and swale topography, and a scarcity of beach deposits. The lower part of the beach, which is excellently developed and, in part at least, derived from deltaic deposits, is characterized by an absence of boulders, excellent ridges and swales, and an abundance of beach deposits. A given volume of till, after being reworked by waves and shore currents, yields a much smaller volume of beach material than does an equal volume of deltaic sediments. Till is, therefore, a poor source for beaches; deltaic deposits are an excellent source.

Because the raised beaches are derived from till and from deltaic deposits that bury till, and because there are no features indicating that they were overrun by the Indlandsis, they must have been formed after the area was deglaciated.

Limestone roundstones are very common on some of the raised beaches. As much as an inch of material has been removed from some of the roundstones by solution. There is little rainfall in Inglefield Land, and much of the snow cover is probably lost by evaporation (Fristrup, 1952–53, p. 55). Solution, therefore, must be a slow process, and these features indicate that the beaches have considerable antiquity (Nichols, 1953 b, p. 269–270).

Fig. 10. The head of the delta in Rensselaer Bugt. A dissected valley train extends approximately 12 miles upstream from the delta. Glaciated Pre-Cambrian rocks with numerous undrained bedrock depressions extend inland for several miles. Oblique photo by U.S. Army.

ELEVATED DISSECTED DELTAS

Uplifted dissected deltas are found in northern Greenland (Bretz, 1935, p. 362; Flint, 1948, p. 332; Troelsen, 1949, p. 20; Krinsley, 1961, p. 749; Washburn and Stuiver, 1962, p. 66–73; Pessl, 1962, p. 73–76; Washburn, 1965, p. 30–31; Sugden and John, 1965, p. 235–247).

Large uplifted dissected deltas are found at Force Bugt, Rensselaer Bugt, Bancroft Bugt, Marshall Bugt, and Dallas Bugt and the lower end of the Minturn Elv. Small uplifted dissected deltas are found in many other places (table 2). The larger deltas have been fed ever since their inception by sediments carried in glacial meltwater streams, whereas the small deltas have been cut off from glacial meltwater at various times during the deglaciation of Inglefield Land.

The delta in Rensselaer Bugt is one of the largest, is the most instructive, and was the most carefully studied (fig. 10). It has been so greatly dissected that it consists only of terrace remnants. There are at least five prominent non-paired terraces and many minor ones (figs. 10, 11). The highest is nearly 200 feet above river level. Because there was

Table 2. *Location of deltas in Inglefield Land, Greenland*

Location	Feature	Altitude	Remarks
Lowest terrace on southwest side of Rensselaer Bugt.	Fore-set beds seen in terrace slope.	At and close to sea level.	
Southwest side of Resselaer Bugt; seen in ravine cut by river 1–2 miles from mouth of Tufts Elv.	Fore-set beds.	Close to sea level.	Considerable thickness of sand and gravel.
Rensselaer Bugt, east side.	Fore-set beds; shells found in place.	0–46 feet	Numerous and excellent examples of fore-set beds; exposed through a vertical distance of 46 feet; maximum dip of $26^1/_2°$; individual beds can be traced for hundreds of feet; interbedded peat; associated shells.
Southwest of Bancroft Bugt; at mouth of river immediately southwest of Bancroft Bugt.	Thick fill of alluvium 150–200 feet thick.		
On irregular peninsula on west side of bay into which the Minturn Elv empties.	Fore-set beds.		Abundant shells; 20° dip; flat initial deltaic surface present; marine cliff cut in delta.
Along Minturn Elv between mouth of bedrock gorge and last tributary to enter it from east; $^1/_4$ mile downstream from entrance to bedrock gorge on east side of river.	Fore-set beds; shells washed out from bottom-set beds.		Fore-set beds seen in terrace slope immediately above river; more than 5 non-paired terraces cut in delta; delta more than 150 feet thick; abandoned channels, oxbows, meander scars, and cusps associated with terraces.
Along river immediately south of Inuarfigssuaq.	Lip present; beach-ridged delta; two small beaches on delta flat near lip; original deltaic slope present; ravine cut into delta.	Lip 240 feet above sea level.	Marine limit approximately 240 feet above sea level; glacial meltwater was diverted from this valley early in the deglaciation of Inglefield Land.

(continued)

Fig. 11. Several non-paired terraces cut in the delta at the head of Rensselaer Bugt. The highest deltaic deposits are nearly 200 feet above river level. Photo by U.S. Army.

Table 2 (continued)

Location	Feature	Altitude	Remarks
Marshall Bugt, immediately upstream from head of bay.	Bottom-set silty beds, containing shells, on west side of river; several excellent terraces.	Highest terrace 160 feet above river level.	Thickness of fill probably more than 200 feet.
Southwest side of Dallas Bugt; near mouth of largest river to reach Dallas Bugt from southwest side of bay.	Beach-ridged delta; many boulders on delta flat–similar to those on modern deltas at Rensselaer Bugt, etc.	Highest beach 210 feet above sea level.	Large delta; shells in place at 162 feet above sea level are found nearby.
Southwest side of Dallas Bugt; beneath southeast end of cliff.	Small delta; lip present; ravine cut into delta.	Lip 212 feet above sea level; a series of beach ridges below toe of delta.	Before beach ridge immediately below toe of delta was formed, growth of delta stopped because glacial meltwater carrying sizable quantities of sediments no longer reached it, due to deglaciation.

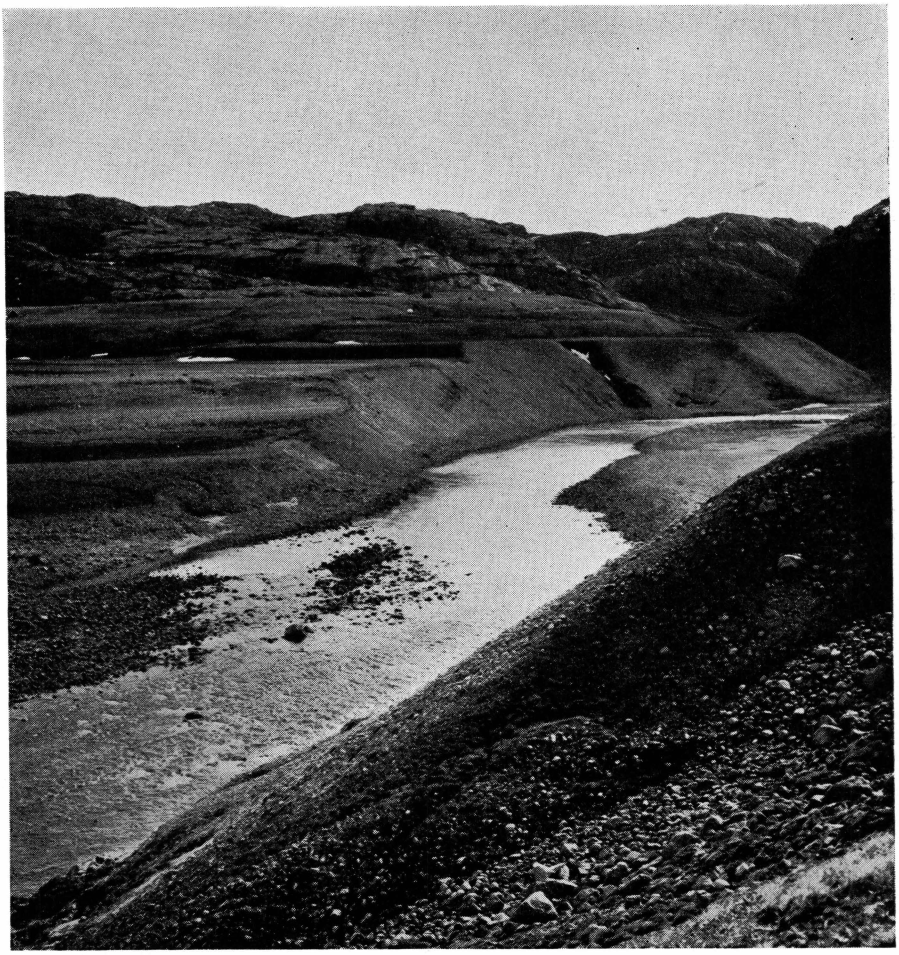

Fig. 12. Non-paired fluvial terraces cut in deltaic deposits near the mouth of the Minturn River. Abandoned channels, oxbows, meander scars, and cusps are present. Photo by Christopher G. Miller.

only time for a reconnaissance study of the delta, no data were obtained on the precise elevations of the terraces. Non-paired terraces are also found at Force Bugt, Bancroft Bugt, and Marshall Bugt and at Minturn Elv (fig. 12). The absence of a stillstand between land and sea, as indicated by the elevated beaches, suggests that the delta progressively extended itself into the bay during the continuous drop of sea level relative to land.

Fore-set beds are found at several places (table 2) around Rensselaer Bugt. At one locality they crop out continuously through a vertical distance of about 46 feet; here individual beds can be traced for hundreds of feet (fig. 13). Dips ranging from 18 to 26 degrees were measured. The deltaic deposits certainly extend below sea level. A conservative

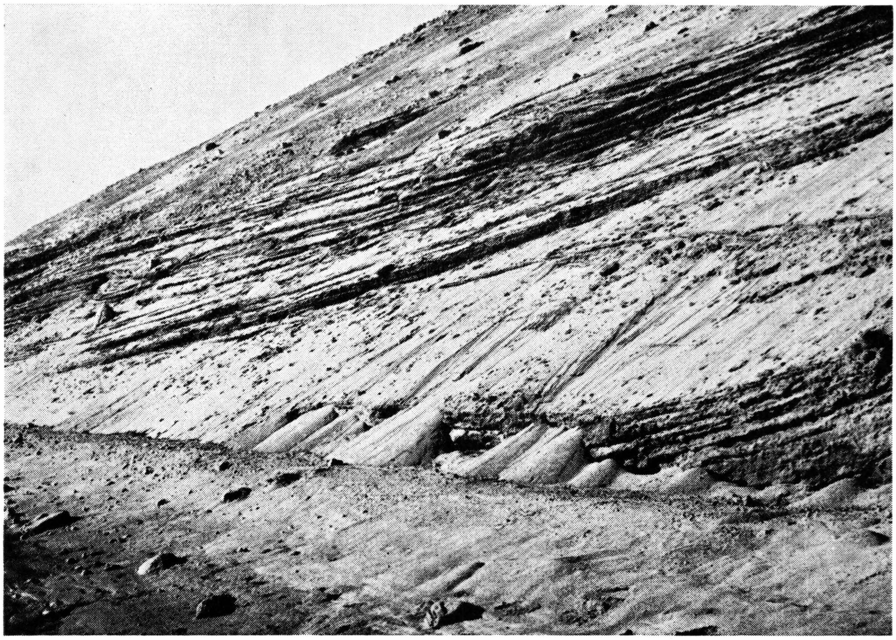

Fig. 13. Fore-set beds in the Rensselaer Bugt delta. They crop out through a vertical distance of about 45 feet. Individual beds can be traced for hundreds of feet. The delta consists mainly of sand with occasional boulders scattered through it.

estimate, based on a reconstruction of the bedrock valley, suggests that they may extend 100 or more feet below it. It seems likely that 200 feet is a minimum thickness for the delta at Rensselaer Bugt; and a similar thickness, based on the height of the terraces and the width of the valleys, is indicated for the deltas at Marshall Bugt and Bancroft Bugt and at the mouth of the Minturn Elv.

The Rensselaer Bugt delta consists mainly of sand, although boulders are scattered through it (fig. 13). The sand at the surface of a terrace about 55 feet above sea level is stained brown down to permafrost.

Many of the modern deltas between Inglefield Land and Thule have a beach ridge at their outer edge. An elevated beach ridge also occurs on the highest terrace of the Rensselaer delta on the east side of the Tufts Elv near the entrance to the gorge; and beach ridges are found on other elevated deltas in Inglefield Land (table 2).

Many layers of silty brown peat a few inches thick are interbedded in the fore-set beds of the delta at Rensselaer Bugt. The vegetation of which the peat is composed grew in wet places on land[1]) and was transported to the delta by the Tufts Elv. The brown color of the peat indi-

[1]) Personal communication from Professor E. S. BARGHOORN of Harvard University.

cates chemical alteration of the original vegetation and antiquity for these deposits. A radiocarbon dating by the Lamont Geological Observatory (L-1091 A, Lamont) shows that the peat is 7800 ± 200 years old.[1])

Abundant shells are found approximately 10 feet above sea level in silts associated with the fore-set beds of the Rensselaer Bugt delta. Single valves are the most common, but some articulated valves are present. Shells are found in many other places around Rensselaer Bugt, where they have been reworked by both running water and wind. *Mya truncata* L., *Serripes groenlandica* Brug., *Hiatella arctica* L., *Macoma calcarea* Gmel., and *Astarte banksi* Leach have been identified.[2]) A radiocarbon dating of the shells in the delta by the Lamont Geological Observatory (L-1091 E, Lamont) indicates that they are 7800 ± 200 years old.[3]) Shells which have been dated have also been found in several other localities (table 3).

The Tufts Elv, beginning from where it emerges from the bedrock gorge upstream from Rensselaer Bugt, has cut a valley with a broad flood plain into the delta. The flood plain is covered with large boulders, and the river is braided. Extending for approximately 2 miles up the gorge and presumably for its total length, there is a discontinuous series of alluvial terraces which extend above the river at the bottom of the gorge about as much as the top of the delta does above the modern flood plain that has been cut in it. Downstream, the flood plain merges into a modern delta built out into Rensselaer Bugt. The surface of the modern delta is flatter than the flood plain; and the fact that this delta is intertidal proves that it has been formed while sea level has been at essentially its present position. Beyond the lip of the delta, which can be easily seen from the air, there is deep water which during the summer months is commonly dotted with bergs as well as pack ice. Similar modern deltas are found at the mouths of the other glacier-fed streams in Inglefield Land and are common in East Greenland (Bretz, 1935, p. 362; Flint, 1948, p. 332; Sugden and John, 1965, p. 244–245).

Large boulders are scattered over the surface of the modern delta. They appear to be quite angular and some are several feet long (fig. 14). There was no opportunity to make a detailed study of them, as a boat was not available. Their large size and the fact that they are surrounded by sand and silt necessitate an explanation of how they reached the delta.

It seems certain that they rest on the surface of the delta and are not protruding from a till plain immediately beneath it.

[1]) Personal communication from Dr. David Thurber of Columbia University, New York City.

[2]) Personal communication from Dr. William Clench of Harvard University.

[3]) Personal communication from Dr. David Thurber of Columbia University, New York City.

Table 3. *Location of shells in Inglefield Land, Greenland*

Location	Species	Elevation	Age	Remarks
Force Bugt				A few shells.
Rensselaer Bugt; east side; in clifflet adjacent to river.	*Mya truncata* L. *Serripes groenlandica* BRUG. *Hiatella arctica* L. *Macoma calcarea* GMEL. *Astarte banksi* LEACH.	5–10 feet above sea level.	7800 ± 200 years old (L-1091 E)[1]	In place; in silt associated with fore-set beds of delta; numerous.
Bancroft Bugt.				Not found in place; 60 feet above sea level; on modern flood plain.
Along Minturn Elv; west side of river; at entrance to bedrock gorge.		Most were found at ca. 35 feet above sea level; highest shells between 50-75 feet above sea level.	7500 ± 250 years old (L-1091 C)[1]	Numerous; not in place; presumably washed out by the river from a solifluction lobe composed of deltaic silts.
Marshall Bugt Elv; west side of river; below well-developed terraces; few hundred yards from head of bay.		10–20 feet above sea level.	6300 ± 150 years old (L-1091 D)[1]	Found in fine-grained sediments; numerous.
Head of Dallas Bugt; southwest side.		Approximately 160 feet above sea level.	5900 ± 150 years old (L-1091 B)[1]	In fine-grained sediments; in place; numerous.
Near mouth of river between Dallas Bugt and Advance Bugt.				

[1]) Radiocarbon dating by Lamont Geological Observatory. Personal communication from Dr. DAVID THURBER.

Fig. 14. The modern delta in Rensselaer Bugt. It is covered with boulders several feet in length which extend out to the lip of the delta. The small bergs are in the deep water beyond the lip. The delta indicates a stillstand of unknown duration. Oblique photo by U.S. Army.

Gravel and large boulders cover the flood plain cut in the uplifted dissected delta immediately upstream from the modern delta. The boulders on the modern delta may have been transported from the flood plain by the Tufts Elv, particularly when it was in flood and when aided by the ebbing tide. This would have been facilitated if the boulders: (1) were floated on top of slabs of river and/or sea ice; (2) were carried while frozen to the bottoms of ice slabs; (3) were pushed by ice slabs from the flood plain onto the modern delta, the slabs of ice being so large that they could easily move the boulders; (4) were transported in channels cut in river and/or sea ice. Boulders transported by the river would be somewhat rounded. If there are any boulders on the modern delta that are composed of rock types not found in the drainage basin of the Tufts Elv, they did not come from the flood plain. Cliffs hundreds of feet high and steep bedrock slopes are found along many parts of the coast; perhaps fragments roll from these cliffs and slopes across the ice foot onto the sea ice and are later rafted onto the modern delta of the Tufts Elv during very high tides. The boulders could not have been depo-

Fig. 15. Small knobs and kettles in fluvial gravels deposited on sea and/or river ice are found near the mouth of the Minturn Elv. The relief is as much as 5 feet, and the gravel-buried ice is only a foot or so below the surface.

sited on the delta by large icebergs, however, as the water is not deep enough. Modern deltas with boulders scattered over them are also found at Marshall Bugt, at the mouth of the Minturn Elv, and elsewhere.

Pitted alluvium is found on the flood plain cut in the dissected delta of the Minturn Elv, upstream from its modern delta. The alluvium has a relief of as much as 5 feet and is characterized by small knobs and kettles composed of sand and gravel about a foot thick beneath which there is ice (fig. 15). The gravels were without doubt deposited on the ice by the Minturn Elv, and the knobs and kettles are due to the differential melting of the ice. Boulders several feet long are scattered over the gravels (fig. 16). The slight rounding of the boulders and the fact that they are probably too high topographically to have been transported by offshore sea ice, together with their association with river-deposited sands and gravels, make it likely that they too were transported and

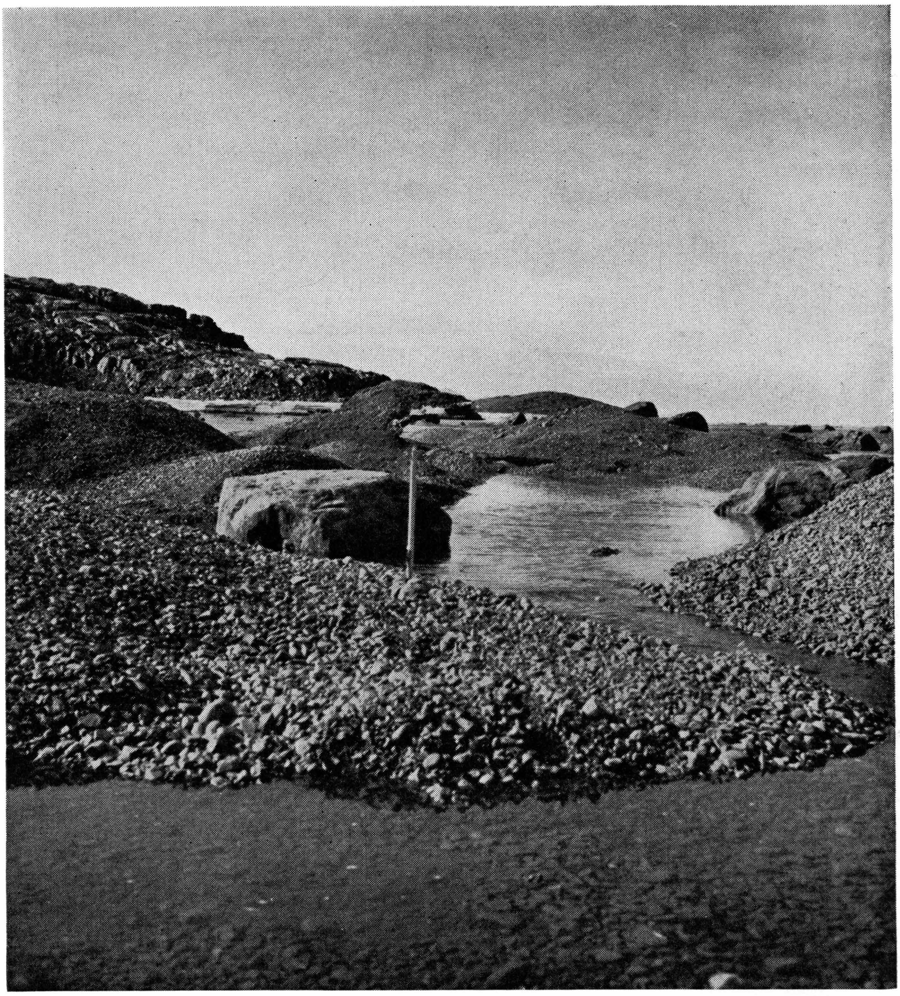

Fig. 16. Subrounded boulders several feet in length are found on the gravels.

deposited by the Minturn Elv. This suggests that perhaps the boulders on the modern deltas in Rensselaer Bugt and elsewhere are also stream-transported and deposited.

It is unlikely that the bottom-set beds of the dissected Rensselaer Bugt delta are being eroded by the Tufts Elv to any considerable degree. This is indicated by the fact that the river is clear and therefore does not carry any significant quantity of silt or clay. The Marshall Bugt elv, however, unlike all the other large glacier-fed streams seen by the writer in Inglefield Land, is dirty, yellow, and unfit for cooking and drinking. It discolors Marshall Bugt. It is not known whether the sediments responsible for the discoloration are derived from (1) yellow sedimentary rocks,

(2) a weathering profile developed on the Pre-Cambrian paleoplain, (3) weathered glacial drift, (4) gossans in the basement complex, or (5) the bottom-set beds of the delta. The bottom-set beds of the delta are a likely source, as is suggested by the presence of several small patches of fine-grained material on the west side of the river a few tens of feet above it and near its mouth. A short traverse along the river might solve the problem. If the source of the yellow sediments carried by the river is the bottom-set beds of the delta, it would be of interest to know whether they were originally yellow, having been derived from older yellow material, or whether they were originally gray and were weathered yellow. It would seem that if the discoloration of the river is due to the erosion of bottom-set beds they must have been originally yellow, because the weathering necessary to change considerable quantities of gray sediments into yellow would have taken time and would probably have had to take place under subaerial conditions. It is difficult to see how the weathering would have had time to take place, under the existing hydrologic and topographic conditions, before the sediments were removed by the river. An answer to this problem may have significance with regard to multiple glaciation.

The highest deltaic sediments are perhaps 50 feet lower than the highest raised beaches in the Rensselaer Bugt area. The raised beaches may have been formed before the delta was developed because an ice tongue afloat in Rensselaer Bugt allowed the highest beaches to form but prevented the early development of the delta. Ice tongues afloat are now present in both Washington Land and Peary Land, North Greenland (KOCH, 1926, p. 99; 1928a, p. 513–514). If the isostatic rebound in Rensselaer Bugt was as rapid at this time as it was initially at Mesters Vig, Northeast Greenland (WASHBURN, 1965, p. 31–37), the ice tongue would have had to persist for only a few centuries. The difference in elevation may, however, be due to the fact that the upper part of the delta has been removed by the Tufts Elv whereas the highest beaches have not been destroyed.

ORIGIN OF TERRACES

It seems certain that most of the terraces cut in the Tufts Elv delta and in other deltas of Inglefield Land have not resulted from the development of a succession of lower and lower deltas as sea level progressively fell with respect to the land in post-glacial time. The erosional origin of terrace risers and treads demonstrates this. Only the tread of the highest terrace, on the east side of the Tufts Elv close to the entrance of the bedrock gorge, is thought to be an original deltaic surface. On this surface there is a small ridge a few feet high which the writer thinks is a beach ridge, and the presence of beach ridges on modern deltas suggests that this surface is deltaic. However, top-set beds are not found on many of the terrace treads and indeed fore-set beds are truncated by them. In addition, several terrace risers are known to be erosional because truncated fore-set beds crop out on them.

It is also certain that all or nearly all of the terraces were formed by the Tufts Elv and its tributaries rather than by marine action. This is proved by the following: (1) The terrace risers and treads in general are parallel to the Tufts Elv and its tributaries and not perpendicular to wave propagation. (2) The Tufts Elv has built an extensive delta during the present stand of sea level, whereas marine erosion and deposition have been insignificant in the Rensselaer Bugt area during this period. It seems likely, therefore, that during all of post-glacial time fluvial processes in Rensselaer Bugt have been quantitatively more important than marine processes. (3) Some terrace risers face inland, where wave attack could not have formed them. (4) Lower terraces are found farther inland than some higher terraces, which would not be expected if marine processes were solely responsible. (5) The terraces are non-paired. (6) Oxbows, abandoned channels, meander scars, and cusps are found on one of the terraces of the Minturn Elv delta. (6) The large non-paired fluvial terraces cut in the valley trains upstream from the deltas suggest that the terraces cut in the deltas are also fluvial.

It is not thought that the terraces indicate a succession of marine stillstands. The continuous slope of the beaches from the marine limit down to the modern strandline precludes this.

FLINT (1948, p. 90, 186) has described a dissected delta in Kempe Fjord in Northeast Greenland. A series of terraces is found along a trench cut in the delta. He reported that the terraces were all fluvial. This delta and its terraces appear, therefore, to be similar to the dissected deltas in Inglefield Land. SUGDEN and JOHN (1965, p. 235–240) have also described terraces cut in an uplifted dissected delta in Kjove Land, East Greenland. These workers, however, believe that the terraces are marine in origin and were formed during a succession of marine stillstands.

MARINE LIMIT

The highest beaches and deltas progressively decrease in altitude from 4 miles southwest of Force Bugt (285 feet) northeastward to Dallas Bugt (210 feet) (tables 1, 2). This suggests that the ice removed during deglaciation progressively decreased in thickness from the southwest to the northeast and/or that the deglaciation to the northeast is more recent (table 3). The writer has not done field work east of Dallas Bugt. No well-developed raised beaches are shown on the aerial photographs of this area. It is characterized by Pre-Cambrian igneous and metamorphic rocks. Bedrock crops out abundantly and mantle rock is presumably thin. The absence of well-developed raised beaches may be due to the scarcity of source materials and/or to the absence of open water during the summer months. The problem merits investigation. Perhaps the most rewarding areas for study would be at the mouths of the rivers, for here elevated and dissected deltas might be present.

Table 4. *Elevated beaches and marine terraces of West Greenland*

Location	Altitude	Remarks	Observer or Authority
Søndre Strømfjord (lat. 66°–67°)	886 feet	Terraces up to 886 feet; not proven marine.	LAURSEN, 1950, p. 101. For areas between 67°30′ and 71°00′ see also LAURSEN, 1944, p. 90–103.
Qasortoq (lat. approx. 69°)	607 feet	Upper marine limit.	LAURSEN, 1950, p. 125
Atâ (lat. 69°45′)	About 656 feet	Upper marine limit.	LAURSEN, 1950, p. 125.
Vaigat (lat. approx. 70°)	656 feet	Upper marine limit.	LAURSEN, 1950, p. 125.
Ũmánaq Fjord (lat. approx. 71°)	755 feet	Upper marine limit.	LAURSEN, 1950, p. 125.
Kap York district between Etah and Melville Bugt.	164–180 feet	The most typical shoreline.	KOCH, 1928a, p. 517–518.

(continued)

Table 4 (continued)

Location	Altitude	Remarks	Observer or Authority
Saunders Ø (ca. lat. 76°35′, long. 70°)	125 feet	Highest preserved marine feature.	KRINSLEY, 1963, pl. 4, p. 59
Saunders Ø	150 feet	A solifluction sheet is found immediately above this beach.	BENDIX-ALMGREEN, FRISTRUP, NICHOLS, 1967, p. 16
North Star Bugt (near Thule)	125 feet	Highest preserved marine feature.	KRINSLEY, 1963, pl. 4, p. 59
Thule (ca. lat. 76°30′, long. 69°)	130 feet	Beach ridges.	NICHOLS, 1953b, p. 269
North of Saunders Ø on the mainland	150 feet	Raised marine terraces	KRINSLEY, 1963, p. 59
Carey Øer (70 miles west of Thule; ca. lat. 76°40′, long. 73°)	265 feet	Highest beach on Northwest Island; many lower beaches.	BENDIX-ALMGREEN, FRISTRUP, NICHOLS, 1967, p. 14
Etah (ca. lat. 78°20′, long. 73°)	40 feet 90 feet 125 feet 200 feet 290 feet 350 feet 400 feet	Several terraces whose origin is not specified; lower ones may be fluvial terraces.	BENTHAM, 1936, p. 428. See also TROELSEN, 1950, p. 14
Northwestern Greenland (lat. 77°–81°)	1050–1800 feet	The more recent work does not substantiate these measurements. They are much too high.	BØGGILD, 1928, p. 248

The work of KOCH (1928a, p. 517–518), KRINSLEY (1963, plate 4, p. 59), NICHOLS (1953b, p. 269), and BENDIX-ALMGREEN, FRISTRUP, and NICHOLS (1968) showed that the marine limit in the Thule, Greenland area is less than 300 feet above sea level (table 4), and the writer's field work in Inglefield Land showed that here too the marine limit is below 300 feet (tables 1, 4). These workers do not support BØGGILD (1928, p. 248; DALY, 1934, p. 142), who reported that the marine limit on the west coast of Greenland between 77°–81° is 1050–1800 feet above sea level. BØGGILD (1928, p. 248) writes: "For the regions round Smith Sound the figure given is about (320 metres), whereas Bessels in Polaris Bay (at Petermann Fiord) found shell layers and driftwood at the unique height of 1800 ft. (about 550 metres)." It seems likely that at Smith Sund uplifted erosion surfaces and/or struc tural terraces were measured instead of raised beaches or other marine features. The shells at Petermann Fjord may have been carried to their present elevation by birds, wind, or glaciers, and the driftwood by man (BRETZ, 1935, p. 221).

The Tilt of the Marine Limit

On the basis of available data on the altitudes of the highest marine limits in West Greenland, DALY (1934, p. 142) concluded that there was a systematic downward tilt from north to south. This, he reasoned, could not be due solely to isostatic rebound consequent upon deglaciation, and he concluded that a recent independent epeirogenic movement must be

Table 5. *Elevated marine features of North Greenland*

Location	Altitude	Remarks	Observer or Authority
Jørgen Brønlund Fjord, south coast Peary Land (lat. approx. 82°15′)	377–394 feet	Mollusk shells in marine clay.	FRISTRUP, 1952, p. 91.
Jørgen Brønlund Fjord	371 feet	Post-glacial marine terraces.	TROELSEN, 1949, p. 19–21 See also LAURSEN, 1954, p. 1–26.
North Greenland	689 feet	Marine limit.	KOCH, 1928a, p. 517.
Peary Land	656 feet	Ice-pushed features in post-glacial marine clay.	KOCH, 1928a, p. 517.
North Greenland	443 feet	Shells.	KOCH, 1928a, p. 517.
Peary Land (eastern part)	541 feet	Driftwood.	KOCH, 1928a, p. 517.
Northeast Greenland; between J.P. Koch Fjord and Dijmphna Sund	656 feet	Marine terraces.	KRINSLEY, 1961, p. 747.
North Greenland; Kronprins Christian Land and Peary Land	Marine silt 225 feet; raised beaches 300 feet	Marine silt; submergence occurred 5400 years ago, based on radiocarbon dating of shells.	DAVIES and KRINSLEY, 1961, p. 3, 7.

responsible for the tilt. The more recent data assembled in tables 1, 2, and 4 indicate, however, that in a general way the marine limit decreases in altitude from south to north. The marine limit may be 886 feet above sea level at 66° N., approximately 755 feet at 71° N., approximately 266 feet at 76° N., and about 210 feet at 79° N. These data can be explained by isostatic rebound consequent on reater deglaciation in the south than in the north. Thus the independent epeirogenic movements postulated by Daly did not take place.

The marine limit in Northeast Greenland (table 5), however, is at least 689 feet above sea level. This is more than twice as high as that in Inglefield Land (tables 1, 2). The systematic downward tilt of the marine limit from south to north apparently does not continue on to Northeast Greenland.

Recent Marine Stillstand

The inter-tidal zone of the modern delta at Rensselaer Bugt is approximately a mile wide and a mile long (fig. 14). Although nothing is known about its rate of growth, it probably took hundreds of years to form and in any event represents a prolonged stillstand of sea level at about its present position. Similar deltas are found at the mouths of the other glacier-fed streams of Inglefield Land, and they are also found in East Greenland (Bretz, 1935, p. 362; Flint, 1948, p. 332). Sugden and John (1965, p. 244–245) report that the inter-tidal zone of the Gurreholms delta in Scoresby Sund, East Greenland, is 1.9 miles wide and 1.2 miles long and necessarily represents a prolonged present-day marine stillstand.

A marine cliff about 40 feet high, cut into elevated beach deposits and formed when sea level was at or near its present position, is found immediately south of the most northerly stream that enters Rensselaer Bugt from the east. Similar cliffs, which cut both elevated deltas and beach deposits, are found on the southwest side of Rensselaer Bugt, at the hook at the northeast tip of Bancroft Bugt, and on the irregular peninsula on the west side of the bay into which the Minturn Elv empties (fig. 1). These cliffs also indicate a stillstand of sea level at its present position of some duration.

Although the precise positions of the succession of sea levels at which these marine cliffs and deltas were formed are not known for certain, it can be said with certainty that they were formed when the sea level was at approximately its present position. Noe-Nygaard (1932, p. 22) noted that in East Greenland marine caverns, holes, and abrasion terraces cut in bedrock are found only along the present high-water mark although sea level was once higher. He also noted that where there is a continuous series of beaches extending from sea level upward, the lowermost beach is larger, higher, and contains more material than any of the beaches above it. He concluded that a stillstand of sea level at its present position was responsible for these features.

WASHBURN's (1965, p. 32–37) and WASHBURN's and STUIVER's (1962, p. 63–73) curves, showing the rate of emergence of the Mesters Vig district in East Greenland in post-glacial time, dramatically prove that for the last few thousand years there has been a stillstand between land and sea at about their present position in this area.

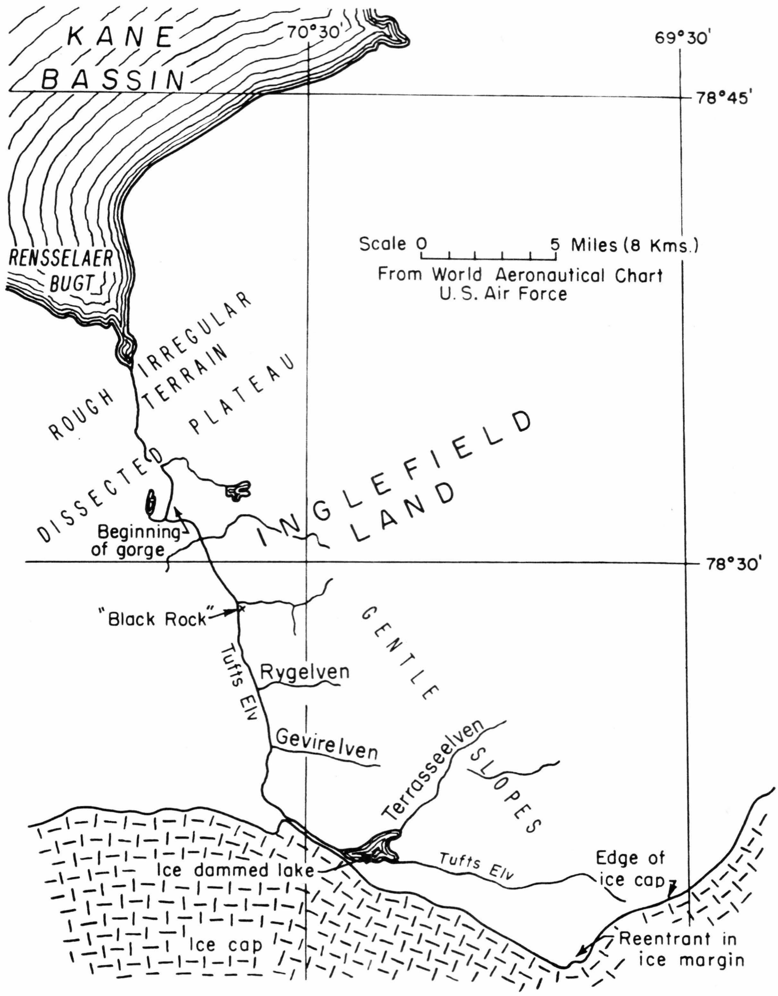

Fig. 17. Map showing the drainage basin of the Tufts Elv.

GLACIATION

Mantle Rock Deposits

Till. Till is the most abundant mantle rock deposit. Flat or gently sloping till plains, more or less devoid of vegetation, cover great areas (figs. 17, 18). Most of the till is ground moraine although a small amount is found in recessional moraines. Striated fragments are very rare in the writer's experience although Mr. RICHARD W. LEMKE found many striated quartzite fragments.[1] In places, the till contains an abundance of the fine sizes; this was shown in one locality by desiccation cracks which

tilbagetræden

[1]) Personal communication from Mr. RICHARD W. LEMKE, U.S. Geological Survey.

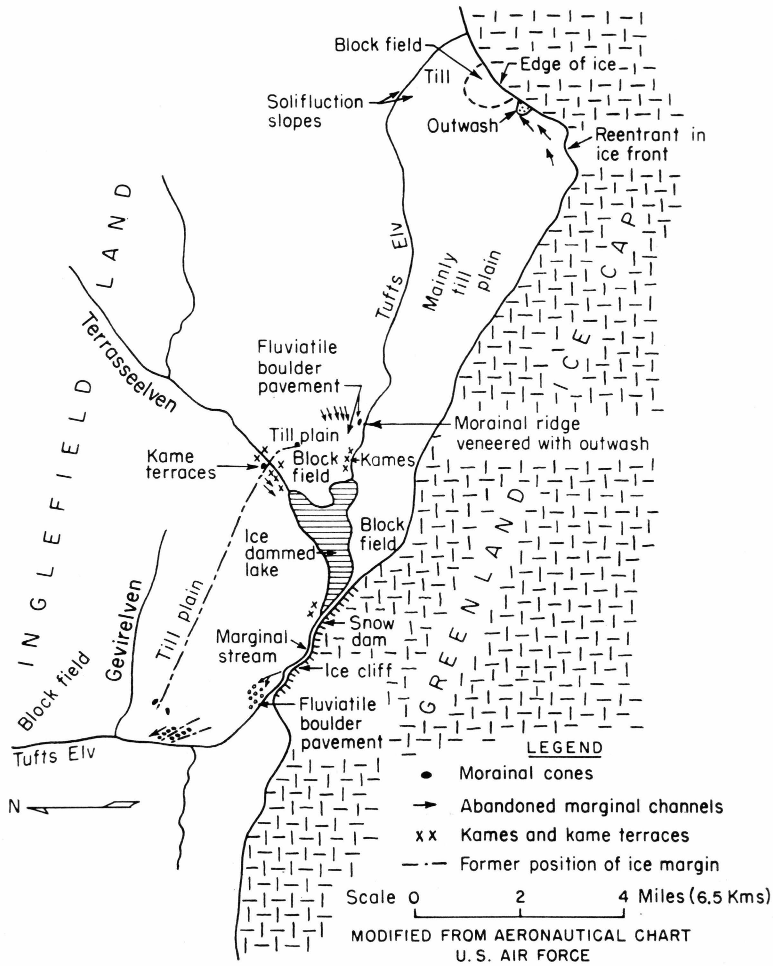

Fig. 18. Map showing ice-dammed lake, till plains, block fields, marginal channels, and kame terraces along the Tufts Elv.

rudely outlined polygons approximately a foot across. The till is stained yellowish brown at the surface, and at one place, about 2 miles from the reentrant in the ice front (figs. 17, 18), blasting showed that it is brown to a depth of 30 inches.

Twelve feet of till was seen in a cut where the marginal river leaves the Indlandsis and flows northward (fig. 19). Till is either very thin or entirely absent over a large part of the area where the Pre-Cambrian basement rocks crop out. On the other hand, seismic blasting near the reentrant in the ice front showed that the mantle rock in this area is approximately 90 feet thick.[1]) The relative abundance of the various

[1]) Personal communication from Mr. Jean Jacques Holtzscherer.

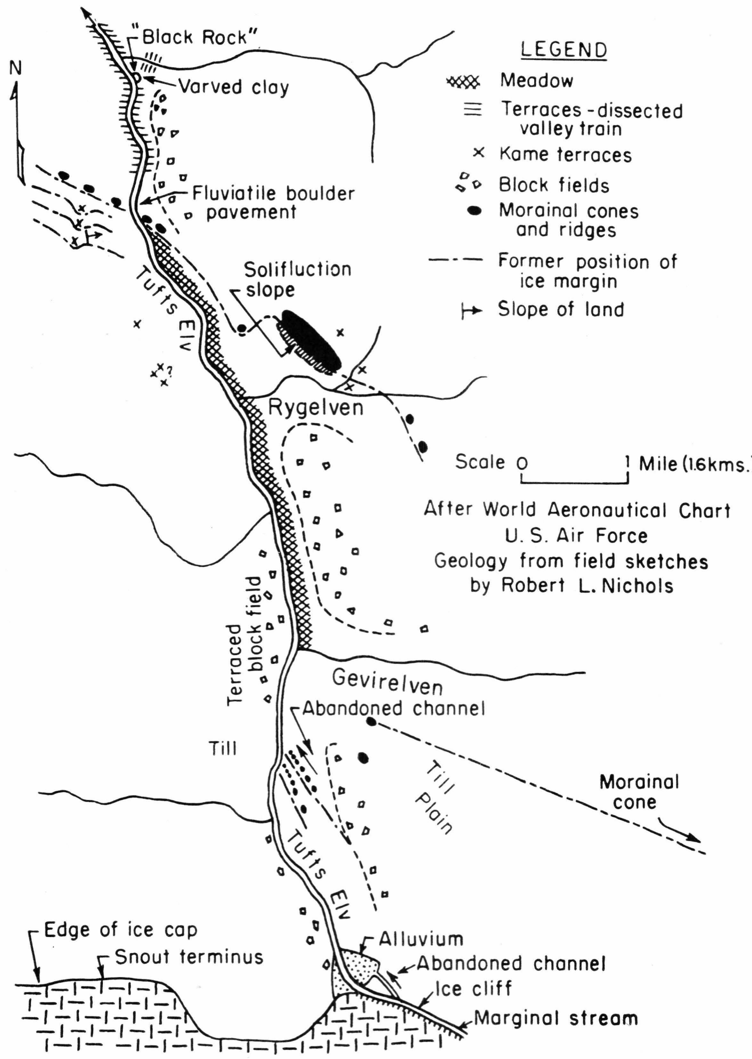

Fig. 19. Map showing the morainal cones and ridges, dissected valley train, block fields, and kames along the Tufts Elv.

mantle rock deposits in the area, and the small chance that any residual mantle is now present, make it seem likely that most of this is till.

In the area along the ice front, a short distance northwest of the reentrant, the mantle rock is striped. Areas having few large fragments but abundant fine material are adjacent to areas with a much higher proportion of large fragments and a much smaller proportion of fine material. The contact between the areas is sharp, straight, and perpendicular to the ice front. The easiest way to explain this distribution is to assume that there are two kinds of till because different kinds of bedrock

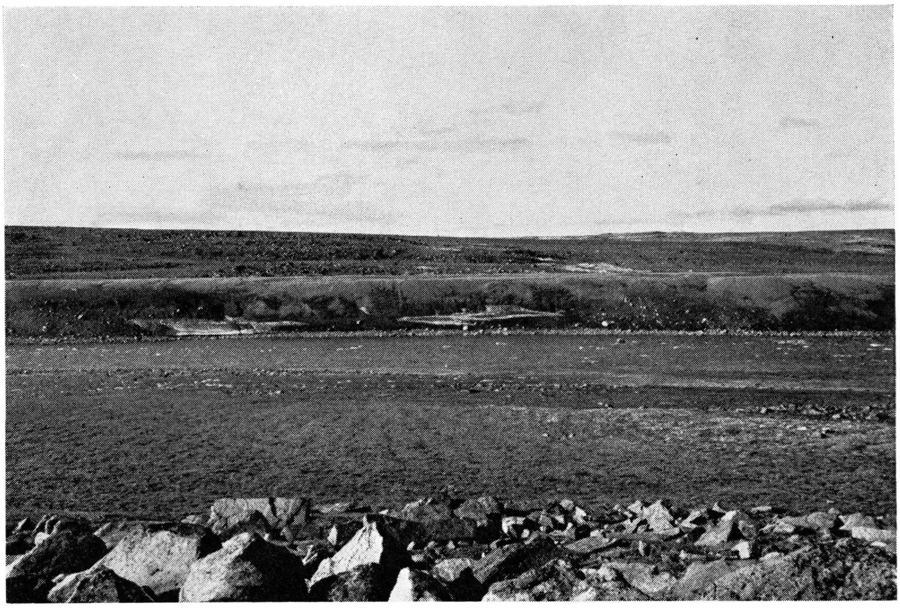

Fig. 20. A fluvial terrace 15–20 feet high cut in the valley train several miles upstream from the delta in Rensselaer Bugt.

are present. A fine example of bedrock control on the lithology of the till is also found near the reentrant in the ice margin. Here, in an area approximately 150 feet wide and 500 feet long, the till is dark colored, whereas it is light colored in the adjacent areas. The dark color results from the presence of mafic rock in the area.

The till has been buried by glacio-fluvial and solifluction deposits, modified by periglacial soil processes, eroded in marginal channels, and transformed into fluvial boulder pavements by meltwater streams.

Stratified Drift. The stratified drift is found principally in thin patches of sand and gravel, small outwash fans, kames, kame terraces, marine deltas, and valley trains. It is very abundant in the deltas at the mouths of the principal rivers and in the dissected valley train which is found on both sides of the Tufts Elv downstream from the aligned morainal cones and ridges (figs. 19, 20). All of the stratified drift studied was stained yellowish brown.

It is the writer's impression that upstream from the morainal cones and ridges the volume of till is perhaps a thousand times that of stratified drift. The outwash in Inglefield Land is derived from superglacial moraine and from erosion of glacial till in front of the Indlandsis. The absence of significant quantities of superglacial moraine may account for the scarcity of stratified drift.

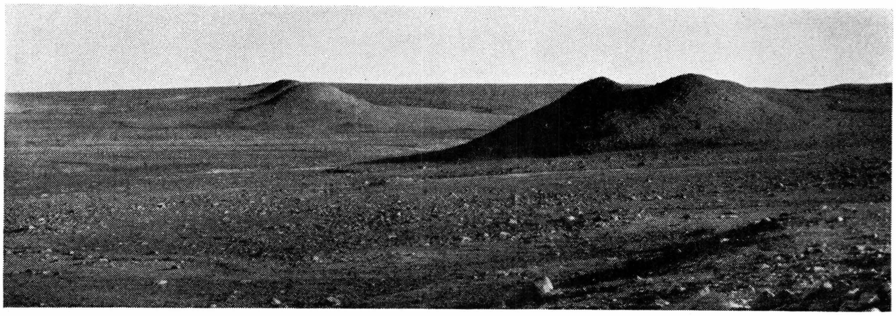

Fig. 21. A line of four morainal cones which extends for more than a mile.

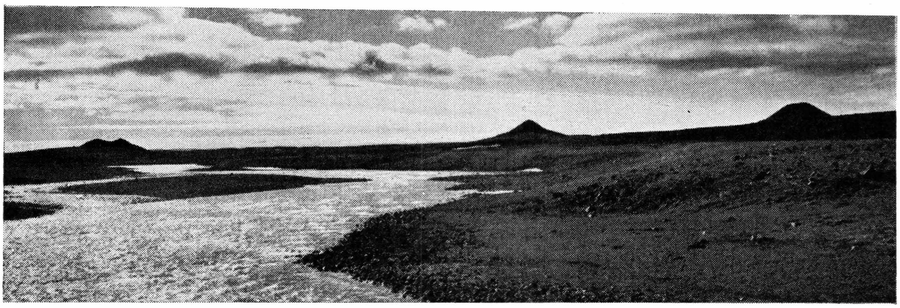

Fig. 22. Three aligned morainal cones.

Varved silt crops out at the bottom of the terrace a few hundred feet upstream from "Black Rock". It extends along a distance of a few hundred feet. A thickness of about 3 feet is exposed, and it is buried by approximately 37 feet of terrace sand and gravel. It is somewhat stained and in places deformed. Nothing is known of the distribution of the water body in which it was deposited. This is the only location where varved silt was seen.

The presence on both till and stratified glacial deposits of a limonite staining which in places extends downward for more than 20 feet makes the glaciation roughly Wisconsin in age. The radiocarbon dates substantiate this conclusion. It is not known why the staining extends down into the permafrost. Perhaps it occurred during the warmer climate for which LAURSEN (1950, p. 110) found evidence.

Morainal Cones and Ridges

Twenty-two morainal cones and three morainal ridges were mapped (figs. 18, 19). The cones range from 32 to 80 feet in height and are composed of unsorted, sandy, bouldery material (fig. 21). Boulders 6 feet long are common on them, and one boulder 12 feet long was measured. The sides of the cones and ridges are at about the angle of repose for the

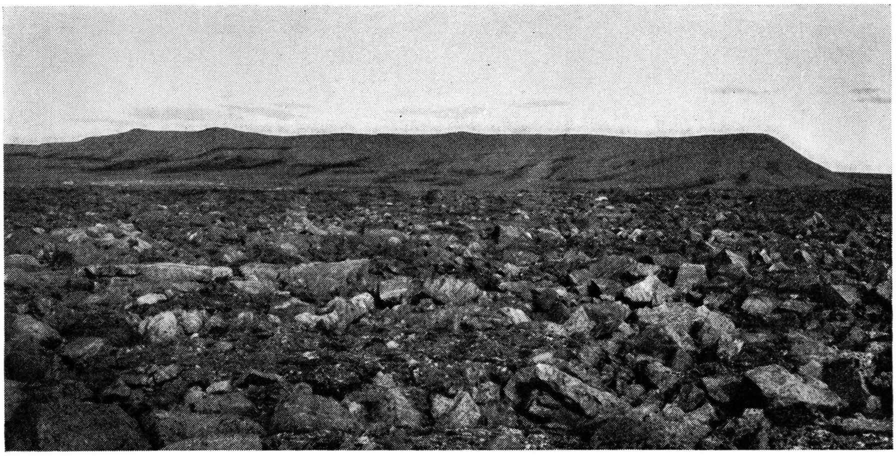

Fig. 23. A morainal ridge approximately 3500 feet long and 160 feet high. A block field is shown in the foreground.

material. One slope was found to be 26 degrees. Most of the cones are completely separated from each other, although coalescing cones are also present. Single simple cones are the rule; however, one cone with a double summit was seen (fig. 22).

The ridges differ in no respect from the cones except that they are elongated. One of the ridges is about 750 feet long, 200 feet wide, and 50 feet high; and its slope where measured was found to be 26 degrees. The largest ridge dwarfs all other morainal features. Its length is approximately 3000 feet along its summit and 3500 feet along its base. At its northwest end it terminates in a steep slope approximately 70 feet high and at its southeast end in a 29-degree slope about 100 feet high; and it has a maximum height of about 160 feet above the surrounding country (fig. 23). Its summit is not flat but is characterized by knobs and swales.

These cones and ridges are bordered by till plains, kame terraces, solifluction slopes, and fluviatile boulder pavements. They have been deposited on both flat and sloping terrain. There is no evidence that either the cones or the ridges have suffered any significant erosion since their formation.

Many of the cones and ridges are aligned in more or less straight lines (fig. 21). The most prominent and important of these alignments contains 9 cones and ridges and extends for between 4 and 5 miles. The largest ridge is in this system, and its long axis is parallel to it (fig. 19). Three other groups of cones and ridges are also aligned (figs. 18, 19). The absence of associated erosional features indicates that the discontinous distribution of aligned morainal material is due not to differential erosion but to differential deposition.

The writer does not believe that these cones and ridges are push moraines. This possibility is ruled out by their shape, their lack of continuity, and the absence of scars associated with them from which the material to form them could have been derived. Their steep sides and the absence of streamlining similar to that seen in drumlins make it certain that they were never overrun by the ice and that they are not lodge moraines. They are, therefore, dump moraines. They were derived mainly from superglacial moraine which had accumulated on and at the border of the Indlandsis.

The border of the Indlandsis at Thule is covered with superglacial moraine not derived from nunataks but brought up along shear planes (BISHOP, 1957; GOLDTHWAIT, 1951, p. 567–571). At one place near Thule, only a few scores of feet away from the edge of the ice, there are four cones composed of unsorted morainal material, the largest of which is more than 20 feet high. There was no evidence that these cones had an ice core. They line up in nearly a straight line which is approximately parallel to the edge of the ice. It seems certain that the morainal material in these cones slid, slumped, and fell off the steep edge of the Indlandsis and was deposited at the foot of the ice. Sags and ravines in the ice front probably concentrated the material, and this, together with a somewhat irregular distribution of moraine on the ice, resulted in a series of discontinuous cones. Superglacial moraine is also found on the margin of the Indlandsis in Inglefield Land, although there it is discontinuous and not as plentiful as at Thule. Nunataks are not present, and the moraine was also brought to the surface along thrust planes which are present in great abundance. It seems likely that the cones and ridges in Inglefield Land were formed in the same way as the cones at Thule.

Two possibilities exist concerning the age relations of the aligned cones and ridges. *First*, the aligned cones and ridges might have been formed simultaneously at one position of the ice margin and therefore constitute a discontinuous recessional moraine. *Second*, the aligned cones and ridges might have been deposited in a persistent reentrant along the ice margin at different times as the ice margin slowly retreated. In this event, although they are aligned they are of different ages, do not mark the position of the ice margin at a single time, and do not form a recessional moraine.

That they were formed at about the same time and therefore constitute a special type of recessional moraine is suggested by the following. *First*, the cones and ridges are aligned in a direction more or less parallel to the ice margin at the present time. *Second*, if the aligned cones and ridges were twice as numerous in the same distance no question would arise about their age relations. It would be generally agreed that they were contemporaneous and that they formed a morainal system. The fact that in a distance of between 4 and 5 miles there are as many as 9 cones and ridges is, for the writer, an argument for contemporaneity. *Third*, if they had been formed at different times it seems likely that the margins of the ice at these times would have been at considerable angles to the nearest margin at the present time. This probability, together with the fact that no evidence for the former existence of these margins was found, suggests a contemporaneity for the cones and ridges. *Fourth*, aligned morainal cones are found at Thule; here the edge of the Indlandsis is parallel to the alignment of the cones and only a short distance away, proving that the cones are essentially synchronous.

For their formation these recessional moraines needed fairly long intermittent halts of the edge of the Indlandsis during the general retreat of the ice. How long the halts were is not known.

Superglacial moraine may have been very much more abundant when these cones and ridges were deposited than at the present time. The superglacial moraine is now formed so slowly that it would take thousands of years to form a ridge as large as the largest one. As this amount of time was probably not available, it seems likely that the ice is now less active and that moraine is being carried up to the surface of the Indlandsis along thrust planes more slowly than when the cones and ridges were being formed.

The presence of these large recessional moraines shows that deglaciation did not take place by stagnation.

Kame Terraces

Kame hillocks and terraces are found at several places (figs. 18, 19). Twenty-four small terraces are located along both sides of the Terrasse-elven, extending from the ice-dammed lake to approximately a mile above it (figs. 18, 24). They are composed of fine gravel which is weathered and stained yellowish brown at the surface. They vary greatly in size. The largest terrace studied is approximately 500 feet long, 250 feet wide, and in places more than 20 feet high. The others range downward in size to thin patches and layers of gravel covering much smaller areas. The terraces are characterized by flat, nearly horizontal upper surfaces, which are bordered by steep slopes except on the side away from the river, where they commonly merge almost without break into the gently sloping valley wall (fig. 25). They were deposited on till and are separated from each other by till slopes modified by frost action and solifluction. One small terrace was deposited on the edge of a morainal cone.

On the surface of a few of these terraces, boulders and blocks are found. They are of two kinds. Those farthest from the river were depo-

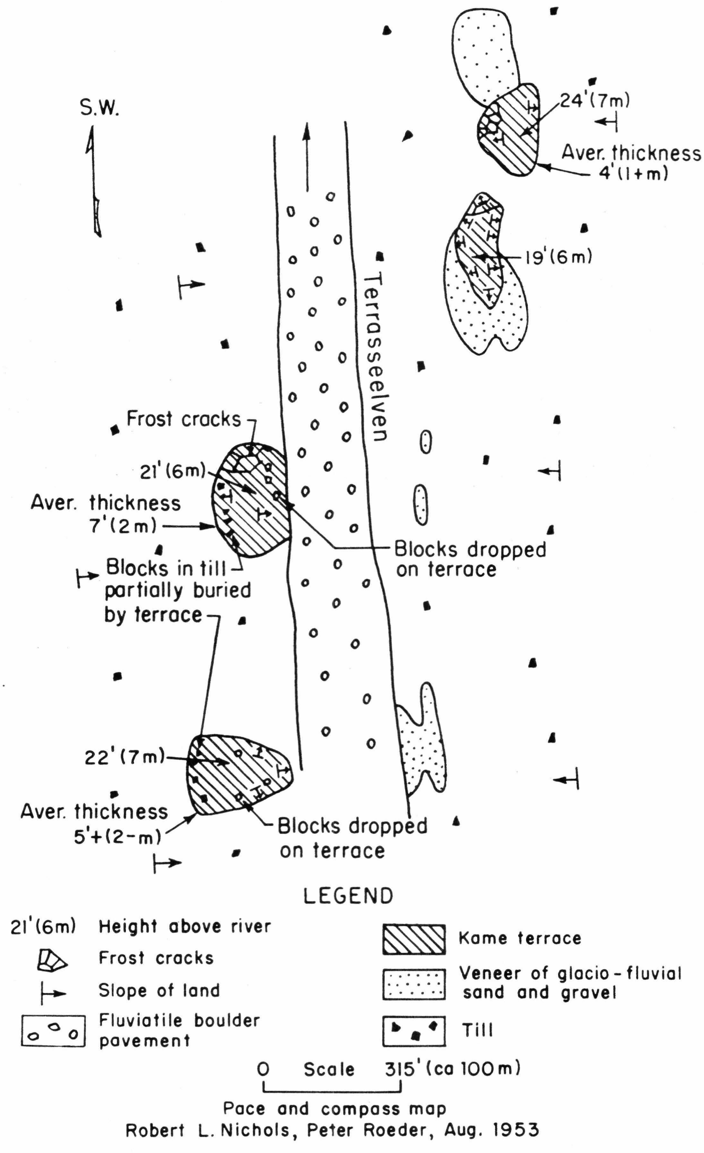

Fig. 24. A map showing the kame terraces and patches of sand and gravel along the Terrasseelven.

sited with the ground moraine and were too large to be completely buried by the feather edges of the terraces. Those nearest the river were dropped on terrace gravel after most of it had been deposited. A small kame hillock was deposited on the tread of one of the terraces. Prominent frost cracks are found on both treads and risers.

The terraces extend from the river up to scores of feet above it. Their area is small compared to that of the terrain between them (fig. 25).

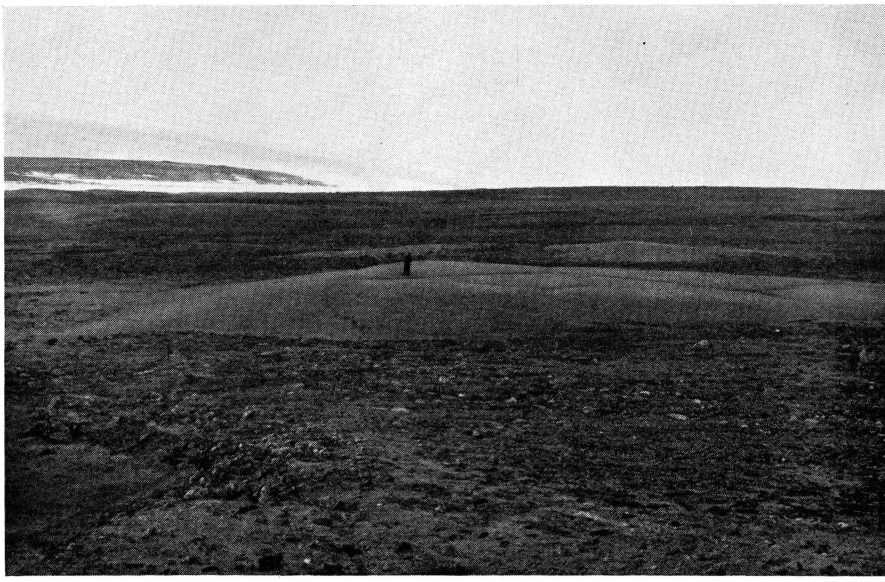

Fig. 25. A kame terrace along the Terrasseelven less than a mile from the northern arm of the ice-dammed lake. A weathering profile has been developed on it, and frost cracks cross it.

The terraces farther up the Terrasseelven valley and consequently farther from the present ice margin are older than those farther down the valley. Unfortunately nothing is known of the attitude of the bedding in these terraces, or of the thickness of the weathered zone. Marginal channels cut in the sides of the valley are associated with the largest of these terraces. The channels do not run down the steepest slopes of the valley wall but are more or less parallel to the river. The largest can be traced for hundreds of feet, is more than 10 feet deep, has the typical cross section and gradient of a channel, and has a boulder pavement at the bottom. These channels do not now carry any significant quantities of water.

Similar terraces were seen at several other places in Inglefield Land (figs. 18, 19).

The possibility was considered that these terraces are: (1) beach deposits formed in the ice-dammed lake when it was higher than now; (2) deltas formed in the ice-dammed lake when it was higher than now by streams running more or less perpendicular to the Terrasseelven; (3) remnants of a dissected valley fill; (4) remnants of a dissected proglacial delta deposited in the ice-dammed lake when it was higher than now. Analysis of the data showed, however, that they are kame terraces formed when the ice front was 2 or 3 miles beyond its present position.

They were formed where streams carried sand and gravel off the glacier into ponds along its margin. The fact that some of the terraces merge imperceptibly with the valley wall on their up-slope sides indicates that the ponds in which they were formed were completely filled with sediment on the side farthest from the glacier. The terrace slopes are either ice-contact slopes or slopes resulting from the deposition of fore-set beds. The large boulders found on the terraces probably slid, rolled, and fell off the steep edge of the ice. The hillside channels associated with some of the terraces were cut by the outlet streams of the marginal ponds. The fact that channels are not associated with every terrace indicates either that some of the outlet streams were too small and acted for too short a time to erode a channel, or that the outlet streams, if the climate was warmer than that at present, ran under the ice.

A large reentrant in the ice running down the Terrasseelven valley would not favor the formation of these terraces, whereas a lobe of ice running up the valley would facilitate their formation.

Valley Trains and River Terraces

Glacio-fluvial deposits cut into a series of non-paired terraces are found in the valleys emptying into Force Bugt, Rensselaer Bugt, Bancroft Bugt, and Marshall Bugt, as well as along the Minturn Elv and the stream that reaches Kane Bassin immediately southwest of Bancroft Bugt (fig. 1).

Prominent and well-developed terraces cut in glacio-fluvial deposits are found for approximately 10 miles along both sides of the lower part of the Tufts Elv. They extend upstream from "Black Rock" more or less continuously for about a mile and downstream from "Black Rock" discontinuously to Rensselaer Bugt (figs. 17, 19).

The deposits are composed mainly of sand. A veneer of loess or a thin pebble pavement resulting from deflation is commonly found at the surface. At one place three feet of varved silt was found beneath 37 feet of sand. Large boulders were seen on the terrace slope on the west side of the river; whether they are a part of the terrace deposit or were deposited on the slope by the river is not known. There is a suggestion, based on a superficial analysis of the size of the roundstones in the pebble pavement on top of the terrace, that the deposits get somewhat coarser going upstream from "Black Rock".

In some places the terraces are devoid of vegetation; in others they are covered with grass and tussocks. Excellently developed frost cracks and the resulting polygons are common on both treads and risers. Several small kettle holes a few feet deep and a few score feet in diameter were seen on the highest terrace, and fore-set beds were seen in two cuts.

The highest terrace on the east side of the river is commonly about 300 feet wide and immediately upstream from "Black Rock" is more than 50 feet above the river; the upper terrace on the west side appears to rise to a similar height. Several lower terraces are located at many places along the west side of the river. In a few places the river has cut through the glacio-fluvial deposits and is running on bedrock.

That the glacio-fluvial deposits have some antiquity is shown by the following: (1) The small side streams have eroded the deposits considerably. (2) The deposits in places are stained yellow to a depth of 40 feet below the surface. (3) Some of the pebbles in the deposits are coated with limestone.

There is no evidence that enough erosion has taken place upstream from the deposits in postglacial time to account for the material found in them. Therefore they must have been deposited by streams which obtained the material from the Indlandsis.

The river terraces are either kame terraces or have resulted from the dissection of a valley train.

The following indicate that they have resulted from the dissection of a valley train. (1) The Tufts Elv and the side streams are large enough to have accomplished the necessary dissection in the available time. (2) The comparative absence of kettle holes on the upper terrace suggests that the fill is a valley train rather than a kame terrace deposit. (3) A series of terraces such as exists along the west side of the Tufts Elv can perhaps be explained better by dissection of a valley train than by dissection of a kame terrace. (4) The valley does not seem deep enough for the formation of a lobe of ice which would have persisted until kame terraces of the necessary size were formed. (5) The horizontal distance between the terraces on one side of the river and those on the other is so small that neither a stagnant nor an active lobe of ice of this width could ever have extended down the valley for 10 miles. (6) At present no lobes extending from the edge of the Indlandsis are as large as the lobe that would have been necessary if the river terraces along the Tufts Elv are kame terraces. (7) The alignment of morainal cones and ridges does not indicate the former existence of extensive lobes of either stagnant or active ice.

The valley train terminates in a delta at Rensselaer Bugt.

The valley train and delta were probably deposited more or less continuously as the Indlandsis retreated from Rensselaer Bugt to the position outlined by the morainal cones and ridges (fig. 19). As indicated above, a considerable halt in the retreat of the Indlandsis was necessary to form this recessional moraine; during this period the younger parts of the valley train and delta were deposited. After the moraine was formed, the Indlandsis continued to retreat with occasional minor stillstands. However, no valley train was formed upstream from the recessional

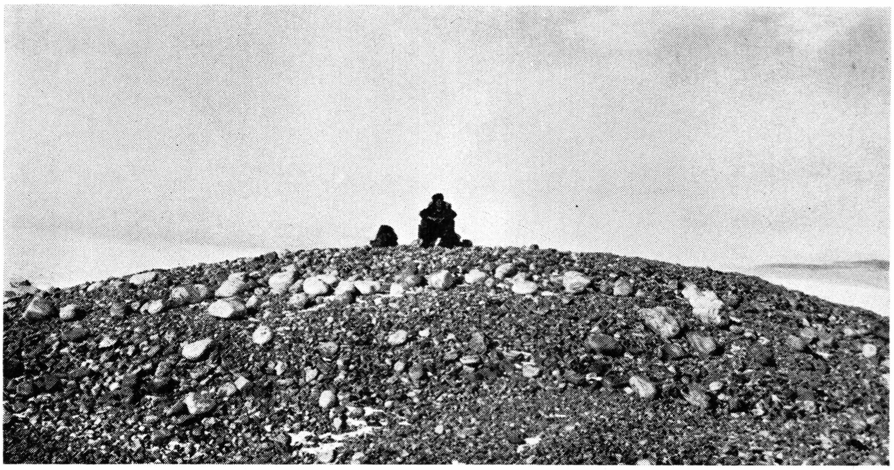

Fig. 26. Superglacial moraine on the Indlandsis near the reentrant in the ice margin about half a mile from the edge of the ice. It consists of sand and gravel, and the roundstones are up to 2 feet in diameter.

moraine (fig. 19). The reason for this is not known for certain, but the following factors bear on the problem.

The glacio-fluvial deposits in Inglefield Land were not derived from meltwater that emerged from beneath the Indlandsis, but from superglacial moraine (fig. 26) and from erosion of subaerial glacial deposits near the ice front. Because there are no nunataks in this part of Inglefield Land, the superglacial moraine is brought from beneath the Indlandsis up to its surface along shear planes. If closely jointed bedrock and an abundance of mantle rock with little interstitial ice beneath the margin of the Indlandsis were followed, because of glacial retreat, by massively jointed bedrock and an absence of mantle rock beneath the margin of the Indlandsis, then the quantity of superglacial moraine would probably decrease, thereby reducing the volume of glacio-fluvial deposits that could be formed. If true, this might help to explain the absence of a valley train upstream from the cones and ridges.

If the Indlandsis had retreated faster upstream than downstream from the cones and ridges, there would have been less time to form superglacial moraine upstream and therefore there would have been less moraine on the Indlandsis. Under these conditions a smaller amount of outwash would have been deposited upstream from the cones and ridges than downstream from them.

The ice-dammed lake in the upper part of the Tufts Elv valley (figs. 17, 18) may be the most important factor in accounting for the absence of a valley train upstream from the cones and ridges.

Meltwater from along more than 15 miles of the Indlandsis empties into this ice-dammed lake at the present time. It may be that ice-dammed lakes did not exist when the ice front stood downstream from the position outlined by the cones and ridges, but that almost immediately thereafter one or more came into being. When this happened, glacio-fluvial material that might otherwise have been transported down the Tufts Elv was trapped in the lake(s) and was therefore not available to form a valley train upstream from the cones and ridges. Any meltwater from the ice front which ran down the Tufts Elv rather than into the ice-dammed lake(s) did not deposit significant quantities of outwash upstream from the cones and ridges. Moreover, the stream(s) running out of the lake(s) were underloaded; and probably the cutting of terraces in the valley fill downstream from the cones and ridges began when the ice-dammed lake(s) were formed.

A difference in stream gradients and valley widths is not involved, however, because the gradient and width of the Tufts Elv are about the same upstream from the cones and ridges as they are downstream from them.

Flint (1947, p. 138) has attributed the lack of outwash in some areas to the presence of till that is rich in clay and poor in stones. This is probably not the case in Inglefield Land, where, as indicated above, the till is in places sandy and stony.

The delta in Rensselaer Bugt continued to grow as the valley train was deposited. The youngest part of it was formed of material removed from the older part of the delta and from the valley train during the formation of the terraces in both delta and valley train.

The other valleys that contain terraced glacio-fluvial deposits have undoubtedly had a similar history.

Marginal Channels

The topography along the edge of the Indlandsis slopes toward the ice in places; here ice-marginal streams, ponds, and lakes are found. The streams run along the edge of the ice until they reach a major drainage line leading away from it. They may hug the ice front closely or may leave it for a distance because there is a sag or swale oblique to the ice edge. They run on both bedrock and mantle rock.

The marginal streams not only remove snowbanks deposited during the winter months and cut back the edge of the glacial ice, but also erode both mantle and bedrock.

The total amount of erosion accomplished by the marginal streams in Inglefield Land, however, is not great. The size of a marginal channel is a function of the following: *First*, the length of time during which the marginal stream has been cutting the channel. This is controlled by the stability of the ice front. *Second*, the volume of water that flows subaerially along the edge of the ice. *Third*, the tools furnished to the river. *Fourth*, the slope of the land toward the ice. The steeper the slope, the more the runoff is concentrated in a narrow channel and consequently the greater the erosion. *Fifth*, the slope along the margin of the ice. This, together with the volume of the marginal stream, controls its velocity. *Sixth*, the kind of material being eroded.

Abandoned marginal channels are found in many places. A marginal channel 13 feet deep cut in till and lacustrine sands is located about a mile northeast of the reentrant in the ice front (fig. 18). Six are located 1–2 miles above the ice-dammed lake, on the north side of the river that runs into the southern arm of the lake. Some of these are $^1/_4$ mile long and nearly 10 feet deep. An extensive fluviatile boulder pavement is found immediately downstream from them (fig. 18).

A large and well-developed abandoned marginal channel is shown in figs. 18, 27. It is more than half a mile long, about 200 feet wide, and in places as much as 30–40 feet deep. It has a channel-like cross section, and for its entire length it slopes more or less uniformly toward its terminus. How much of this channel resulted from glacial erosion and was therefore inherited by the meltwater, and how much of it is due to fluviatile erosion is not known. The water-worn boulders along the bottom look cleaner and less weathered than the fragments on the channel walls and on the flat above. A large fluviatile boulder pavement is located on the lowland downstream from the lower end of the channel; it was probably formed from till by the unconfined waters after they left the channel. The upper end of the channel hangs between 35–45 feet above the present-day marginal stream. A water-washed surface is found along the stream, downstream from the upper end of the abandoned channel, where the till and other mantle rock deposits have been removed. It extends approximately as high above the river as the abandoned channel hangs above the river, but it does not extend higher because the abandoned channel acts as a spillway in times of flood and prevents water from rising in the present channel much higher than the floor of the abandoned channel (fig. 27).

Older and much larger abandoned marginal channels are found near Bancroft Bugt and elsewhere. At Bancroft Bugt they are cut in bedrock, are hundreds of feet deep, and hang high above the present-day rivers (fig. 4).

The details of the glacial retreat in the areas labelled A, B, and C in figure 4 were not worked out. The V-shaped valley labelled A is 300 feet deep in places, slopes downward toward Kane Bassin, is not occupied by a through-going stream, and hangs above the modern drainage. These cha-

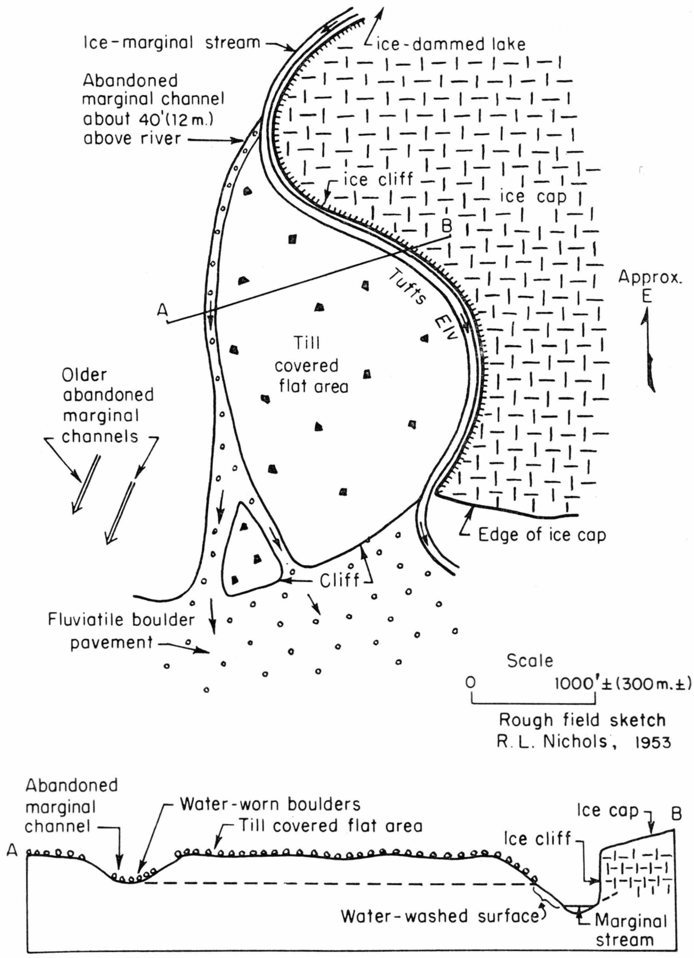

Fig. 27. Map and cross-section showing an abandoned marginal channel a short distance downstream from the ice-dammed lake.

racteristics prove that it was formed by ice-marginal drainage after the area was partly deglaciated. The V-shaped valley labelled B has similar characteristics and was formed in the same way. The difference between these V-shaped ice-marginal channels and the broad pre-glacial valley shown in figure 4, which is characterized by structural terraces, is striking.

The V-shaped gorge labelled C in figure 4 is just as striking as those at A and B. At the cliff end it hangs hundreds of feet above sea level, and at the other end, high above the valley labelled D. Boulder gravels in places along the bottom of the valley (D) show that it was formerly occupied by a river much larger than the present one. The V-shaped gorge (C) was cut by a marginal stream held up by a lobe of ice that occupied Bancroft Bugt.

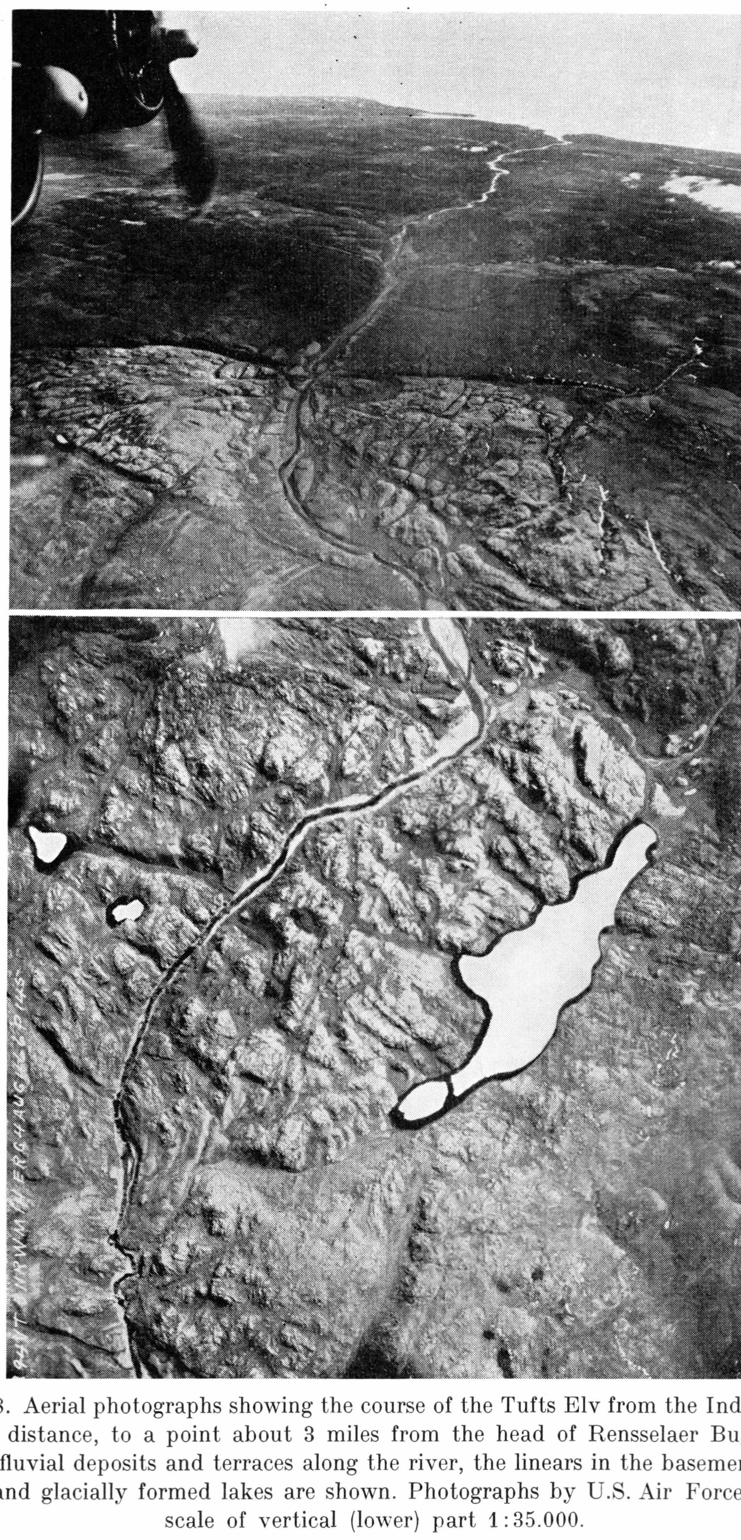

Fig. 28. Aerial photographs showing the course of the Tufts Elv from the Indlandsis, in the distance, to a point about 3 miles from the head of Rensselaer Bugt. The glacio-fluvial deposits and terraces along the river, the linears in the basement complex, and glacially formed lakes are shown. Photographs by U.S. Air Force. Appr. scale of vertical (lower) part 1:35.000.

Fig. 29. The area in the vicinity of Rensselaer Bugt in southwest Inglefield Land. Paleozoic sedimentary rocks on the right and Pre-Cambrian igneous and metamorphic rocks on the left. Criss-cross linears are located in the area of the older rocks. Oblique photograph by U.S. Air Force.

Linears

The topography in the areas of Pre-Cambrian igneous and metamorphic rocks is markedly different from that found on the Pre-Cambrian and Paleozoic sedimentary rocks, where a dendritic drainage somewhat modified by glacial action covers much of the area. Linear topographic features which are generally absent on the sedimentary rocks are common on the igneous and metamorphic rocks (fig. 28).

These linears are relatively straight, some are as much as a few miles in length, and in general they are valleys rather than ridges. In places they are more than 100 feet deep. The topography in these areas is so irregular and fluviatile boulder pavements, block fields, talus, and steep bedrock cliffs are so common that travel for men on foot is slow and tedious. On a basis of orientation there are more than three sets of linears.

A criss-cross topography is found in some areas where linears are at right angles to each other (fig. 28, 29, 30). The writer has seen similar topography in Labrador, and it is also found in other parts of the Canadian Shield. It has been made the object of an interesting study by WILSON (1941, 1948). SPENCER (1959, pl. 1, 2) has studied similar features in Wyoming and Montana, as has TANNER (1938, fig. 129) in Finland.

There are three problems concerning the origin of the linears: (1) What structural control was responsible for them? (2) What processes were involved in their formation? (3) When were they formed? An attempt to answer these questions will now be made.

Linear patterns can be formed by the deposition of till, and rectilinear patterns have sometimes resulted from the formation of crevasse fillings in stagnant ice. Linear patterns can also be formed by elevated beach ridges. Moreover, criss-cross patterns can be formed by the deposition of beach ridges on small hogbacks formed in bedrock (NICHOLS, 1953b, p. 274). The linears in Inglefield Land, however, were not formed in these ways, as they are bedrock features.

Narrow linears resulting from diabase dikes have been identified on the Canadian Shield (WILSON, 1948, p. 701). If the linears in Inglefield Land were due to parallel cross-cutting dikes, a very high percentage of some of the area would have to be dike rock. However, as dikes are not common in Inglefield Land the linears must have some other structural control.

Because only igneous and metamorphic rocks, as far as is known, are found in the area where the linears are located, and because the linears have a rectilinear pattern, sedimentary bedding cannot explain some of them and probably can explain none.

Jointing, faults, and foliation are probably responsible for all or nearly all of the linear topographic features. In the Canadian Shield the joints in massive granite have a cross-hatched pattern (WILSON, 1948, p. 701). Perhaps the shorter linears in Inglefield Land are due to joints and the longer ones to faults.

Several geomorphic processes were involved in the formation of the linears and the topography adjacent to them. As indicated above, an erosion surface was cut on the Pre-Cambrian basement complex before the deposition of the Pre-Cambrian sedimentary rocks of the Thule Group (COWIE, 1961, p. 17, 40). In discussing this surface COWIE (1961, p. 16) writes: "Metamorphic rocks are also found as far north in Inglefield Land as the Humboldt Gletscher, and in Bache Peninsula and neighbouring parts of the Canadian coast. It is characteristic of all these localities that the older rocks were peneplained to a remarkably level surface . . .".

If the sedimentary rocks initially deposited on the peneplain were marine, the residual soils and other terrestrial mantle rock deposits that

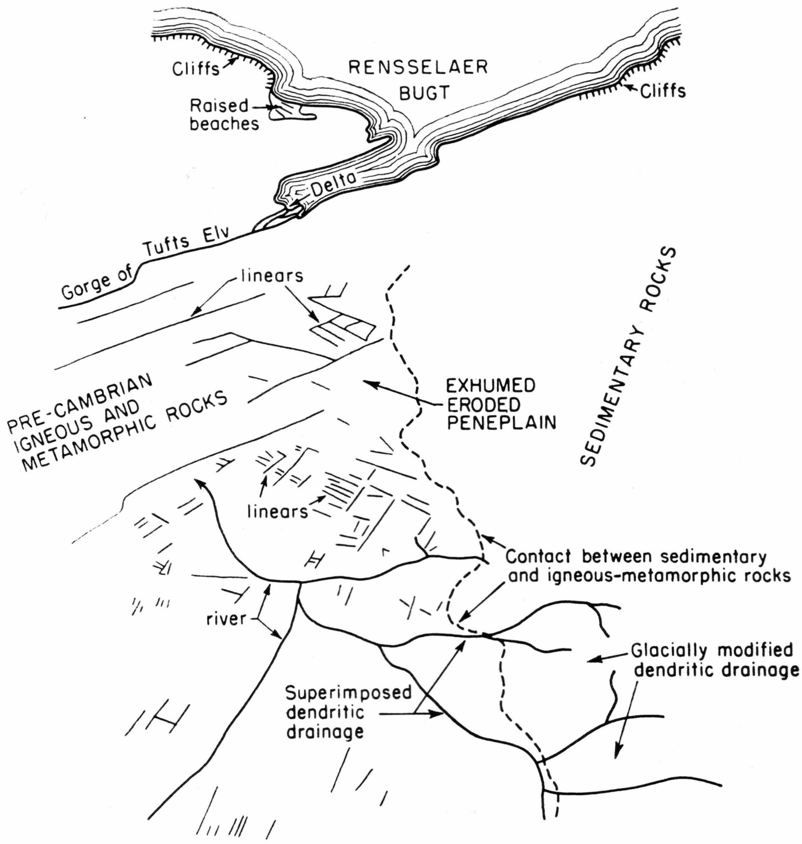

Fig. 30. A map showing the three principal sets of linears and the dendritic drainage pattern superimposed from the sedimentary rocks onto the Pre-Cambrian basement complex. Geology and geography taken from the oblique aerial photograph shown in Figure 29.

may have veneered the peneplain could have been removed by wave action and the relief on the peneplain may have been reduced still further.

The presence of such a flat surface makes it most unlikely that the linears existed in anything like their present size in Pre-Cambrian time. Remnants of the sedimentary rocks of the Thule Group at and near the bottoms of the prominent linears would prove, of course, that the linears were as well developed in Pre-Cambrian time as at present. However, no such remnants have been found.

As mentioned above, a peneplain was cut in Inglefield Land in the late Tertiary. Following this an uplift of several hundred feet initiated a period of fluviatile erosion. Over large areas, and particularly along the major drainage lines, the sedimentary rocks were removed; and the drainage pattern established on them was superimposed on

Fig. 31. A linear inland from the head of Rensselaer Bugt. The steep bedrock slopes, the fluviatile boulder pavements, the talus, and the size of the linear are well shown.

to the exhumed Pre-Cambrian peneplain. The dendritic drainage on the Pre-Cambrian rocks inland from Rensselaer Bugt (fig. 30) may have been formed in this way. Continued erosion destroyed the peneplain (fig. 2) in most areas, and at this time the linears were initiated. At first they were shallow with graded longitudinal profiles and systematic variation in valley width; as dissection continued they progressively increased in width and depth.

Following this, the area was glaciated. The linears oriented in the same direction as the movement of the ice were probably deepened and widened, and those at angles to the direction of ice movement probably had their walls plucked on the upslope side and abraded on the downslope side. Glacial lakes and roches moutonnées in the area where the linears are located, and the poorly graded longitudinal profiles and non-systematic variation in the width of the linears, all show that they are due in part to glaciation.

During the retreat of the Indlandsis, meltwater streams further modified the linears and formed fluviatile boulder pavements from the till deposited in them.

Minor changes resulting from frost processes, gravity, and weathering have taken place since the meltwaters disappeared (fig. 31).

Linears are not common in the area near the reentrant in the ice front (figs. 17, 18) although here too Pre-Cambrian igneous and metamorphic rocks are present (fig. 28). Their absence may be due to a thick cover of ground moraine.

Extent of Glaciation

The presence of glacial striations on the coastline between Dallas Bugt and Advance Bugt, glacial striations and numerous well-developed roches moutonnées around Dallas Bugt, glacially planed surfaces on the coastline on the southwest side of Marshall Bugt, glacial striations around Rensselaer Bugt (table 6), erratics on the plateau between Rensselaer Bugt and Dallas Bugt, and glacially-formed lakes along much of the coastline — all associated with glacial drift only slightly weathered — proves that during the full-bodied stage of Wisconsin glaciation the Greenland Indlandsis completely covered Inglefield Land and filled at least parts of Smith Sund, Kane Bassin, Kennedy Kanal and Robeson Kanal (fig. 5).

Table 6. *Distribution of glacial striations and roches moutonnées in Inglefield Land, Greenland*

Location	Feature	Orientation	Remarks
Half a mile inland from head of Rensselaer Bugt, east side of Tufts Elv	Roches moutonnées; striations.	Striations perpendicular to coast	Abundant
East side of Rensselaer Bugt, between two streams which enter bay from east	Roches moutonnées; striations	Striations perpendicular to coast	
At mouth of Minturn Elv, west side	Striations	Roughly same as valley; perpendicular to coast	Considerable area
Marshal Bugt at Inuarfigssuaq (Eskimo camp)	Glacially planed bedrock; mammillary forms	Indicate ice moved perpendicular to coast	Ice must have extended far out into Kane Bassin
In valley immediately west of Marshall Bugt and south of Inuarfigssuaq	Striations	Running in same direction as valley; perpendicular to coast	
Dallas Bugt, southwest side	Roches moutonnées; grooves; striations; glacially planed bedrock	Parallel to Dallas Bugt	Large areas; many localities
Dallas Bugt, northeast side near head of bay	Excellent striations	Perpendicular to coast	Cover large area
Between Dallas Bugt and Advance Bugt, close to river between them	Striations	Perpendicular to coast	

The Greenland Indlandsis also extended beyond the Carey Øer, which are approximately 70 statute miles west of Thule as shown by erratics (BENDIX-ALMGREEN, FRISTRUP, and NICHOLS, 1967 p. 12); and at about the same time the ice sheet in southern Greenland extended in places as much as 100 miles beyond its present terminus, as shown by numerous glacial lakes which occur in the Søndre Strømfjord area from the terminus of the glacier out to the coast. These facts suggest that during the full-bodied stage of the Wisconsin, Smith Sund, Kane Bassin, Kennedy Kanal and Robeson Kanal were probably completely filled with the Greenland and Ellesmere Island Ice Sheets.

Glacial Retreat

The radiocarbon age determinations of the shells found in the coastal areas (table 3) show that deglaciation occurred at Rensselaer Bugt more than 7800 ± 200 years ago and at Dallas Bugt more than 5900 ± 150 years ago. They also show that the duration of post glacial time progressively decreases from the southwest to the northeast.

An analysis of the distribution of kame terraces, morainal cones and ridges, marginal channels, and valley trains shows that the characteristics of the ice margin varied from place to place during the retreat of the Indlandsis. In some areas streams carried material off the ice and deposited it in lakelets in front of the ice, forming kame terraces. In other areas superglacial moraine rolled, slid, and fell from the ice to the ground, and morainal cones and ridges were formed. A short distance away, only ground moraine was deposited. Not only did the margin of the ice vary at any one time from place to place, but in some cases the same sector of the margin varied from time to time during the retreat. This is demonstrated by a kame terrace which was deposited on the feather edge of a morainal cone.

Glacial Readvance

A glacial readvance has taken place in the North Star Bugt area, North Greenland, within the last 8500 years and perhaps following the climatic optimum (KRINSLEY, 1963, p. 61, 66; WHITE, 1956, p. 24–30). Ice tongues readvanced along the major fiords in Kronprins Christian Land and Peary Land, North Greenland, about 3700 years ago (DAVIES and KRINSLEY, 1961, p. 3; KRINSLEY, 1961, p. 747). A kame terrace deposited on marine silt and clay in Jørgen Brønlund Fjord after the fiord had been deglaciated led TROELSEN (1952, p. 219–220) to conclude that a small glacier had readvanced down Jørgen Brønlund Fjord when sea level was 213 feet higher than at present. LAURSEN (1950) in his studies of the marine Quaternary deposits in West Greenland found an assemblage of shells in Horizon F that indicated a climate somewhat warmer than at present. He was also able to show that whereas the marine limit in Disko Bugt is about 656 feet above sea level, Horizon F was deposited in this area when sea level was approximately 131 feet higher than at present. It seems likely that the deterioration of climate following the deposition of LAURSEN's Horizon F was responsible for the recent readvance of the glaciers in North Greenland.

In view of this evidence for a recent deterioration of climate in West Greenland and a recent glacial readvance in North Greenland, it is not surprising that many features in Inglefield Land also suggest a recent glacial readvance. Evidence for this readvance follows.

1. Superglacial Moraine

A patch of superglacial moraine between $^1/_4$ and $^1/_2$ mile from the edge of the ice and less than a mile from the reentrant in the ice front was studied (fig. 17). It is more than 100 feet across and is composed of material ranging in size from sand to blocks 3–4 feet long. The sand and some of the coarser material are stained yellow, and excellently rounded boulders up to 2 feet in diameter are common.

These gravels could have been formed in pre-glacial, interglacial, or interstadial time, or after the last major glacial advance if following it the Indlandsis retreated some distance behind its present position and later readvanced. The writer doubts that any pre-glacial gravels existed when the superglacial gravels were picked up and brought to the surface of the Indlandsis; it would seem as if they must have been removed long since by glacial action and perhaps by other processes, and he feels that the same argument applies to interglacial and perhaps to interstadial gravels. In view of the evidence for a readvance in the last few thousand years in northern Greenland, it seems likely that the superglacial gravels were deposited after the retreat that followed the last major advance and were later picked up by the Indlandsis during the recent readvance. These gravels were definitely not formed in a subglacial stream because the Greenland Indlandsis is a polar glacier at present and probably has been one since long before the gravels were deposited.

2. Deposition of Weathered Moraine

Yellowish-brown weathered morainal material is falling to the ground from within the glacier at the ice cliff a short distance northeast of the reentrant in the ice front (fig. 17). It seems likely that this material was derived from a weathering profile that was developed, on the deposits laid down by the last major glacial advance, after the ice front had retreated behind its present location but before it had readvanced to its

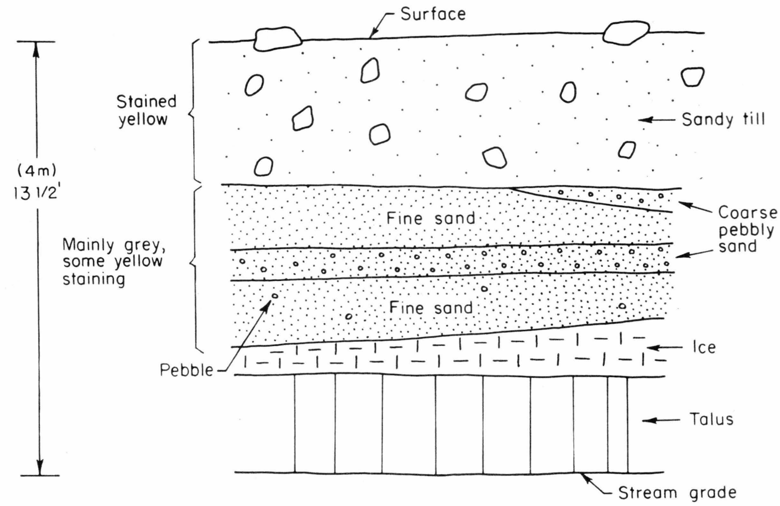

Fig. 32. Weathered till and outwash seen on the walls of a gully cut by a marginal stream near the reentrant in the margin of the Indlandsis.

present position. With this history, weathered material could be brought to the surface of the Indlandsis along thrust planes during a readvance.

3. Weathered Till

The section shown in figure 32 was seen in a cut made by one of the marginal streams. It is located not more than 50 feet from the edge of the glacier and only a short distance from the pronounced reentrant in the ice front (figs. 17, 18). The upper 54 inches is unstratified and composed of material varying in size from sand to blocks. It is probably sandy till, somewhat modified by frost action. It is stained yellow from top to bottom. There is no evidence that any significant thickness of till has been removed by marginal streams, as no boulder pavement is found on it.

Below the till there is approximately 5 feet of fine gray sand which appears horizontal in the section. Pebbles are scattered through it, and in the middle there is a thin bed of coarse, pebbly sand. Although the sand is mainly gray, it is stained slightly yellow in places. It is a lacustrine and/or fluviatile deposit.

The till may have been weathered *in situ*, it may have been already weathered when deposited, or the lower part of it may have been weathered *in situ* and the upper part already weathered when deposited. If the Indlandsis had retreated to its present position and remained there while the weathering profile was developed in front of it, one might expect to find good sized moraines and/or marginal channels at the edge of the Indlandsis, as thousands of years were probably necessary to form this profile. Large moraines and marginal channels, however, do not exist at the margin of the Indlandsis in this area.

It therefore seems likely that the weathering profile, if formed *in situ*, was developed after the ice front had retreated behind its present terminus, and that the front readvanced to its present position after the profile had been developed.

Thus if the till was weathered in place, a readvance of the Indlandsis has occurred. For reasons indicated above, the writer also believes that a readvance has occurred if the till was weathered when deposited.

The marginal meltwater streams in this area are yellowish brown rather than the usual gray, suggesting that the weathered till is widespread and continuous. Moreover, in the writer's experience the glacial deposits everywhere in Inglefield Land appear to be weathered about as much as the till in this section. He therefore thinks that the till in the section was laid down during the last major advance of the Indlandsis and was weathered *in situ* after deglaciation.

The depth of the weathering is presumably greater than the depth to permafrost. Perhaps some of the weathering took place during the amelioration of climate that characterized the climatic optimum, when the depth to permafrost was greater than at present.

4. Lacustrine Strandlines

A series of horizontal lacustrine terraces is located not far from where the river enters the northern arm of the ice-dammed lake (fig. 18). These terraces extended up to 34 feet above the level of the lake on August 8, 1953 and 35 feet above the level of the lake on August 9, 1963. They are hundreds of feet long, a few feet wide, and a few feet apart, one below another. They are found on the ice-contact slopes of several kame terraces and on the sides of one morainal cone. That they were formed in a very short time is proved by their small size and by the fact that they were formed in summer on unfrozen, unconsolidated kame terrace deposits and till. Their small size, and the fact that they have been cut on a slope formed on unconsolidated material, mean that they are ephemeral features.

These lacustrine terraces were formed recently by the ice-dammed lake when it stood higher and long after the kame terraces and morainal cones had been deposited. This is proved by the following: *First*, the sand and gravel in the kame terraces and the till in the morainal cone are stained yellow for more than 2 feet below the surface, indicating that these deposits have considerable antiquity. Because the lacustrine terraces are ephemeral, it does not seem likely that they would have persisted for as long a time as was necessary to weather these deposits. *Second*, the flat surfaces of the kame terraces have numerous frost cracks, and cracks are also found on the kame terrace slopes that lack lacustrine terraces. The cracks do not cross the lacustrine terraces, however, because the terraces were formed subsequent to the cracks. The lacustrine terraces are therefore younger than the frost cracks and still younger than the kame terraces. *Third*, some of the kame terraces have been considerably eroded by running water. The lacustrine terraces, however, have suffered only an insignificant amount of stream erosion although they are lower topographically than the eroded kame terraces and therefore have larger ephemeral streams running next to them. Their lack of erosion is due to the fact that they are youthful features, formed much later than the kame terraces.

The lacustrine terraces have resulted from a recent readvance of the Indlandsis. When the kame terraces and morainal cone were formed, the Indlandsis was 2–3 miles more advanced than it is now. Following this, it retreated an unknown distance until it reached a position behind its present terminus. After the weathering profile had been developed on the kame terraces and morainal cone, the Indlandsis readvanced to a point somewhat beyond its present position, and the highest lacustrine terrace was formed. After its formation, a recent minor retreat of the Indlandsis lowered the lake, successively lower and lower lacustrine terraces were formed, and finally the present position of the ice-dammed lake was established. It is not known whether the ice front is still retreating.

5. Patterned Ground

Well-developed patterned ground is found in places at the edge of the ice front. It is similar to patterned ground found elsewhere, and the polygons are not elongated parallel to the ice front as would be expected if they had been formed immediately in front of it. Patterned ground, therefore, appears to have been overridden and buried by a readvance of the Indlandsis. The presence at the ice margin of fluvial and/or lacustrine deposits which incompletely bury non-sorted circles also suggests a readvance of the Indlandsis.

In addition, the absence of a large marginal channel(s) in the mantle rock deposits adjacent to the ice front near the reentrant in the ice margin (fig. 17), and the fact that the meltwater from the glacier is not overloaded, suggest that the glacier has been at its present position for only a short time.

Historic Retreat of Glaciers in Northwest Greenland

Field investigations and photographic interpretation indicate that in North Greenland during the last 68 years the ice margins of scores of glaciers and local ice caps have retreated (DAVIES and KRINSLEY, 1962; WRIGHT, 1939). Thus it is not surprising to find evidence for glacial retreat in Inglefield Land. The present ice-dammed lake (fig. 18) is cutting small clifflets and terraces on the glacial deposits along its margin. Above these lacustrine features, however, similar features extend 20–25 feet higher. All of these strandline features are ephemeral and could be historic, as they are cut on unconsolidated deposits. In any event, an analysis of the outlet of the lake suggests that a minor retreat of the Indlandsis could account for the lowering of the lake.

Early Deglaciation of Inglefield Land

A problem of interest to geomorphologists is why Inglefield Land is completely deglaciated when the lands around it are still covered by ice. No definitive answer can be given at present because not enough is known about the meteorology and bedrock topography of the region. It is obvious, however, that this early deglaciation must be due in part to local factors.

It seems likely that the bedrock, at equal distances from the coast, is lower beneath the Humboldt Gletscher than in Inglefield Land (World Aeronautical Chart, Smith Sound, Canada-Greenland, No. 20, 1952). Also, an east-west seismic profile across Prudhoe Land indicates that Inglefield Land in general is higher than the ice-covered bedrock of Prudhoe Land.[1]) Therefore the ice sheet that formerly covered Inglefield Land was probably always thinner than that which covered Prudhoe Land and the area now occupied by the Humboldt Gletscher. This facilitated the early deglaciation of Inglefield Land.

Nothing is known of the position of bedrock beneath the Indlandsis southeast of Inglefield Land. If there were an ice-covered bedrock ridge extending in a southwest-northeast direction, more or less parallel to the southeast margin of Inglefield Land, and running from the southwest side of the Humboldt Gletscher to the reentrant in the ice front on the northeast side of Prudhoe Land, it would restrict the flow of ice into Inglefield Land and could be a very important factor in its early deglaciation (figs. 1, 5). A threshold mechanism similar to this one has been used in Antarctica to explain certain anomalous deglaciation patterns (Nichols 1966, p. 16–18).

[1]) Personal communication from Mr. Jean Jacques Holtzscherer.

PERIGLACIAL FEATURES

Permafrost

On July 22, 1953 the depth to frozen ground in brown weathered till at three different places, all within a mile of the reentrant in the Indlandsis (figs. 17, 18), was 30, 22, and 18 inches. Differences in the moisture content of the till and in the snow cover were probably in part responsible for the differences in depth. Because the end of the summer thaw does not occur until early in August, it seems likely that the depth to permafrost in this area and terrain is between 2 and 3 feet.

Solifluction

In Inglefield Land, as in other parts of the Arctic (Washburn, 1947, p. 88–95; Nichols, 1953b, p. 270–272), solifluction deposits are widespread and very common (fig. 33). They are derived mainly from till and perhaps from loess; but beach deposits, deltas, and glacio-fluvial deposits are generally not sources. The margin of a small active solifluction lobe is located in places at the bottom of a cliff on the south side of the Minturn Elv near its mouth. The talus that falls from the cliff onto the surface of the lobe is carried downslope by it. Depending on the source, some solifluction deposits are composed mainly of fine-grained material while others contain an abundance of large fragments. Some are boulder banked. Solifluction deposits are not found where the topography is flat nor where the mantle rock is dry. They bury till, terrace gravels, and deltaic and beach deposits. Because the higher beaches are buried by solifluction sheets, it is commonly impossible to determine the marine limit. Vegetation more or less completely covers some deposits but little vegetation is found on others. Vegetation apparently grows on those which hold moisture because they are composed of an abundance of fine-grained material. The moisture which gives them mobility comes from the active zone and also from the meltwater of snow patches. The extreme fluidity, during the summer months, of the material composing some of the solifluction deposits makes it easy to see why movement takes place on gentle slopes. Solifluction sheets, lobes, and clifflets are common. The sheets and lobes are commonly less than 5 feet thick but some are thicker, as clifflets as much as 12 feet high occur. Some steep slopes, when seen from a distance,

Fig. 33. Solifluction sheets similar to this one near the mouth of the Minturn Elv are common in Inglefield Land.

are characterized by a series of lines parallel to the slopes. The furrows made at the contact of adjacent solifluction lobes may be responsible for these lines. A solifluction lobe was noted with low marginal levees and a central channel similar to those found on mud flows. The material in the channel was finer-grained than that in the marginal levees. It probably retained its moisture and consequent mobility longer, and drained away after the coarser, drier material in the levees had become inactive. Fragments on solifluction deposits with travertine on their upper surfaces demonstrate overturning.

A gentle, continuous, uniform, smooth slope, half a mile or more in length and with an inclination of a few degrees, is located a few miles northeast of the source of the Tufts Elv (figs. 17, 18). As seen from a distance, neither hillocks nor hollows appear on it. Although the writer did not have a chance to study it, he does not believe it is a till slope reflecting a gently-sloping bedrock surface below. The irregularities of the bedrock where exposed, and the abundance of solifluction deposits in this part of Inglefield Land, support this conclusion and suggest that the slope has been graded by solifluction. A smooth, well-graded solifluction slope is found at the foot of the morainal ridge shown in figure 19; and similar slopes occur in other places. At a distance these solifluction slopes look like the alluvial slopes of Nevada, California, and elsewhere.

Sorted and Nonsorted Circles and Polygons

Nonsorted circles and polygons are very common. They are less than a foot high and from a few feet to more than 20 feet in diameter. They are found on till plains, solifluction deposits, and beaches, and nestled between the fragments of block fields. They are commonly composed of wet, very fluid material. In places they make it difficult for anyone on foot, because unless he walks very rapidly across them he will sink in over his shoes. They were therefore avoided whenever possible. On beach deposits they bring up smooth unweathered limestone roundstones that are very easy to differentiate from the pitted and ridged weathered limestone fragments generally found on the surfaces of the beaches. The presence of nonsorted circles and polygons on beaches suggests that the beach deposits are not thick and that immediately below them there is either till or fine-grained offshore marine deposits.

Sorted circles and polygons and striped patterns are also common. They appeared to be similar to those found elsewhere.

Frost-Crack Polygons

Frost furrows similar to those described by Leffingwell (1919, p. 205–212), Washburn (1947, p. 102–103), Lachenbruch (1960, p. B404–B406), and others, were seen in many places. They are commonly 3 or more feet wide and up to 2 feet deep. They extend for scores of feet in more or less straight lines, and their intersection one with another results in frost-crack polygons some of which are more than 100 feet across (figs. 6, 25). They are found on horizontal ground and also on slopes up to 15–20 degrees. Almost all of them occur where there is no vegetation, but a few were seen in tundra. The furrows are found on kame terraces, valley trains, dissected deltas, alluvial fans, modern flood plains, and solifluction and beach deposits; but they were not seen on the broad expanses of till that are so common in Inglefield Land. It is generally agreed that they have resulted from the cracking of frozen ground with an appreciable ice content, due to thermal contraction during periods of intense cold (Washburn, 1956, p. 850–851). If frost cracks are formed in this way, they would naturally be better developed in well-sorted material with high porosity, like sand and gravel, than in till, which has a lower porosity because it is poorly sorted and generally contains large boulders. Studies in Alaska, Antarctica, and elsewhere have shown that beneath the furrows there are either ice wedges (Lachenbruch 1960, p. B404–B406) or sand wedges which have progressively filled the contraction cracks (Péwé, 1959, p. 545–552). The furrows are due to the

Fig. 34. A frost furrow with marginal levees on a fan near Dallas Bugt. Between the levees is a miniature furrow formed in windblown sand by the trickling of sand down into an open contraction crack.

melting of the ice wedges or to the falling of surficial material into the contraction cracks.

Many of the furrows in the area around Thule have been nearly filled with wind-blown sand so that only a slight surface sag remains. Holes 4 feet long, 1 foot and more in depth, and 1 foot across are found in some of these furrows. They may be deflation pits, small thermokarsts resulting from the melting of ice wedges, or holes formed by the trickling of the sand down into contraction cracks.

The surface of the ground adjacent to the furrows has not, in general, been significantly deformed. However, levees are found along both sides of some furrows (fig. 34). The levees may be 2–3 feet high, and the width from the outer edge of one levee to that of the other may be as much as 20 feet. The roundstones on levees on elevated beaches are devoid of lichens, whereas the roundstones on other parts of the beaches are covered with lichens. A thin veneer of wind-blown sand is commonly found in the furrows between the levees. Less commonly, miniature V-shaped trenches a few inches wide, the sides of which are at the angle of repose, may be found in the wind-blown sand. They result from the trickling of the wind-blown sand down into open contraction cracks. They prove that the furrows and levees are being formed at the present time, as does the absence of lichens on the roundstones on the levees. If a contraction crack extends through the active zone and down into the permafrost, a sand wedge similar to those described by Péwé (1959) will be formed. The levees are formed because the material in the polygons between the contraction cracks, which has progressively increased in volume during the growth of either ice wedges or sand wedges, expands during the warm season and causes the deposits next to the contraction cracks to be deformed and pushed upwards (Lachenbruch, 1960, fig. 186.2).

Miniature holes and tunnels, small cracks in the ground, and collapsed grass, associated with the wind-blown sand found in some of the furrows, also prove that the formation of the furrows is a current process.

Block Fields

Large block fields are very common; they were encountered in places too numerous to mention. Good examples are located in front of the Indlandsis about 2 miles northeast of the reentrant in the ice front, along the front of the Indlandsis southeast of the ice-dammed lake, in the area between Gevirelven and Rygelven, on the upland between the two rivers that empty into the ice-dammed lake, and elsewhere (figs. 17, 23). They are difficult to negotiate on foot and were therefore avoided whenever possible.

These block fields are characterized by (1) large blocks, some of which are 12 feet long. (2) Fine-grained nonsorted circles, covering only a small fraction of the block fields, are nestled between blocks, indicating till at depth. (3) Granular disintegration has removed as much as an inch of material from the corners and edges of some blocks. (4) The block fields are found on more or less flat topography. (5) They are found on the basement complex and not on the sedimentary rocks.

Five theories for the origin of these block fields will be considered.

First, they are areas of very blocky till. Subglacial till deficient in fines and high in blocks was formed because of a widely-spaced joint system in crystalline bedrock. Superglacial moraine is found near the edge of the Indlandsis where it has been brought to the surface along shear planes. In places it consists of only a thin veneer of angular blocks about the same size as many of the fragments in the block fields. The deposition of this material on blocky subglacial till would produce areas similar to many of the block fields.

Second, they are areas of blocky till on which there are numerous fragments which were brought from depth by frost action (Taber, 1943, p. 1452–1456). It is recognized, of course, that unless the active zone has been much thicker in the past than it is now, this process would not be important in forming the block fields (Laursen, 1950, p. 110.)

Third, they are areas of blocky till from which meltwater has removed the fines, leaving a concentration of blocks at the surface.

Fourth, they have resulted from frost-shattering of both bedrock and blocks after the area was deglaciated. If, on deglaciation, numerous bare bedrock knobs surrounded by lower areas covered with till were present, post-glacial frost-shattering might break up the bedrock outcrops so that fragments might slide out over the till and bury it. Continued frost-shattering might destroy the knobs, and the bedrock platforms formed from them might be concealed by angular fragments. Non-sorted circles, composed of material derived from the till and of different composition from that of the frost-riven fragments, might form between them. Fragments with both weathered and unweathered surfaces, angular fragments with sharp edges and corners, the common occurrence in fragments of open cracks with unweathered walls, and the fact that adjacent fragments can commonly be fitted together, all prove that frost-shattering is a current process. The block fields have been carefully studied by Mr. Richard W. Lemke, U.S. Geological Survey. He believes that some of them have resulted from frost acting on bedrock outcrops after the area was deglaciated.[1])

Fifth, they may have resulted from a combination of these processes.

As these block fields have probably been formed in a number of different ways, each one will have to be studied individually in order to determine its origin. However, it seems likely that some of them are areas of till which were originally blocky.

Debris-Capped Fragments

Fragments capped with unsorted material are common. This material, which ranges in size from silt to small boulders, may completely cover the tops of the fragments or may form only a thin scattered veneer. The debris may be a foot or more above the ground surrounding the fragments. Debris-capped fragments are found on both flat and sloping terrain.

Three theories for the origin of the debris will be considered. Where the blocks are found on till plains, it might be supposed that the debris was deposited by glacial action. Because the fines can be removed quickly and easily from the tops of the fragments by wind and perhaps by rain, it is certain that the debris veneers have no antiquity. Thus the debris

[1]) Personal communication from Mr. Richard W. Lemke.

on fragments that are at a considerable distance from the ice margin was not formed in this way. In many places the debris-capped fragments are surrounded by solifluction lobes and sheets. Debris could have been deposited on a fragment when it was overrun by a solifluction lobe. The surface surrounding the fragment might later have been lowered by continual solifluction so that finally the debris came to be some distance above the surface of the ground. This mechanism does explain the cap on some fragments in the Thule area. Here block-banked terraces are common. Solifluction lobes running down slopes have been dammed up by blocks and small terraces have been formed. These terraces may be 4 or more feet high, 20 or more feet long, and several feet wide. In some cases the top of the terrace is considerably below the top of the block, in others it is level with it but no material has been deposited on the block, and in still others the top of the block has been completely overrun by the lobe and a debris cap has been formed. In Inglefield Land the debris-capped fragments that are surrounded by flat terrain were probably not formed in this way because solifluction lobes and streams do not form on flat terrain and because it would be difficult to subsequently lower the surface around the blocks. These debris-capped fragments are commonly surrounded by flat nonsorted nets. Although these nets bury low-lying fragments and also deposit a thin veneer of unsorted material on fragments which stand somewhat higher above the surface, they do not as a rule grow high enough to cover the tops of the fragments under consideration. These debris-capped fragments may once have been buried by unsorted material and may have later grown upward because of frost action (Taber, 1943, p. 1452–1456).

Earth Hummocks

Small round earth hummocks cover acres of ground in places. They may be a little over a foot in height and as much as 2 feet across at the base. They are closely spaced, the distance between them being considerably less than the width of one hummock (fig. 35). They are fatiguing to walk on, as the tops are rounded and it is hard to avoid slipping into the spaces between them. The sides are steep, some vertical or even overhanging. The hummocks are more or less completely covered with a veneer of vegetation which in general is less than an inch thick. Their shape indicates that they are composed of fine-grained material which has considerable cohesion; digging showed that this material is loess. The steep sides require explanation. The hummocks are not frozen in the summer, and it seems unlikely that the thin veneer of vegetation is responsible, as it is absent in some places. Perhaps the cohesion which makes the steep sides possible is due to the carbonate content of the loess,

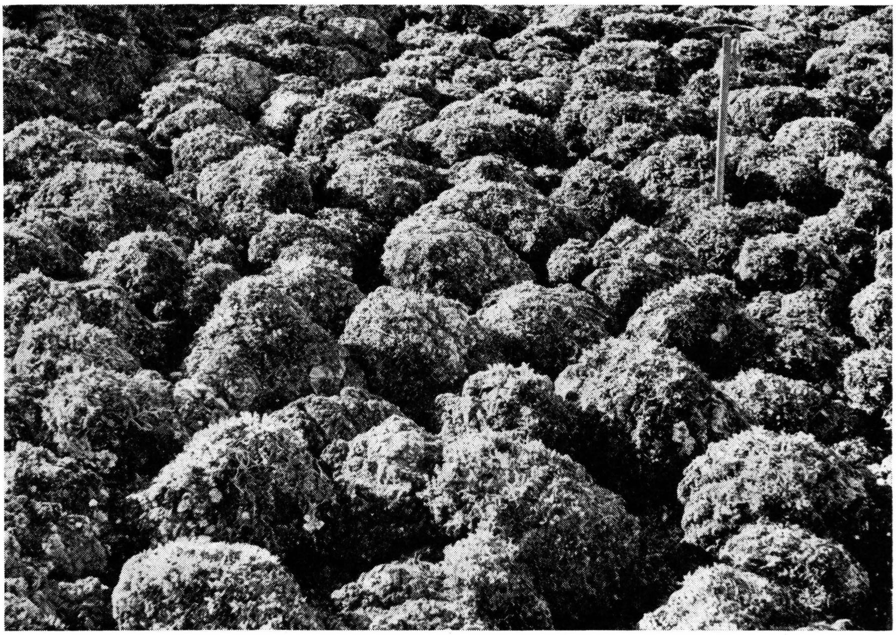

Fig. 35. Closely spaced earth hummocks. They are commonly a foot in height, 2 feet across at the base, and composed of loess. Ice axe gives scale.

perhaps also to moisture. Because limestones are common in Inglefield Land (Troelsen, 1950, geol. map) it seems certain that the loess contains calcium carbonate as does that described by Böcher (1949, p. 26) from the Søndre Strømfjord area. The hummocks occur both on flat areas and on slopes in excess of 15 degrees. Although they were found on "Black Rock", on the terraces immediately upstream from the gorge of the Tufts Elv, on a terrace slope cut near the head and on the west side of the dissected delta at Rensselaer Bugt, and elsewhere, they are not common. Their distribution is perhaps controlled by the distribution of the loess. They are found in Iceland; in St. Elias Range, Yukon Territory, Canada; at Resolute Bay, Cornwallis Island, N.W.T.; in other parts of Greenland, and elsewhere (Sharp, 1942, p. 282–284; Washburn, 1956, p. 830–831; Nichols, 1953b, p. 275). They are thought to be due to differential freeze and thaw (Sharp, 1942, p. 297–299).

Miniature Nivation Cirques

Although all of Inglefield Land is below the climatic snowline, perennial snow patches are common. They are in general less than 100–200 feet long and are inset a few feet into the topography on which they rest. These miniature nivation cirques (Flint, 1947, p. 94; Nichols, 1963a)

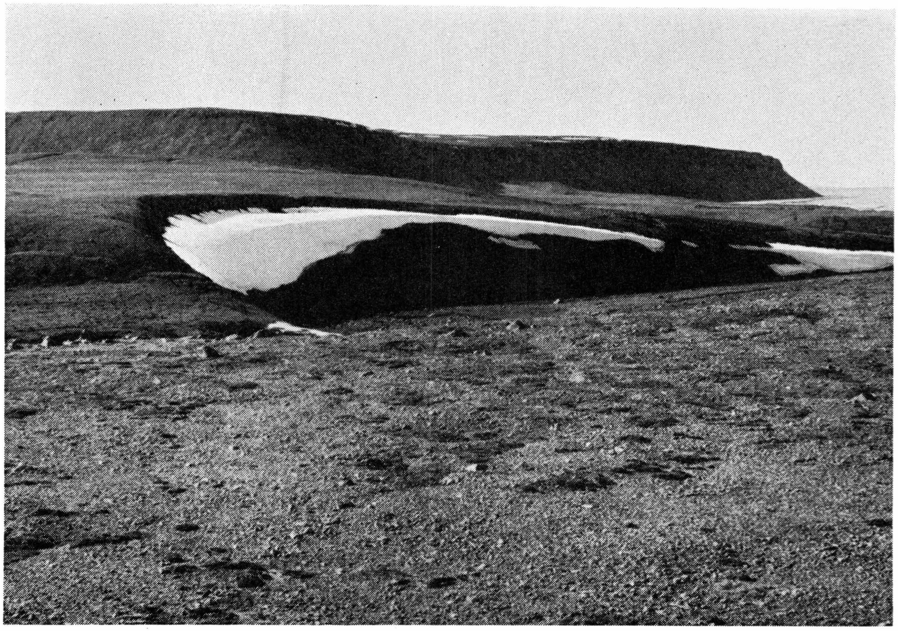

Fig. 36. A small nivation cirque inset a few feet into an elevated delta near Dallas Bugt.

are found on till and glacio-fluvial deposits (fig. 36). Small sand flows with channels, levees, and lobate termini superimposed one upon another, together with small fans, are found both below and marginal to the snow patches. They are formed by meltwater from the snow patches. The ground immediately below the snow patches may be soft and soupy during the summer months. Here the meltwater from the snow patches facilitates solifluction. Nivation is an unimportant process at the present time in Inglefield Land; the volume of the material moved is insignificant and the distance it is moved is generally only tens of feet. Nivation may have been more important earlier in post-glacial time when the snowline was lower.

Weathering of Rock

In general the bedrock is fresh and unweathered. In some places, however, a small amount of granular disintegration has occurred. At one place not more than a mile from the Indlandsis the rocks were rough and crumbly and some residual mantle had formed. Fretwork with a relief of 2 inches was seen on a garnetiferous granite. The fragments on some of the block fields have been somewhat rounded by spalling. The older surfaces on the fragments tend to be lighter in color, whereas the newer surfaces, exposed by spalling, are darker. Some of the fragments have been roughened on top by weathering but have remained smooth underneath because of the lack of it.

The writer knows of no studies made to assess the relative importance of physical and chemical weathering in northern Greenland. However, recent work on weathering in a part of Antarctica where the climate is about the same as in Inglefield Land indicates that chemical weathering is not important (KELLY and ZUMBERGE, 1961).

WORK OF WIND

Wind-Blown Sand

Although most of Inglefield Land is devoid of vegetation, the work of wind is of very minor importance. This is probably due to: (1) a scarcity of sand because of extensive felsenmeers and till plains, (2) a winter snow cover, (3) the irregularity of much of the terrain, (4) dampness during summer months.

Wind-blown sand, in the writer's experience, is almost non-existent (MALAURIE, 1955, p. 212). Small quantities hardly worth mentioning are found: (1) in a gully incised in a sandy terrace cut in the uplifted dissected delta at Rensselaer Bugt, (2) in the shadow dunes associated with the boulders on the surface of this terrace, (3) in the frost cracks on this terrace, (4) in an eolian veneer derived from sandy alluvium deposited by a stream adjacent to the Indlandsis, (5) in eolian sand sheets veneering modern kettled fans at the foot of cliffs between Dallas Bugt and the river immediately north of it, (6) in the small dunes, a few feet high, covering a few acres, which are adjacent to an ice-dammed lake (figs. 17, 18). The lake level drops between 15–20 feet late in the summer because of the partial removal of the snow and ice dam by the outlet stream. This lowering uncovers the sandy lacustrine sediments from which the dunes were derived. The raised marine beaches are not sources of eolian sand, being composed principally of gravel. Small dunes several feet high are found along the rivers and coast in Peary Land, but they are not common. Still larger dunes are present around Søndre Strømfjord (FRISTRUP, 1952–53, p. 60).

Loess

Fine-grained yellow weathered material is found in many places along the Tufts Elv. It veneers "Black Rock" (figs. 17, 19), the slope adjacent to it, and the nearby terrace. It is found approximately 100 feet above the river near the upper end of the gorge. Quantitatively it is of little importance, although on the slope adjacent to "Black Rock" it is about 2 feet thick. It is intermediate in size between fine sand and clay. This is shown by the following: (1) If thrown into the air when dry, it is blown away by the wind. (2) It has some cohesion when damp and will retain for a time the shape into which it has been molded. (3) It cracks on drying. The fact that it is weathered suggests some antiquity for the deposit. The following prove that it is loess: (1) It is too well-sorted and fine-grained to be a stream deposit. (2) It is found above the highest stream terraces and in places where post-glacial streams have not existed.

Its proximity to the valley train, into which the terraces have been cut, proves that it was derived from the valley train. During the winter months, when there was little or no meltwater and when the valley train sediments deposited during the previous melt season were dry and not covered with snow, the wind winnowed the loess from the valley train. The presence of loess in the gorge of the Tufts Elv, where no terraces are now found, probably indicates that they were once present but have been removed by the river. No loess was found near the extensive uplifted dissected delta in Rensselaer Bugt, although before dissection this must have been one of the most extensive areas of outwash in Inglefield Land. The orientation of the ventifacts on the delta suggests that most of the fine-grained material winnowed from the topset beds of the delta was probably blown out into the bay. Detailed field work would probably reveal loess which had been blown from the delta on to adjacent areas. Böcher (1949, p. 25–28) has described the loess in the Søndre Strømfjord area.

Ventifacts

Numerous well-developed ventifacts are found in the eolian area close to the ice-dammed lake. Many fragments which are not buried by the dunes have been wind-cut as much as 4 feet above ground level but only on the side that faces the lake and the glacier beyond. The leeward surfaces of fragments not submerged by the high stage of the lake are stained by limonite; those surfaces which are submerged are veneered with yellow lacustrine silt. The windward sides are gray, as the yellow lake sediments and any limonite that is formed are annually removed by sand-blasting. This difference in color is so pronounced that it is easily recognized.

Wind-cutting takes place during the summer months; coarse sand was being moved on August 8, 1963 when the writer visited the area, and wind velocities in excess of 40 miles per hour occurred a few miles from this area on August 9, 1963. It is not known how effective eolian action is in this area during the winter months, as nothing is known about the thickness and continuity of the snow cover at that time.

The location of the dunes and the distribution of the wind-cut surfaces prove that winds which blow off the glacier, in part perhaps kata-

Fig. 37. Wind-eroded frost mounds.

batic, are responsible for the eolian features. The fact that the blocks are cut on only one side proves that they do not move.

Numerous ventifacts — not spectacularly developed, however — are found on one of the terraces cut in the uplifted dissected delta in Rensselaer Bugt. Kanters, pits, cups, and polishing are present. Several inches have been removed from some fragments, and cutting has taken place as much as 1–2 feet above ground level. The windward sides of the ventifacts are clean and devoid of lichens, whereas the leeward sides are stained and covered with lichens. Thousands of shining fragments may be seen when the observer looks towards the sun. Most of the wind-cut surfaces are not highly polished, perhaps because material smaller than the sand sizes is not abundant. Wind-cutting is favored by the fact that the terrace is old, sandy, dry, flat, and without vegetation. The ventifacts, the small shadow dunes on the terrace, and the presence of eolian sand in a gully cut on the Kane Bassin side of the terrace, prove that here the wind blows principally from Inglefield Land out into Kane Bassin.

Wind-polished and wind-cut stones are very common in parts of Peary Land (Troelsen, 1949, p. 22–23; Fristrup, 1952–53, p. 56–60) and are also well-developed in Søndre Strømfjord (Jensen, 1889).

Wind-Eroded Earth Hummocks

At several places, earth hummocks have been truncated so that the original upper curved surfaces have been removed and flat erosional surfaces have been formed (fig. 37). Parallel grooves are found on the ero-

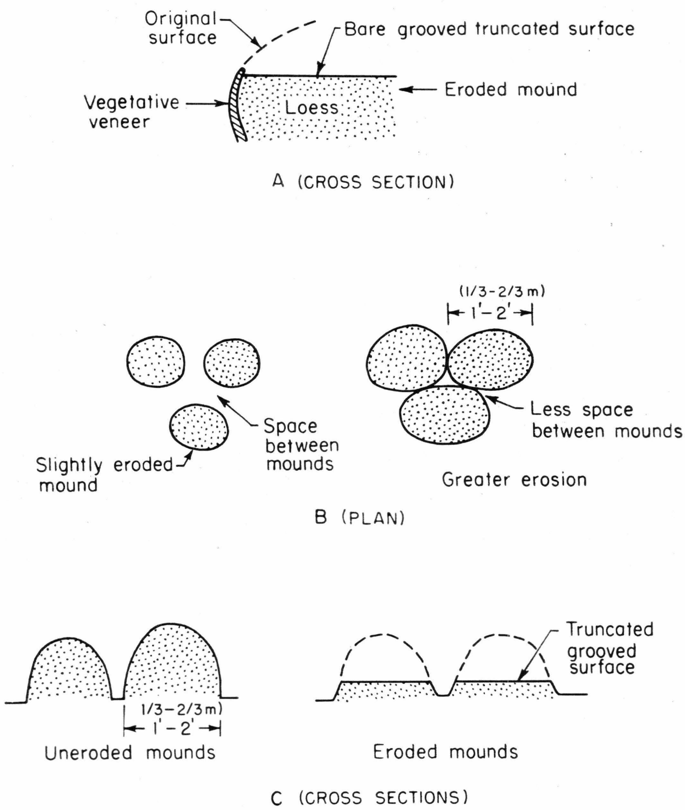

Fig. 38. Cross-sections and plan views of wind-eroded frost mounds.

sional surfaces. Small undercut clifflets less than an inch in height are also present, and occasionally a granule has protected the loess in its lee. A rim of vegetation which was resistant to erosion commonly surrounds the truncated surface (fig. 38). The erosional surfaces studied were devoid of vegetation, although vegetation is found on the sides of the truncated mounds. Apparently the hummocks were covered with vegetation at the time the flat surfaces were cut, but due to the recency of their formation no vegetation has established itself on them. In places many inches of loess have been removed. Where dissection has been deep, the flat surfaces are not separated by wide spaces but are almost in juxtaposition. They resemble truncated pillow lava.

A preliminary analysis of these features suggested that wind or sheet wash formed them, as it seemed most unlikely that they had resulted from the sag or slump of wet viscous material. The following facts, however, prove that they were formed by wind erosion: (1) The rainfall of Inglefield Land is not great, and presumably heavy rain is not common. (2) These truncated surfaces are found on areas so flat that it seems unlikely that sheet wash could be an effective erosive agent. (3) In places the parallel grooves on the truncated surfaces are inclined to the steepest slope. (4) The undercut clifflets are similar to the undercut slopes on yardangs. (5) The spaces between the hummocks are not eroded, nor are they filled with material from the hummocks. It is not known whether the presence of frozen ground or dampness prevented further erosion and was responsible for the flatness of the erosional surfaces.

The sequence of events resulting in the formation of the wind-eroded earth hummocks is as follows: (1) deposition of loess; (2) formation of mounds; (3) mounds covered by vegetation; (4) mounds truncated by wind erosion.

Distribution of Lichens

Lichens are found on the fragments in the till and fluviatile boulder pavements near the place where the outlet stream from the ice-dammed lake flows northward from the margin of the Indlandsis (figs. 17, 18). They do not, however, uniformly cover the exposed parts of the fragments. They are common on the side away from the glacier, but are usually absent on the side toward the glacier. Because of this distribution of lichens, the deposits look fresh and youthful when viewed looking away from the Indlandsis, and weathered and much older when viewed looking toward the Indlandsis. It seems unlikely that this distribution has resulted from differential moisture or insolation. The side toward the glacier is probably blasted by wind-driven snow, ice crystals, and perhaps other material, so that lichens cannot grow here although they can grow on the lee side. The lichens found near the ground on some fragments on the side facing the glacier may be due to protection by low snowdrifts.

Eolian Micro-Topography

Following two days of strong winds, small-scaled topographic features were seen on the fine-grained material between the blocks on till plains. Ridges an inch or more in length separated by valleys $^1/_8''$ to $^1/_4''$ in depth and width were seen over an area of several square miles. The valleys and ridges were all oriented in the same direction. The valleys were formed by erosion and the ridges were residuals between them. The valleys commonly started from nothing, progressively increased in width and depth, and terminated in small vertical or overhanging clifflets. The following prove that they resulted from wind erosion: (1) they were seen after strong winds; (2) they were not found where standing or running water was present during the wind storm; (3) they were everywhere parallel to the prevailing wind direction as indicated by the presence of small shadow dunes behind the blocks. They are obviously ephemeral features.

Fig. 39. Surface efflorescence near Marshall Bugt. It is white, soluble, and approximately $^1/_8$ inch thick. It veneers fine-grained sediments and is mainly sodium chloride. Photograph by CHRISTOPHER G. MILLER.

EVIDENCE FOR ARIDITY

Meteorological data, surface efflorescences, deposition of travertine from small streams, the presence of travertine on the bottoms of fragments, and solution-facetted and solution – shaped fragments all prove that the climate of Inglefield Land has been arid for thousands of years (TEDROW, 1966).

Surface Efflorescences

Small patches of surface efflorescences covering hundreds of square feet were seen in Inglefield Land, although they are not common (fig. 39). They are white, soft, soluble, bitter, and approximately $^1/_8''$ thick, and they veneer fine-grained sediments. They have not been formed by eva-

Fig. 40. Banded efflorescences on a steep slope cut in a dissected delta. The bands are a few inches wide and scores of feet long.

poration of standing water, as they are found on sloping ground. Their thinness makes one wonder if they persist throughout the year or whether they are dissolved during the wet season and re-precipitated during the dry season. At Rensselaer Bugt the efflorescences are found in more or less parallel and horizontal bands on a steep slope cut into an uplifted dissected delta. The bands are a few inches wide and scores of feet long (fig. 40). The delta is composed of layers of sand and silt. The efflorescences cover the silt. Professor Cornelius S. Hurlbut, Jr. of Harvard University, using both the petrographic microscope and the X-ray diffractometer has identified NaCl from a patch near the head of Marshall Bugt. The efflorescence did not taste like pure NaCl to the writer. If any other soluble salt is present, however, it must, according to Professor Hurlbut, be in quantities less than five per cent by weight of the NaCl. Here, as at Rensselaer Bugt, the efflorescences are associated with deltaic sediments; perhaps the source of the salts is connate water or, if not, they may have been wind-blown from the ocean.

Thick efflorescences composed of chlorides and sulphates of magnesium and calcium are found in Peary Land, where they occur around ephemeral lakes and streams (Troelsen, 1949, p. 23; Fristrup, 1952, p. 92). Böcher (1949, p. 33–41) reports the presence of saline lakes, saline soils, and surface efflorescences in the Søndre Strømfjord area. The sur-

Fig. 41. Solution-shaped beach stones on an elevated beach. Grooves, ridges, saucer- and bowl-shaped hollows. Pencil gives scale.

face efflorescences are composed of the chlorides, sulphates, and carbonates of calcium, sodium, magnesium, and potassium. Gypsum efflorescences have been reported from the Thule region of Greenland (NICHOLS, 1953b, p. 270), and surface efflorescences form under similar conditions in Antarctica (NICHOLS, 1963b, p. 27–28).

Surface efflorescences are found only in arid regions, where they are formed by the evaporation of solutions which move upward by capillarity in fine-grained sediments.

Solution-Shaped Limestone Fragments

The limestone fragments on the older raised beaches have suffered considerable solution. Commonly as much as an inch of material has been removed from the unburied, upper surfaces. Solution-formed surfaces characterized by saucer- and bowl-shaped hollows and pits separated by sharp-edged divides have been developed (fig. 41). The hollows are usually less than an inch in diameter and half an inch in depth (LAHEE, 1931, p. 32). The divides are so sharp that they rapidly wear out the soles of shoes. On steep faces the pits and hollows are replaced by grooves separated by sharp, narrow ridges. They are commonly a few inches long, and individual grooves and ridges may be a quarter of an inch in width and depth. Although significant thicknesses of limestone have been removed from the upper, unburied surfaces, little or no material has been removed from the buried surfaces; indeed, travertine is sometimes deposited on them. Where the unburied surfaces of limestone roundstones have been truncated, solution-facetted pebbles similar to those described by BRYAN (1929) have been formed.

Similar pits and hollows have been developed under a variety of conditions. They have been formed in lakes under water (BELL, 1895; HUDSON, 1910) and subaerially in regions with approximately 9 inches of rainfall (SMITH and ALBRITTON, 1941, p. 64), and it seems likely that they would form on bare limestone in humid climates. Those in Inglefield Land, however, could have developed only in an arid climate, as the soil-buried surfaces have suffered little or no solution. In a humid climate, the buried surfaces would have been considerably corroded, perhaps as much or even more than the unburied, upper surfaces.

The absence of organic acids formed from decaying vegetation, the scarcity of rainfall based on other evidence, and the loss of much of the snow cover by evaporation suggest that these features may have taken thousands of years to form and that therefore the aridity, as at Thule, has had considerable duration (NICHOLS, 1953b, p. 269–270).

Travertine on Subsurface Portions of Fragments

A thin coating of travertine veneers the buried portions of many fragments not composed of limestone. The veneer is formed mainly by precipitation due to evaporation of water that moves upward in the mantle rock by capillarity. Pre-Paleozoic and Paleozoic limestones are found in Inglefield Land, much of the mantle rock is therefore composed partly of limestone, and this limestone is the source of the travertine. Travertine veneers do not form in humid climates, where the water in the mantle rock generally moves downward and the mantle rock rarely dries out. They are common, however, in arid and semi-arid regions (BRYAN, 1929, p. 194–195; SCOTT, 1947, p. 149; MILLAR, TURK, and FOTH, 1958, p. 437; NICHOLS, 1963b, p. 26–27). Moreover, it is well known that while the zone of accumulation of calcium carbonate in soils is at or near the surface of the ground in arid climates, it increases in depth with increasing precipitation and disappears with still greater precipitation (MILLAR, TURK, and FOTH, 1958, p. 140). These veneers indicate, therefore, that arid conditions have existed in Inglefield Land during the period of their formation.

Travertine also veneers the buried portions of many limestone fragments on both till plains and elevated beaches. This veneer on limestone fragments is thicker than on non-limestone fragments; it may be as much as half an inch thick.

Fig. 42. The white material is calcium carbonate precipitated on fragments near Rensselaer Bugt by evaporation of meltwater.

Some of the travertine came from the limestone fragments to which it is attached, as indicated by solution-formed grooves, ridges, pits and hollows on the upper, unburied surfaces of the fragments, and by an analysis of the processes taking place in the soil beneath them. It is not known whether the greater thickness of travertine on limestone fragments is due mainly to more abundant limestone in the mantle rock or to the fragments themselves being sources of travertine.

Travertine on Unburied Surfaces of Fragments

Travertine thinly veneers the fragments that project above the surface of a shallow, slow-moving stream near Dallas Bugt (fig. 42). When seen by the writer on July 19, 1965, the veneers were parallel to the surface of the stream and extended in general an inch or so above it but hardly at all below it. The stream was not saturated with calcium carbonate; if it had been saturated, the travertine would have covered all submerged parts of the fragments.

The travertine was formed mainly by the evaporation of water (1) drawn up from the surface of the stream onto the sides of the fragments by capillarity, (2) washed up onto the sides of the fragments by small waves, (3) left on the sides of the fragments because of a fluctuating

Table 7. *Selected stations for North Greenland and adjacent Arctic Canada*

Station	No. Lat.	W. Long.	Years	*Mean Precipitation (inches of water)* Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Mean annual total
Alert	8230	6220	12	.19	.21	.26	.21	.36	.44	.51	.94	1.10	.64	.23	.25	5.34
Jørgen Brønlund Fjord	8210	3130	2–3	.05	.10	.05	T	.05	.30	.21	.22	.69	.49	.20	.05	2.41
Buchanan Bay*)	7855	7521	3–4	.14	.35	.09	.21	.22	.12	.34	2.04	.45	.82	.58	.04	5.40
Etah*)	7822	7242	1–2	.20	.21	.16	.41	.22	(.07)	—	—	(.30)	(.27)	.47	.28	—
Eureka..................	8000	8556	15	.11	.07	.07	.06	.11	.11	.61	.41	.42	.26	.09	.08	2.40
Inglefield Bay	7726	6630	4	.11	.08	.29	.12	.22	.33	.74	.72	.33	.12	.19	.08	3.33
Lady Franklin*)	8144	6445	2–3	.42	.13	.44	.17	.40	.18	.66	.38	.35	.24	.20	.30	3.88
Lake Hazen	8149	7118	1–5	(.23)	(.01)	(.02)	(.05)	(.08)	.12	.05	.25	(.20)	(.23)	(.03)	(.04)	1.31
Thule AFB (Dundas)	7632	6845	14	.24	.25	.14	.14	.21	.27	.62	.64	.56	.44	.48	.20	4.19

( ) single year.
*) combined data from more than one station in the same area.

Table 8. *Selected stations for North Greenland and adjacent Arctic Canada*

Station	No. Lat.	W. Long.	Years	*Mean Snow Depth (inches accumulation)* Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Mean annual total
Alert	8230	6220	12	16.5	18.0	18.0	16.1	14.9	0.9	0.2	2.0	9.1	13.0	14.0	15.9	11.6
Buchanan Bay*)	7855	7521	3	15.4	19.2	23.4	18.2	13.0	3.0	0.0	0.0	3.6	5.5	12.0	13.5	10.6
Etah*)	7822	7242	1	6.7	7.5	7.2	6.4	4.4	0.3	0.0	0.0	0.8	2.4	3.1	5.9	3.7
Eureka..................	8000	8556	15	6.7	7.2	7.7	8.4	7.6	0.7	0.0	0.0	2.4	4.1	5.3	6.2	4.7
Lake Hazen	8149	7118	1–5	(10.0)	(10.0)	(10.0)	(12.0)	(13.0)	0.0	0.0	0.0	(3.0)	(8.0)	(9.0)	(9.0)	7.0
Thule AFB (Dundas)	7632	6845	14	9.1	9.6	9.2	7.3	3.6	0.3	0.0	T	0.9	3.2	6.8	8.6	4.9

( ) single year.
*) combined data from more than one station in the same area.

Table 9. *Selected stations for North Greenland and adjacent Arctic Canada*

Station	No. Lat.	W. Long.	Years	*Mean Snowfall (inches of snow)* Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Mean annual total
Alert	8230	6220	12	2.0	2.2	2.6	2.1	3.6	2.4	2.6	6.1	10.9	6.4	2.3	2.5	45.7
Jørgen Brønlund Fjord	8210	3130	2–3	0.5	0.9	0.5	0.1	0.5	1.3	0.0	0.2	7.0	5.0	2.0	0.4	18.3
Buchanan Bay*)	7855	7521	3	1.4	3.5	0.9	2.1	2.0	0.8	0.0	0.0	4.5	8.2	5.8	0.4	29.6
Etah*)	7822	7242	1	1.1	0.6	2.2	3.1	0.7	0.1	T	T	2.9	2.7	5.3	2.5	21.2
Eureka..................	8000	8556	15	1.0	0.7	0.7	0.6	1.0	0.3	0.2	0.5	4.3	2.6	0.9	0.8	13.6
Lake Hazen	8149	7118	1–4	(2.3)	(0.1)	(0.2)	(0.5)	(0.8)	0.9	T	0.1	(2.0)	(2.3)	(0.3)	(0.4)	9.9
Thule AFB (Dundas)	7632	6845	14	2.2	2.9	1.8	1.6	2.2	0.5	T	T	3.6	6.5	5.0	3.0	29.3

( ) single year.
*) combined data from more than one station in the same area.

Table 10. *Selected stations for North Greenland and adjacent Arctic Canada*

Station	No. Lat.	W. Long.	Years	*Mean Temperature* (°F.) Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Mean annual temp.
Alert	8230	6220	12	−26	−28	−27	−12	12	32	39	33	15	−5	−15	−23	0
Jørgen Brønlund Fjord	8210	3130	2–3	−24	−20	−17	−8	16	38	44	39	23	−3	−12	−18	4
Buchanan Bay*)	7855	7521	4–5	−27	−25	−22	−7	17	36	40	37	23	3	−12	−23	3
Etah*)	7822	7242	5–9	−18	−19	−14	−3	22	36	41	37	25	8	−3	−9	9
Eureka..................	8000	8556	15	−35	−36	−34	−16	14	37	42	38	20	−7	−23	−32	−3
Inglefield Bay	7726	6630	5	−14	−13	−10	5	28	34	39	37	28	13	2	−11	12
Lady Franklin*)	8144	6445	3–4	−39	−40	−31	−15	12	33	37	34	17	−9	−22	−27	−4
Lake Hazen	8149	7118	1–5	(−39)	(−41)	(−39)	(−27)	8	35	44	41	(18)	(−11)	(−35)	(−48)	−8
Rensselaer Harbour	7837	7053	1–2	−32	−34	−37	−10	14	(30)	(38)	(32)	13	−4	−22	−31	−4
Thule AFB (Dundas)	7632	6845	14	−11	−14	−15	0	25	37	43	40	28	15	1	−11	12

( ) single year.
*) combined data from more than one station in the same area.

stream level. Although the fragments were heated very slightly by insolation and by the ambient temperatures, the water next to the fragments was not heated sufficiently to cause precipitation of travertine. Moreover, in such a slowly moving stream, loss of carbon dioxide due to agitation was not a factor in the precipitation (Twenhofel, 1932, p. 323). If the travertine had been precipitated by organisms, it would have veneered all submerged parts of the fragments. Solution of the travertine by rainfall, by the unsaturated stream waters, and by meltwater from local snowbanks was not sufficient to prevent it from forming a veneer. Similar deposits are found adjacent to Rensselaer Bugt and near the mouth of the Minturn Elv (fig. 42).

The factors which favor the formation of the travertine are: (1) evaporation in excess of rainfall, (2) high concentration of calcium carbonate in the streams, (3) streams which start to flow at the beginning of the melt season, progressively increase in height with the oncoming of the warm season, remain fairly constant at their highest level for some time, and then continuously decrease in height and finally freeze up with the oncoming of the cold season. Under such a regimen the travertine veneer will be formed at the high stage of the streams and will only rarely be flooded and redissolved by the unsaturated river water.

An arid climate is commonly defined as one in which potential evaporation exceeds precipiation (Thornthwaite, 1948). The presence of the travertine does not of itself demonstrate aridity. However, the formation of the travertine is definitely favored by aridity.

Meteorologic Aridity

The meteorologic data also suggest that the climate of Inglefield Land is arid (Malaurie, 1955, p. 211). Table 7 gives the mean precipitation in inches of water for several selected stations in North Greenland and Arctic Canada. An analysis of these data suggests that the mean annual precipitation in the coastal areas of Inglefield Land is equivalent to approximately 4 inches of water.

Fristrup (1952–53, p. 55) writes: "In the arid parts of Peary Land the evaporation is very pronounced and even in the winter the relative humidity of the air is rather low . . . it was found, that about $^1/_3$ or perhaps even more of the winter precipitation will have been evaporated before the heat of the summer starts". This conclusion is also reached from an analysis of Tables 8, 9, and 10, which give the mean monthly snow depths, snowfall, and temperatures for several arctic stations. At Thule the mean snow depth for February is 9.6 inches and it progressively decreases to 3.6 inches for May (Table 8). During the interval from February 15 to May 15 approximately 5.9 inches of snow falls (Table 9).

The mean temperature for this interval is approximately −3 °F (Table 10). Thus most of this 12-inch loss of snow, which in water equivalent equals approximately $^1/_4$ the mean annual precipitation, is due to evaporation. The mean temperature for Thule between February 15 and May 15 is approximately −3 °F, and between May 15 and August 15 approximately 37 °F (Table 10). Because the rate of evaporation, other parameters remaining equal, approximately doubles for each 18 °F rise in temperature, the potential evaporation at Thule between May 15 and August 15 may be more than 4 times that between February 15 and May 15. It seems likely, therefore, that the potential evaporation for Thule is greater than the precipitation, and that the climate of Inglefield Land and adjacent areas is arid (THORNTHWAITE, 1948).

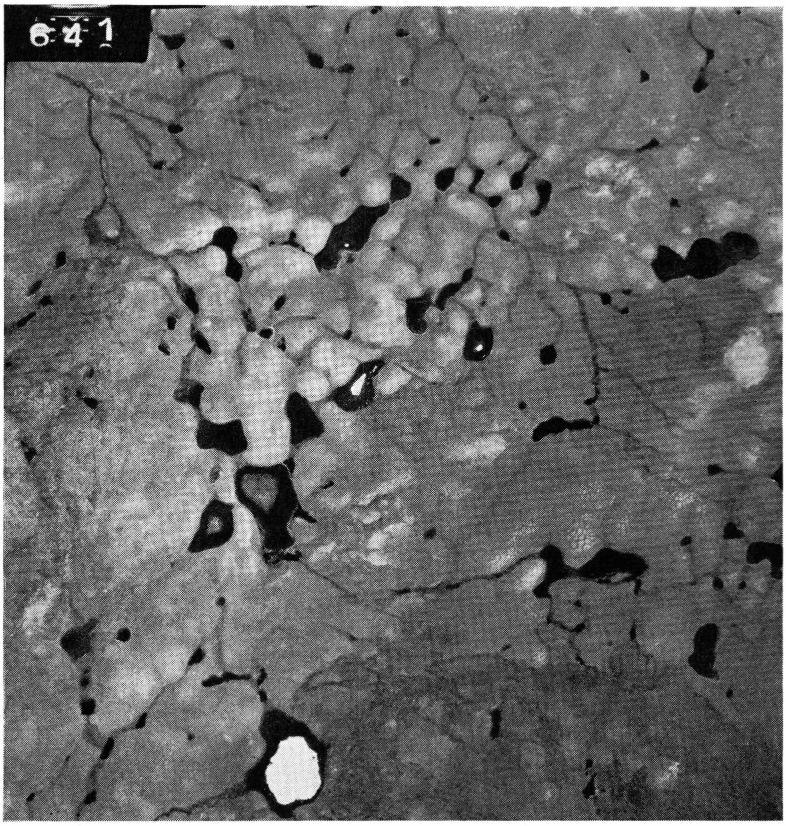

Fig. 43. An area covering 9 square miles containing over 50 lakes which are presumably moraine-dammed. Oblique photograph by Geodætisk Institut, Copenhagen. Appr. scale 1:50.000.

LAKES

There are hundreds of lakes in Inglefield Land. More than fifty were counted in an area of approximately 9 square miles (fig. 43). Their basins were formed by: (1) differential glacial deposition, (2) differential glacial erosion of bedrock, (3) ice dams, (4) the swales between beach ridges (fig. 9), (5) solifluction deposit dams, (6) a combination of these methods. The great majority, however, are moraine-dammed.

Some are perennially frozen; others have no ice covering during the warm season. Without exception the perennially frozen lakes have a moat of water between the ice and the shoreline during the warm season. If a lake is not as deep as the thickness of lake ice that can be melted in the warm season, it will not be perennially frozen.

Fig. 44. A perennially frozen ice-dammed lake 2–3 miles long near the source of the Tufts Elv. The Indlandsis terminates in the lake in an imposing cliff in places 50–100 feet high. Vertical photograph by Geodætisk Institut, Copenhagen. Appr. scale 1:90.000.

Ice-Dammed Lakes

Ice-dammed lakes, although not numerous, are present. The one near the headwaters of the Tufts Elv was briefly visited (fig. 18). It is between 2 and 3 miles long and about a mile wide at its widest point (fig. 44). The Greenland Indlandsis, which in most places in Inglefield Land ends in a ramp or snout, terminates in the lake in an imposing cliff 50 to 100 feet high (fig. 45). When first seen on July 21, 1953, the lake was frozen over except near the shoreline where a thaw margin had been formed. Small bergs derived from the ice cliff were frozen into the lake ice. On August 8 a stranded slab of lake ice between 4 and 5 feet thick was measured. This was close to the end of the melting season, so that it seems certain that not all the lake ice was melted and that the lake is therefore perennially frozen. The lake is fed by water which runs directly into it from the Indlandsis, by drainage from along its margin, and by two rivers. The southern river is fed mainly by meltwater from

Fig. 45. An ice-dammed lake adjacent to the Greenland Indlandsis near the source of the Tufts Elv.

the glacier; the water in the northern river comes from rainfall, melted snow, and melted ground ice. The lake empties into a stream that flows along the edge of the ice for between 2 and 3 miles and then leaves the margin of the ice and flows northward to Rensselaer Bugt. During the height of the melting season this outlet river is a rapidly-moving, sizable stream; it has eroded the margin of the Indlandsis and formed an ice cliff which in places is approximately 100 feet high (fig. 46). When seen on July 31, 1953, it was carrying large quantities of glacial ice, and the bumping in the rapids of this ice against the bottom of the river could be heard for considerable distances.

The lake owes its origin to the slope of the glacially modified preglacial topography toward the margin of the Indlandsis. The two arms or bays at the eastern end were formed because two valleys were drowned by the lake.

The level of the lake dropped more than 10 feet between July 25 and 27, 1953. This was shown by the presence of stranded slabs of lake ice and bergs, by the changed outline of the lake, and by levelling (fig. 47). The lake level was 15 feet lower on August 1 than on July 25, and during a 9-hour interval on August 1 the lake fell between 2 and 3 inches. On August 8 the level was approximately the same as on August 1.

The lake level dropped between 15 and 20 feet a short time before August 8, 1963. This was shown on August 8 by the presence immediately above the lake of: (1) stranded lacustrine ice, (2) miniature lacustrine clifflets and terraces, (3) wet sand in places covered with thin sheets of water, (4) perfectly preserved rill marks and ripple marks.

On July 31, 1953, a hurried traverse near the outlet of the lake may have shown why the level dropped more than 15 feet in about a week. On July 31 the lake was dammed by both the glacier and bedrock. About

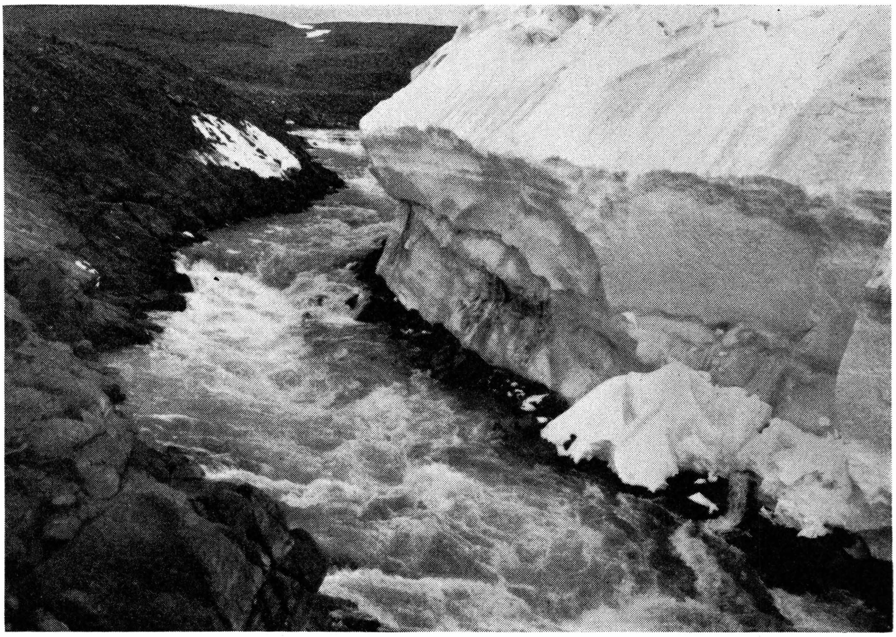

Fig. 46. The outlet stream of the ice-dammed lake near the headwaters of the Tufts Elv. At left are water-washed bedrock surfaces formed during the flood stage of the stream.

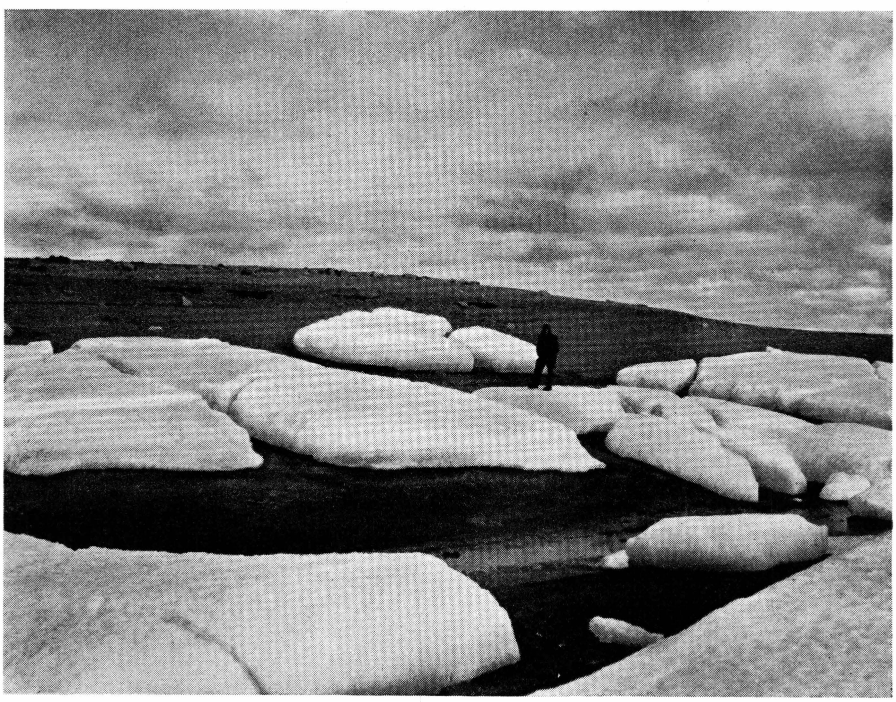

Fig. 47. A 15-foot lowering of the lake level between July 25 and August 1, 1963 stranded hundreds of slabs of lake ice and small icebergs.

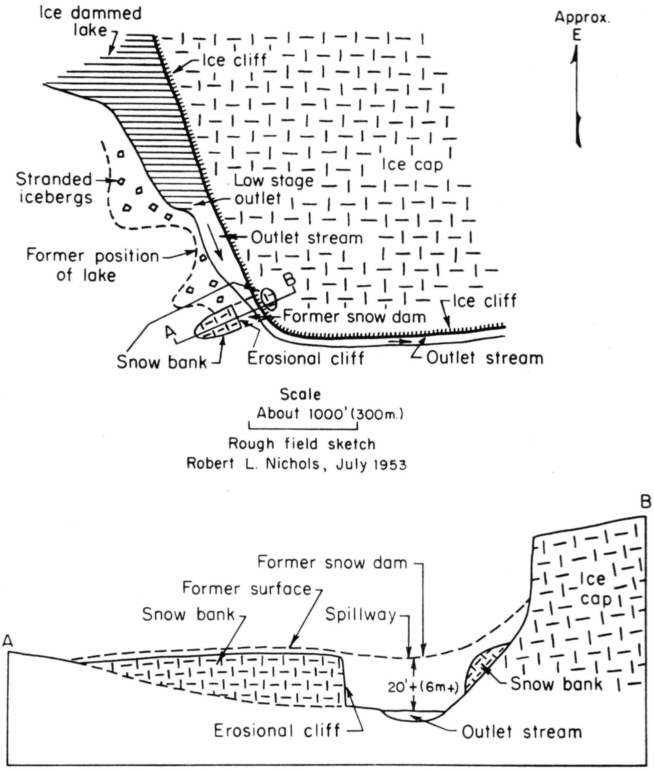

Fig. 48. Outlet of the ice-dammed lake.

1000 feet downstream from the outlet of the lake, where the ice cliff makes a right angle turn, the outlet river was running between the ice cliff and a large snowbank (fig. 48). The river had eroded the snowbank and formed a vertical cliff. A mass of snow was clinging to the ice cliff at a point somewhat higher than the highest part of the snowbank.

During the cold season, when there is no water running out of the lake and when it is dammed by both the glacier and bedrock, snow is blown off the Indlandsis and a snowbank is built which stretches from the Indlandsis across the channel of the outlet stream and for some distance beyond. This snowbank is more than 20 feet high at the channel of the outlet stream (fig. 48). After the melting season arrives, the level of the lake slowly rises because water runs into it, and now it is the snowbank that dams the lake. The lake level continues to rise until finally the snow dam fails. The level then drops rapidly until the ice cliff and bedrock a few hundred yards upstream again act as a dam, whereupon the lake level becomes more or less stationary.

By July 31, 1953, much of the snow dam had been removed but parts of it remained on both sides of the river. It is not known whether water running across the snow dam, water running beneath and through it, or a combination of these processes caused the dam to fail.

At one place on July 31 water-rafted pieces of glacial ice were found about 30 feet above the level of the river. The surge of water responsible for this rafting must have been associated with the destruction of the snow dam. On July 29, approximately 9 miles downstream, clifflets cut in fluviatile sand and stream-rounded boulders of sod containing fresh flowers had been found 3 feet above the level of the river, which was then in flood. These features probably also resulted from the failure of the snow dam which held in the lake.

Ice-Scratched Lacustrine Deposits

The drop in lake level stranded hundreds of slabs of lake ice and small bergs on the shore (fig. 47). The lowering also showed that the pebbly, sandy lake bottom was covered in many places with hundreds of scratches and grooves formed by the dragging and scraping of slabs of lake ice and bergs. The ice that formed any particular marking was present in some cases; in others it had either floated or melted away. The markings were oriented in all possible directions to the strandline. They varied greatly in size and shape, depending on whether a corner of an ice slab, a prong on the bottom of an iceberg, an icicle-like protuberance on lake ice, or a rock held in the bottom of an iceberg was responsible.

Three types of marks were seen. The commonest were scratches or grooves 10, 20, or more feet long, an inch or more deep, and a few inches to $1^1/_2$ feet wide (Hume and Schalk, 1964, plate 1). They were both straight and crooked, and were margined by levees an inch or more high. The largest of this type was 20 feet long and 10 inches deep. It was margined by a levee on one side only and was formed by a pronounced prong at the bottom of a small berg. Another type resembled the marks made by a metal rake in soft soil. The many more or less parallel grooves terminated in a levee at the end toward which the ice moved. These marks were apparently formed by a number of icicle-like prongs of the kind common on lake ice which has been considerably melted. The third type consisted of a crescentic ridge associated with a scar from which the material in the ridge was derived (Nichols, 1953 a, p. 175). These are small features, the ridges being at most only several inches high.

Similar features have been described in modern sedimets by Twenhofel (1932, p. 672–674); Clarke (1918, plates 15, 16, 17) has described markings in Devonian sandstones which he thought were formed in the same way.

Pyritic Sediments

On digging into the wet, fine-grained sand where the glacier-fed stream enters the southeast arm of the lake (figs. 17, 18), it was found that the upper $^1/_2$–1 inch was brown and that immediately below there was $^1/_4$–$^1/_2$ inch of black sand, below this $^1/_2$–1 inch of grey sand, and still lower brown sand.

The black color is due to the presence of iron sulfide, probably of the type described as hydrotroilite ($FeS\text{-}nH_2O$), an amorphous, hydrous monosulfide of iron. It is precipitated by sulfate reduction induced by *Desulfovibrio* in the sediments. It is not surprising that sulfate-reducing bacteria are found so far north under such rigorous conditions, as they have been found in Antarctica (BARGHOORN and NICHOLS, 1961). When the black sand was allowed to dry out under oxidizing conditions it turned gray, due to the oxidation of the iron sulfide.

RELATIVE IMPORTANCE OF THE GEOMORPHIC PROCESSES

On the basis of the writer's reconnaissance field work, it may be impossible to evaluate correctly the relative importance of the geomorphic features and exogenous processes in Inglefield Land. However, a tentative evaluation can be made (Table 11). It will be helpful to consider for each process, in a semi-quantitative way, the mass of material it has moved and the average distance it has moved this mass.

The Pre-Cambrian erosion surface which has been exhumed in parts of Inglefield Land (COWIE, 1961, p. 40–41) is also found, though not exhumed, in Northeast Greenland (HALLER and KULP, 1962, fig. 7). It is certain, therefore, that it once covered all parts of Inglefield Land. It cuts coarse-grained igneous and metamorphic rocks, and in its formation the detritus of several miles of rock was probably transported hundreds of miles by fluvial action. It is the most important geomorphic feature in Inglefield Land from the point of view of the magnitude of geologic work necessary to form it.

The youngest rocks now known in Inglefield Land (TROELSEN, 1950, geol. map) are Ordovician limestones and shales. They are found in eight or more areas none of which is over 5 miles long, but it is certain that they were formerly much more extensive. Whether younger Paleozoic, Mesozoic, and Cenozoic rocks were once present is not known for certain.

In any event, there was a period of fluvial erosion following the Ordovician and preceding the oldest glaciation. Ordovician, Cambrian, and Pre-Cambrian rocks were removed in places, the Pre-Cambrian erosion surface was exhumed, and the spectacular elements of the topography, including cliffs, plateaus, and structural terraces, were formed. Undoubtedly more than a thousand feet of rock was removed over large areas. More geomorphic work was accomplished and more important geomorphic features were formed during this period than any other, with the exception of the Pre-Cambrian erosion period.

As indicated above, the whole of Inglefield Land was glaciated during the Wisconsin. Little data is available, however, on the thickness of the glacially transported material. The Pre-Cambrian rocks close to Rensselaer Bugt crop out almost everywhere, and the till is therefore thin. On

Table 11. *Relative quantitative importance of geomorphic features and processes*

Process	*Features formed*	*Area covered distribution*	*Thickness removed or deposited*	*Average distance transported*	*Relative importance*
fluvial	Pre-Cambrian erosion surface	all of Inglefield Land	perhaps up to several miles	perhaps hundreds of miles	1
fluvial	post-Ordovician — pre-glaciation erosion — peneplain, cliffs, structural terraces, mesas	all of Inglefield Land	a maximum of more than 1000 feet	unknown distance	2
glaciation	till, moraines, erratics, glacial striations, roches moutonnées, and planation	all of Inglefield Land	on the average perhaps 10 feet	several miles	3
periglacial features	permafrost, solifluction deposits, sorted and nonsorted circles, frost cracks, block fields, earth hummocks, nivation cirques	common and widespread	less than 40 feet	less than a mile	4 or 5
glacial meltwater	marine deltas, valley trains, kame terraces, marginal channels	widespread	some deltas hundreds of feet thick	up to 10 miles	4 or 5
waves and shore currents	bay and ocean-side beaches	discontinuous along the coast; less than $^1/_4$ mile wide	less than 15 feet thick	a mile or less	6
wind	ventifacts, wind-eroded frost mounds, small dunes, and loess	few square miles at most	less than 10 feet	less than a mile	7

the other hand, a seismic survey less than a mile from the Indlandsis near the source of the Tufts Elv (figs. 17, 18) showed that the till there is approximately 90 feet thick.[1]) The coastal plateaus are underlain by Pre-Cambrian and early Paleozoic sedimentary rocks. Glacial erratics composed of Pre-Cambrian igneous rocks are everywhere present. Some of these erratics may have been transported 10 and more miles (TROELSEN, 1950, geol. map). Probably no great error will be made as to the importance of glaciation if we conclude that all of Inglefield Land is covered

[1]) Personal communication from Mr. JEAN JACQUES HOLTZSCHERER.

with several feet of till that has been transported, on the average, several miles (Flint, 1957, p. 77; 1947, p. 114–116).

The important features formed by glacial meltwater are marine deltas, valley trains, kame terraces, and marginal channels. Solifluction deposits, block fields, and nonsorted circles are the most important of the periglacial features. The periglacial features have a wider distribution and are very much more common than the meltwater features, but the material has not been moved so far and the deposits are not so thick. The meltwater features, as in the case of deltas, may be hundreds of feet thick, and the sand and gravel may have been transported as much as 10 miles. The writer finds it impossible to evaluate the relative importance of these two groups of features. Neither group is as important as glaciation; both are more important than the shoreline processes and the work of wind.

Waves and shore currents have not been important geomorphic processes in Inglefield Land in post-Wisconsin time. In general, only ocean and bayside beaches are present. They are found along only a fraction of the coast because, except in the drowned valleys, cliffs hundreds of feet high veneered with talus slopes are present almost everywhere. The beaches are less than a quarter of a mile wide and probably less than 15 feet thick on the average; and it seems likely that the material composing them has been transported by shore currents less than a mile in most cases. They are, however, more important than the eolian features.

The work of the wind has obviously been much less important than any of the other processes in shaping the landscape. The total area covered by eolian deposits could be only a few square miles at most and it might well be less than one square mile; and the average thickness of the deposits is probably less than 10 feet.

tvivlsom – man ser ikke resultatet idanar at vindpust magler.

CONCLUSIONS

The following are the important conclusions resulting from this study:

(1) Pre-Cambrian igneous and metamorphic rocks are widespread in Inglefield Land. So too are late Pre-Cambrian and early Paleozoic sedimentary rocks.

(2) A Pre-Cambrian peneplain was developed which cuts the igneous and metamorphic rocks and which was buried by the late Pre-Cambrian and early Paleozoic sedimentary rocks.

(3) A late Tertiary peneplain was developed which cuts the igneous and metamorphic rocks as well as the sedimentary.

(4) The coastal cliffs, bedrock valleys, and structural terraces resulted from the uplift and fluvial dissection of the Tertiary peneplain.

(5) The drowned valleys, promontories, and islands along the coast, and perhaps the straits and channels between Ellesmere Island and North Greenland, are due to the submergence of the uplifted dissected Tertiary peneplain by the weight of the Greenland Indlandsis.

(6) Glacial striations and grooves, roches moutonnées, glacially truncated surfaces, erratics, till, morainal cones and ridges, valley trains, kame terraces, varved clay, fluvial boulder pavements, marginal channels, bedrock basin lakes, moraine-dammed lakes, ice-dammed lakes, waterfalls, and other features resulting from glaciation are present.

(7) Inglefield Land was completely glaciated and the Carey Øer, 70 miles offshore from Thule, were covered by the Greenland Indlandsis, proving that during the Wisconsin the straits and channels between Ellesmere Island and Greenland were nearly or completely filled with glacial ice.

(8) C-14 measurements indicate that deglaciation occurred at Rensselaer Bugt more than 7800 ± 200 years ago and at Dallas Bugt more than 5900 ± 150 years ago. They also show that the duration of post glacial time progressively decreases from the southwest to the northeast.

(9) The presence of large aligned morainal ridges and cones proves the existence of a glacial stillstand of considerable duration during the deglaciation of Inglefield Land.

(10) Weathered glacial deposits, the absence of significant end moraines and marginal channels, and other features at the margin of the Greenland Indlandsis indicate that its present position is due to readvance rather than to retreat.

(11) The linears found in the pre-Cambrian igneous and metamorphic rocks are probably due to joints and faults, and they have been accentuated by fluvial erosion and glaciation.

(12) Elevated deltas and beaches show that the marine limit is approximately 285 feet near Force Bugt and 210 feet at Dallas Bugt 70 miles to the northeast. The marine terraces reported by Bøggild at 1050–1800 feet above sea level between 77–81 degrees do not exist, and the post-Wisconsin epeirogenic movement postulated by Daly to account for them did not take place.

(13) The non-paired terraces cut in the elevated dissected deltas are not due to marine erosion but to lateral planation and rejuvenation of streams consequent on isostatic rebound and on a decrease in stream loads.

(14) The absence of significant marine stillstand between the marine limit and the modern shoreline is indicated by the raised beaches.

(15) The size of the modern deltas and the presence of marine cliffs at the modern strandline indicate a significant stillstand of sea level at about its present position.

(16) Permafrost, solifluction, block fields, nivation cirques, earth hummocks, nonsorted circles and polygons, and ice-wedge and sand-wedge furrows are common.

(17) Although small dunes, loess, ventifacts, and wind-eroded earth hummocks are present, the work of wind is of minor importance.

(18) Surface efflorescences, travertine on the bottoms of fragments, solution-faceted and solution-shaped fragments, and deposition of travertine from ponds and streams all prove that the climate has been arid for thousands of years.

(19) The sediments of an ice-dammed lake contain sulphate-reducing bacteria.

REFERENCES CITED

Ambrose, J. W. (1964). Exhumed Paleoplains of the Precambrian Shield of North America. Amer. Jour. Sci., vol. 262, pp. 817–857.

Ashley, G. H. (1931). Our Youthful Scenery. Bull. Geol. Soc. Amer., vol. 42, pp. 537–545.

Barghoorn, E. S. and Nichols, R. L. (1961). Sulfate-Reducing Bacteria and Pyritic Sediments in Antarctica. Science, vol. 134, No. 3473, p. 190.

Bell, R. (1895). Honeycombed Limestones in Lake Huron. Bull. Geol. Soc. Amer., vol. 6, pp. 297–304.

Bendix-Almgreen, S. E., B. Fristrup and R. L. Nichols (1967). Notes on the Geology and Geomorphology of the Carey Øer, North-West Greenland. Meddr Grønland, Bd. 164, Nr. 8, pp. 1–19.

Bentham, R. (1936). Appendix 1: Geology, pp. 427–431 *in* Humphreys, N., E. Shackleton, and A. W. Moore, Oxford University Ellesmere Land Expedition. Geogr. Journ., vol. 87, No. 5, pp. 385–443 (London).

Bishop, B. C. (1957). Shear Moraines in the Thule Area, Northwest Greenland. U.S. Snow, Ice and Permafrost Research Establishment, Research Report No. 17, 46 pp. (Wilmette, Ill.).

Blackadar, R. G. (1956). Geological Reconnaissance of Admiralty Inlet, Baffin Island, Arctic Archipelago, Northwest Territories. Geol. Surv., Canada, Paper 55–6, 25 pp.

Böcher, T. W. (1949). Climate, Soil and Lakes in Continental West Greenland in Relation to Plant Life. Meddr Grønland, Bd. 147, Nr. 2, pp. 1–63.

Bøggild, O. B. (1928). The Geology of Greenland. Vol. 1, pp. 231–255 in Greenland, edited by M. Vahl *et al.*, pub. by the Commission for the Direction of the Geological and Geographical Investigations in Greenland, C. A. Reitzel, publisher, 575 pp. (Copenhagen).

Bretz, J. H. (1935). Physiographic Studies in East Greenland. Pp. 159–245 *in* Boyd, L. A., The Fiord Region of East Greenland, Am. Geog. Soc. Spec. Pub. 18, 369 pp.

Bryan, K. (1929). Solution-Facetted Limestone Pebbles. Amer. Journ. Sci., vol. 18, 5th series, pp. 193–208.

Clarke, J. M. (1918). Strand and Undertow Markings of Upper Devonian Time as Indications of the Prevailing Climate. New York State Museum Bull., No. 196, pp. 199–238.

Compton, P. A. (1964). Recent Changes of Sea-Level Along the Northeast Coast of Brodeur Peninsula, Baffin Island, N.W.T., Canada. Arctic, vol. 17, No. 4, pp. 279–285.

Cooke, H. C. (1929). Studies of the Physiography of the Canadian Shield, I. Mature Valleys of the Labrador Peninsula. Trans. Royal Soc. Canada, 3rd series, vol. 23, parts I and II, Section IV, pp. 91–120.

— (1930). Studies of the Physiography of the Canadian Shield, II. Glacial Depression and Post-Glacial Uplift. Trans. Royal Soc. Canada, 3rd series, vol. 24, part I, Section IV, pp. 51–87.

Cooke, H. C. (1931). Studies of the Physiography of the Canadian Shield. Trans. Royal Soc. Canada, 3rd series, vol. 25, Section IV, pp. 127–180.

Cowie, J. W. (1961). Contributions to the Geology of North Greenland. Meddr Grønland, Bd. 164, Nr. 3, pp. 1–47.

Daly, R. A. (1934). The Changing World of the Ice Age. Yale University Press, 271 pp. (New Haven, Conn.).

Davies, W. E., and D. B. Krinsley (1961). Evaluation of Arctic Ice-Free Land Sites Kronprins Christian Land and Peary Land, North Greenland, 1960. Geophysics Research Directorate, Air Force Cambridge Research Laboratories, Office of Aerospace Research, Air Force Surveys in Geophysics, No. 135, 51 pp.

— — (1962). The Recent Regimen of the Ice Cap Margin in North Greenland. Extract of publication No. 58 of the I.A.S.H., Commission of Snow and Ice, pp. 119–130.

— D. B. Krinsley and A. H. Nicol (1963). Geology of the North Star Bugt Area, Northwest Greenland. Meddr Grønland, Bd. 162, Nr. 12, pp. 1–68.

Flint, R. F. (1947). Glacial Geology and the Pleistocene Epoch. John Wiley and Sons, Inc., 589 pp.

— (1948). Glacial Geology and Geomorphology. Pp. 91–210 in Boyd, L. A., The Coast of Northeast Greenland, Am. Geog. Soc. Spec. Pub. 30, 339 pp.

— (1957). Glacial and Pleistocene Geology. John Wiley and Sons, Inc., 553 pp.

Fristrup, B. (1952). Danish Expedition to Peary Land, 1947–1950. Geogr. Review, vol. 42, No. 1, pp. 87–98.

— (1952–53). Wind Erosion within the Arctic Deserts. Særtryk af Geografisk Tidsskrift, Bd. 52, pp. 51–65.

Goldthwait, R. P. (1951). Development of End Moraines in East-Central Baffin Island. Journ. Geol., vol. 59, pp. 567–577.

Haller, J. and J. L. Kulp (1962). Absolute Age Determinations in East Greenland. Meddr Grønland, Bd. 171, Nr. 1, pp. 1–77.

Holtedahl, O. (1917). Summary of Geological Results. Report Second Norwegian Arctic Expedition in the "Fram", 1898–1902, vol. 4 (36), (Kristiania, Oslo).

Hudson, G. H. (1910). Joint Caves of Valcour Island. — Their Age and Their Origin. New York State Museum, Bull. 140, pp. 161–196.

Hume, J. D. and M. Schalk (1964). The Effects of Ice-Push on Arctic Beaches. Amer. Journ. Sci., vol. 262, pp. 267–273.

Jensen, J. A. D. (1889). Undersøgelse af Grønlands Vestkyst fra 64° til 67° N. B. 1884 og 1885. Meddr Grønland, Nr. 8, pp. 33–122.

Kelly, W. C. and J. H. Zumberge (1961). Weathering of a Quartz Diorite at Marble Point, McMurdo Sound, Antarctica. Journ. Geology, vol. 69, pp. 433–446.

Koch, L. (1926). Ice Cap and Sea Ice in North Greenland. Geographical Review, vol. 16, pp. 98–107.

— (1928a). The Physiography of North Greenland. Vol. 1. pp. 491–518, in Greenland, edited by M. Vahl *et al.*, C. A. Reitzel, publisher, 575 pp. (Copenhagen).

— (1928b). Contributions to the Glaciology of North Greenland. Meddr Grønland, Bd. 65, Nr. 15, pp. 181–464.

— (1933). The Geology of Inglefield Land. Meddr Grønland, Bd. 73, Nr. 2, pp. 1–38.

Krinsley, D. B. (1961). Late Pleistocene Glaciation in Northeast Greenland. Pp. 747–751, vol. II in Geology of the Arctic, edited by Gilbert O. Raasch, University of Toronto Press, 1196 pp. (Toronto, Canada).

— (1963). Glacial Geology, *in* W. E. Davies, D. B. Krinsley and A. H. Nicol, Geology of the North Star Bugt Area, Northwest Greenland, Meddr Grønland, Bd. 162, Nr. 12, pp. 48–68.

LACHENBRUCH, A. H. (1960). Thermal Contraction Cracks and Ice Wedges in Permafrost, *in* Geological Survey Research 1960, U.S. Geological Survey Prof. Paper 400-A, pp. B404–B406.

LAHEE, F. H. (1931). Field Geology. McGraw-Hill Book Co., Inc., 789 pp. (New York and London).

LAURSEN, D. (1944). Contributions to the Quaternary Geology of Northern West Greenland Especially the Raised Marine Deposits. Meddr Grønland, Bd. 135, Nr. 8, 125 pp.

— (1950). The Stratigraphy of the Marine Quaternary Deposits in West Greenland. Meddr Grønland, Bd. 151, Nr. 1, pp. 1–142.

— (1954). Emerged Pleistocene Marine Deposits of Peary Land (North Greenland). Meddr Grønland, Bd. 127, Nr. 5, 26 pp.

LEFFINGWELL, E. DE K. (1919). The Canning River Region, Northern Alaska. U.S. Geol. Surv. Prof. Paper 109, 251 pp.

MALAURIE, J. (1955). The French Geographical Expedition to Thule, 1950–1951: a Preliminary Report. Arctic, vol. 8, No. 4, pp. 203–214.

MILLAR, C. E., L. M. TURK and H. D. FOTH (1958). Fundamentals of Soil Science, John Wiley and Sons, Inc., 526 pp. (New York).

NICHOLS, R. L. (1953a). Marine and Lacustrine Ice-Pushed Ridges. Journ. Glaciology, vol. 2, No. 13, pp. 172–175.

— (1953b). Geomorphologic Observations at Thule, Greenland and Resolute Bay, Cornwallis Island, N.W.T. Amer. Journ. Sci., vol. 251, pp. 268–275.

— (1961). Characteristics of Beaches Formed in Polar Climates. Amer. Journ. Sci., vol. 259, pp. 694–708.

— (1963a). Miniature Nivation Cirques Near Marble Point, McMurdo Sound, Antarctica. Journ. Glaciology, vol. 4, No. 34, pp. 477–479.

— (1963b). Geologic Features Demonstrating Aridity of McMurdo Sound Area, Antarctica. Amer. Journ. Sci., vol. 261, pp. 20–31.

— (1966). Geomorphology of Antarctica. Pp. 1–46 *in* Antarctic Soils and Soil Forming Processes, Antarctic Research Series, vol. 8, American Geophysical Union, 1–177 pp.

NOE-NYGAARD, A. (1932). Remarks on *Mytilus edulis* L. in Raised Beaches in East Greenland. Meddr Grønland, Bd. 95, Nr. 2, pp. 1–24.

PESSL, F., JR. (1962). Glacial Geology and Geomorphology of the Sortehjørne Area, East Greenland. Arctic, vol. 15, pp. 73–76.

PÉWÉ, T. L. (1959). Sand-Wedge Polygons (Tesselations) in the McMurdo Sound Region, Antarctica — A Progress Report. Amer. Journ. Sci., vol. 257, pp. 545–552.

SCOTT, H. W. (1947). Solution Sculpturing in Limestone Pebbles. Bull. Geol. Soc. Amer., vol. 58, No. 2, pp. 141–152.

SHARP, R. P. (1942). Soil Structures in the St. Elias Range, Yukon Territory. Journ. Geomorphology, vol. 5, No. 4, pp. 274–301.

SMITH, J. F., JR. and C. C. ALBRITTON, JR. (1941). Solution Effects on Limestone as a Function of Slope. Bull. Geol. Soc. Amer., vol. 52, No. 1, pp. 61–78.

SPENCER, E. W. (1959). Geologic Evolution of the Beartooth Mountains, Montana and Wyoming. Bull. Geol. Soc. Amer., vol. 70, pp. 467–508.

SUGDEN, D. E. and B. S. JOHN (1965). The Raised Marine Features of Kjove Land, East Greenland. Geographical Journal, vol. 131, pt. 2, pp. 235–247 (London).

TABER, S. (1943). Perennially Frozen Ground in Alaska: Its Origin and History. Bull. Geol. Soc. Amer., vol. 54, No. 10, pp. 1433–1548.

TANNER, V. (1938). Die Oberflächen Gestaltung Finnlands. Bidrag Till Kännedom af Finlands Natur och Folk, H. 86, pp. 1–762.

Tedrow, J. C. F. (1966). Polar Desert Soils. Proceedings Soil Science Society of America, vol. 30, No. 3, pp. 381–387.

Thiel, E. C. (1962). The Amount of Ice on Planet Earth. Geophysical Monograph Series No. 7, Amer. Geophysical Union, pp. 172–175.

Thornthwaite, C. W. (1948). An Approach Toward a Rational Classification of Climate. Geogr. Rev., vol. 38, pp. 55–94.

Troelsen, J. C. (1949). Contributions to the Geology of the Area round Jørgen Brønlunds Fjord, Peary Land, North Greenland. Meddr Grønland, Bd. 149, Nr. 2, pp. 1–29.

— (1950). Contributions to the Geology of Northwest Greenland, Ellesmere Island and Axel Heiberg Island. Meddr Grønland, Bd. 149, Nr. 7, pp. 1–86.

— (1952). Notes on the Pleistocene Geology of Peary Land. Meddelelser fra Dansk Geologisk Forening, Bd. 12, Hefte 2, pp. 211–220.

Twenhofel, W. H. (1932). Treatise on Sedimentation. The Williams & Wilkins Company, 926 pp. (Baltimore).

U. S. Navy Dept. Hydrographic Office (1963). Borden Island to De Long Fiord. Hydrog. Office pub. no. 6606.

Washburn, A. L. (1947). Reconnaissance Geology of Portions of Victoria Island and Adjacent Regions, Arctic Canada. Geol. Soc. America Mem. 22, 142 pp.

— (1956). Classification of Patterned Ground and Review of Suggested Origins. Bull. Geol. Soc. Amer., vol. 67, No. 7, pp. 823–865.

— (1965). Geomorphic and Vegetational Studies in the Mesters Vig District, Northeast Greenland: General Introduction. Meddr Grønland, Bd. 166, Nr. 1, pp. 1–60.

Washburn, A. L. and M. Stuiver (1962). Radiocarbon-Dated Postglacial Delevelling in Northeast Greenland and its Implications. Arctic, vol. 15, pp. 66–73.

Wenk, E. (1961). On the Crystalline Basement and the Basal Part of the Pre-Cambrian Eleonore Bay Group in the Southwestern Part of Scoresby Sund. Meddr Grønland, Bd. 168, Nr. 1, 54 pp.

White, S. E. (1956). Glaciological Studies of Two Outlet Glaciers, Northwest Greenland, 1953. Meddr. Grønland, Bd. 137, Nr. 8, pp. 1–31.

Wilson, J. T. (1941). Structural Features in the Northwest Territories. Amer. Journ. Sci., vol. 239, No. 7, pp. 493–502.

— (1948). Some Aspects of Geophysics in Canada with Special Reference to Structural Research in the Canadian Shield. Amer. Geophysical Union, Trans., vol. 29, No. 5, pp. 691–726.

— (1963). Continental Drift. Scientific American, vol. 208, No. 4, pp. 86–100.

Wordie, J. M. (1938). An Expedition to North West Greenland and the Canadian Arctic in 1937. Geogr. Jour., vol. 92, pp. 385–421.

World Aeronautical Chart (1952). Smith Sound, Canada—Greenland, No. 20. USAF Aeronautical Chart and Information Center, St. Louis, Missouri.

Wright, J. W. (1939). Contributions to the Glaciology of North-West Greenland. Meddr Grønland Bd. 125, Nr. 3, 43 pp.

Færdig fra trykkeriet den 21. juli 1969.