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Institute of Polar Studies
Report No. 40

GLACIAL GEOLOGY OF THE BURROUGHS GLACIER AREA, SOUTHEASTERN ALASKA

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David M. Mickelson
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ABSTRACT

Burroughs Glacier is a sprawling, stagnating remnant of a much larger Neoglacial ice mass in Glacier Bay National Monument. In 1892, when the earliest photographs were taken of it, an ice plateau 10 km by 25 km was present. Since that time the ice surface has downwasted as much as 750 m and its calving margin has retreated 27 km. The ice mass separated into the Burroughs Glacier with its margin on land and Plateau Glacier with its terminus at sea level. Between 1960 and 1970 ice surface lowering averaged 9.5 m/yr at 205 m elevation (in 1960) to 4.6 m/yr at 440 m elevation (in 1960). Retreat of the land-based ice front was up to 140 m/yr and of the calving ice margin up to 350 m/yr during this time.

During deglaciation, the ice-flow direction changed because of the emergence of hills. Vertical sections of till show changes in lithology and fabric direction with depth. Because the rate of change of ice-flow direction is known the rate of till deposition has been calculated. Rates at seven localities on stoss and lee sides of nunataks range from 0.4 cm/yr to 2.8 cm/yr. Most till deposition took place after 1890 except on the stoss side of nunataks where thick till has been deposited. Hilltop striations were cut until just before the hills emerged.

Moraine ridges up to 2 m high have been squeezed into crevasses from below near the calving Plateau Glacier margin. Within the ridges, till fabric diagrams have modes with azimuths that are not necessarily perpendicular to the ridges but are parallel to the horizontal component of the last ice-flow direction.

With the emergence of nunataks near the southeastern Burroughs terminus, small, stagnant bodies of ice were left in the lee of nunataks. Here sand and gravel eskers up to 100 m long and 4 m high and with sharp crests head at the break in slope between the valley wall and valley bottom. In one case they were deposited between 1959 and 1960 by streams flowing at the base of the ice in the same direction as the ice surface gradient. Another esker, composed of laminated silt and sand at its downstream end, was deposited between 1965 and 1967 in a closed tube under hydrostatic pressure.

In some cases water flowing around nunataks did not flow under the ice but flowed along the ice margin depositing kame terraces, or into the ice leaving a hummocky blanket of sand and gravel on the till.

Three ice-dammed lakes were present at the southeast terminus of Burroughs Glacier at times between 1941 and 1960. Sand and silt up to 2 m thick remain, but distinct shorelines were not formed. Meltwater flowing directly from the glacier has deposited outwash both on and off ice. The major stream began depositing outwash between 1941 and

1948. By 1970 outwash at the 1948 margin position had been incised about 23 m. During this interval the ice withdrew about 1300 m up the valley. The subaerial part of the delta at the mouth of the stream increased in area by 3.3×10^5 m^2 during the period. Outwash deposited on ice between about 1956 and 1964 had collapsed by 1970.

Water flowing from the ice has cut marginal channels in till to depths of 20 m. Sets of three and four large channels were occupied 10 to 15 years. Smaller marginal channels, 1-2 m deep, were cut along the stoss sides of several hills. At least 8 successively lower channels up to 100 m in length were formed in one year. The gradients of these channels reflect the ice margin gradient.

ACKNOWLEDGMENTS

This project was supported by National Science Foundation Grant GA-12300 which was administered by The Ohio State University Research Foundation (Project 2809) and the Institute of Polar Studies. I am indebted to Dr. Richard P. Goldthwait, Professor of Geology at The Ohio State University, who was Principal Investigator of the grant and who initiated the project. I offer my sincerest thanks for his guidance both in the field and during the preparation of this dissertation.

Special thanks are also due to the U. S. National Park Service for logistical support. The help of Mr. Robert Howe, Superintendent of Glacier Bay National Monument, Mr. Charles Janda, Chief Ranger, Captain James Saunders and the crew of the M. V. Nunatak, and the remainder of the staff at Glacier Bay National Monument is greatly appreciated. I was capably assisted in the field in 1969 by Mr. Robert Votaw and in 1970 by Messrs. Grahame Larson, Robert Ward and John Rose.

I would also like to thank Mr. W. O. Field of the American Geographical Society whose numerous photographs and observations greatly enhanced this study. Aerial photographs taken by Mr. Austin Post of the U. S. Geological Survey were also extremely useful in deciphering the history of deglaciation. Photographs taken in 1959 and 1960 by Dr. L. D. Taylor and Dr. R. P. Goldthwait were also very useful.

Various members of the Department of Geology and the Institute of Polar Studies at The Ohio State University were very helpful. I would like to thank Dr. Charles E. Corbató for his assistance in constructing till fabric diagrams and for carefully reading this manuscript. I also extend my thanks to Dr. Wayne A. Pettyjohn for his advice in the field and helpful comments when reviewing this manuscript. I extend my gratitude to Dr. Colin Bull, Dr. Kaye Everett, Dr. Garry McKenzie, Dr. Valter Schytt, Mr. John Splettstoesser, Mr. Marcus Hoyer and Mr. Ian Whillans for many helpful suggestions. I am also indebted to Dr. L. P. Wilding of the Department of Agronomy for his suggestions and his help in the identification of clay minerals.

The 1970 base map was constructed from aerial photographs taken by the U. S. Coast and Geodetic Survey. I extend my thanks to Dr. S. K. Ghosh and Mr. H. H. Brecher for the construction of the map.

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CHAPTER I

INTRODUCTION

Purpose and Nature of Investigation

The importance of downwasting and stagnation of ice in the retreat of Pleistocene ice sheets in hilly areas has been discussed extensively in the literature. This is especially true in the case of ice sheets in New England where Flint (1929, 1932, 1934, 1942), Goldthwait (1938), Goldthwait (1968) and others have suggested that stagnation and downwasting were important facets of deglaciation. Others, such as Lougee (1940, 1956), have argued that deglaciation was accomplished by backwasting of a well-defined ice margin.

The Burroughs Glacier, in Glacier Bay, southeastern Alaska, is a stagnating remnant of a much larger, active Neoglacial ice mass that exists in an area with topography similar to parts of New England. Kame terraces, outwash deposits, eskers, ice-marginal channels, and minor moraines have recently formed and are still forming around the perimeter of the Burroughs Glacier. The processes and duration of formation of these geomorphic features are documented in this study. These data should be useful in reconstructing the history of late Pleistocene deglaciation in many areas with similar topography.

The Burroughs Glacier was chosen for the study not only because of its topographic situation but also because maps and photographs dating from 1890 to 1970 are available to aid in the interpretation of the history of deglaciation. The work of Taylor (1962a, b) on ice movement, ablation, and ice structures on the Burroughs Glacier in 1959 and 1960 has also been valuable in the interpretation of numerous features around the present ice margin.

Location of Field Work

The Burroughs Glacier is located in the north central part of Glacier Bay National Monument between 58° 71' and 59° 2' N Latitude and 136° 10' and 136° 25' W Longitude in southeastern Alaska (fig. 1). The glacier, shown as part of the Cushing Plateau by Reid (1896) and connected to the Cushing and Plateau Glaciers on the U. S. Geological Survey Mt. Fairweather (D-1), Mt. Fairweather (D-2), and Skagway (A-4) Quadrangles is presently an isolated ice mass below the firn line.

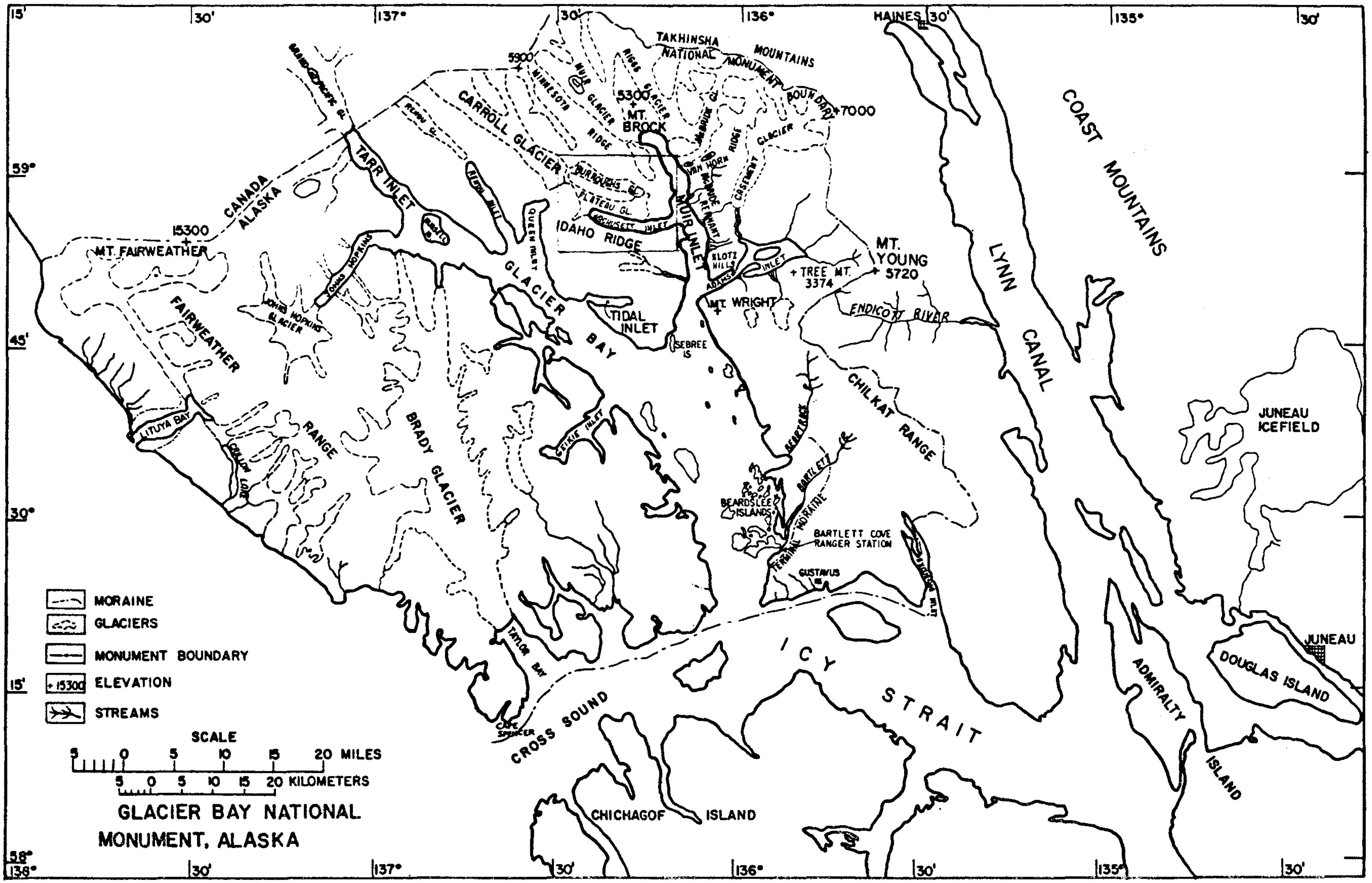


Figure 1. Index map of Glacier Bay, southeastern Alaska.

Detailed investigations were made at the southeast terminus of the Burroughs Glacier and along a nearly isolated ice mass directly north of and parallel to Wachusett Inlet (fig. 7). This ice mass is best called the Plateau Remnant to distinguish it from the active, calving part of the Plateau Glacier. Reconnaissance mapping was done along both sides of the Bruce Hills, the northwest terminus of the Burroughs Glacier, and the southwest side of Wachusett Inlet (fig. 7). No topographic map of the entire region more recent than 1948 is available, but 1970 ice margins are superimposed on the 1948 map in Figure 7.

Bedrock Geology

Rocks in the vicinity of the Burroughs Glacier are primarily diorite, grandodiorite, light and dark dike rocks, and metamorphosed limey clastic rocks of Paleozoic and Mesozoic ages (fig. 2). The Curtis Hills, most of the Bruce Hills, the northern half of Minnesota Ridge, and Nunatak A are Cretaceous stocks consisting mainly of medium-grained diorite that intrudes Paleozoic rocks (Rossman, 1963; MacKevett et al., 1971)(fig. 2, 10). No granite or syenite was found by the writer on the Bruce Hills or at the southeast end of the Burroughs Glacier, although some were identified in pebble lithology counts.

Metasedimentary rocks underlie most of the southeast end of the Burroughs Glacier. The most abundant are beds of alternating lavender and green metasiltstones that range from 0.1 to 1 m in thickness. These are interbedded with a few light and dark metamorphosed shale units as much as 1 m thick. A dark-green metamorphosed limestone was noted in one outcrop on Nunatak E (fig. 10) and evidently underlies the siltstone and shale sequence. A similar sequence, presumed to be of Devonian age, is described by Twenhofel (1946) across Muir Inlet at Nunatak Knob. Many light-colored and mafic dikes as much as 5 m thick occur in the area, cut both the plutonic and metasedimentary rocks, and are presumed to be Cretaceous in age (Rossman, 1963). These were not mapped in detail so they are not shown in Figure 2.

Some correlation between bedrock distribution and lithologies found in pebble counts, heavy minerals, and clay minerals can be made. This is useful in determining changes in ice-flow direction and is discussed in detail in Chapter III.

Climate and Ablation

Climate

The climate of Glacier Bay has been summarized by Loewe (1966) and more recently by McKenzie (1968, 1970). Although no permanent metorological stations are maintained in Glacier Bay, observations have been

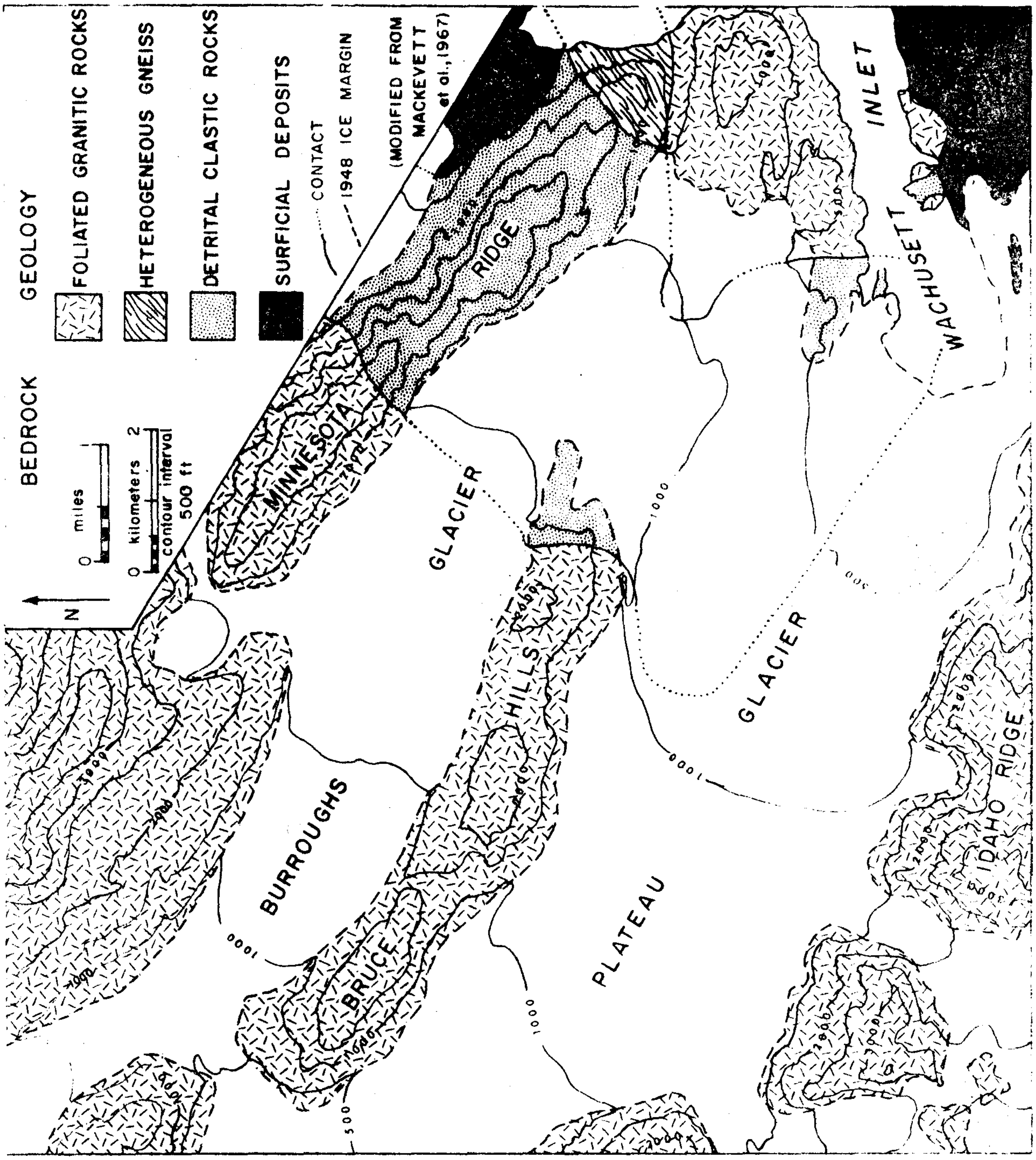


Figure 2. Bedrock geology map of Burroughs Glacier area.

made by field parties at the Burroughs Glacier in 1960 (Taylor, 1962a, b) and during the summers of 1969 and 1970 (this paper). Observations have been made on the east side of Muir Inlet by Haselton (1966, 1967) in 1963 and 1964, by Price (1964) in 1962, by Peterson (1969, 1970) in 1965, 1966 and 1967 and McKenzie (1968, 1970) in 1966 and 1967. Temperature and precipitation data for these stations are given in Table 1.

Abundant precipitation, cloudiness, high humidity, small temperature variations, and relatively strong winds characterize the region. Loewe (1966) concluded that summer observations at different points around Muir Inlet show no notable differences except for the station right next to Casement Glacier at an elevation of 300 m. The data given in Table 1 are probably representative of conditions at the Burroughs Glacier at those times. It is probable, however, that the Wachusett Inlet station in 1960 had slightly lower temperatures than would be expected at sea level because of its proximity to the Burroughs Glacier.

The years 1969 and 1970 were characterized by slightly lower temperatures than were recorded in Adams Inlet in 1966 and 1967 and about the same as those recorded at the Nunatak Station in 1963 and 1964. Total precipitation during July and August was higher than that recorded by earlier field parties in this section of Glacier Bay.

Ablation

Ablation was measured during the summers of 1969 and 1970 at the southeastern end of Burroughs Glacier on a weekly basis at elevations ranging from 120 m to 276 m. Total ablation between August 25, 1969 and August 23, 1970 ranged from 503 cm at an elevation of 276 m (early August 1970) to 701 cm at an elevation of 189 m. Ablation at lower elevations was higher but it could not be measured. A comparison with the 1960 ablation data given by Taylor (1962a) for approximately the same stake positions is given in Figure 3. Ablation and climate data, and a discussion of their relationships will be given by Larson (in preparation).

Glacial History Before the Neoglacial Maximum

Glacial stratigraphy in Glacier Bay National Monument has been described by Cooper (1923, 1937), Lawrence (1958), Goldthwait (1963, 1966a,b), Haselton (1966, 1967) and McKenzie (1968, 1970). The general stratigraphy of deposits in Wachusett Inlet is presented in Figure 4 and a brief description of each unit and its role in the glacial history of the area is given below. The distribution of surficial deposits is shown in Figure 5.

Table 1. Summary of summer meteorological data in Glacier Bay.

Month/ Year	Location	Elevation (m)	Maximum Temperature (°C)	Minimum Temperature (°C)	Monthly Mean Temperature (°C)	Total Precipitation (cm)	Total Number of Days	Mean Maximum Temperature (°C)	Mean Minimum Temperature (°C)
7/1960	Wachusett Inlet	30	-	-	10.0	-	-	-	-
8/1960	"	30	-	-	9.7	19.6	31	-	-
6/1963	Nunatak Cove	0	20	4	10	3.2	8	15	6
7/1963	"	0	21	3	10	7.5	31	15*	5*
8/1963	"	0	22	2	10	6.0	31	15*	5*
6/1964	"	0	17	2	9	3.2	11	14	5
7/1964	"	0	22	4	10	20.8	31	*	*
8/1964	"	0	20	1	9	15.3	31	*	*
9/1964	"	0	17	0	9	0.0	12	16	3
7/1965	Casement Glacier	300	-	-	-	-	-	10.0	4.4
8/1965	"	300	-	-	-	-	-	11.5	4.1
6/1966	Adams Inlet	10	22.7	3.3	11.4	0.4	10	17.1	5.8
7/1966	"	10	23.6	4.5	12.6	5.7	31	17.7	7.5
7/1966	Casement Glacier	300	-	-	-	29.8*	49*	12.1	2.5
8/1966	"	300	-	-	-	*	*	8.3	1.8
8/1966	Adams Inlet	10	22.4	3.3	9.7	23.2	31	13.8	5.6
6/1967	Adams Inlet	10	21.1	5.3	11.8	2.6	8	15.8	7.8
7/1967	"	10	21.8	5.3	11.8	8.9	31	16.3	7.2
8/1967	"	10	20.0	4.4	11.4	20.9	25	15.5	7.3
6/1969	Wachusett Inlet	30	25	5	13	0.0	6	19	7
7/1969	"	30	17	4	10	19.8	31	14	6
8/1969	"	30	18	2	9	18.4	31	14	4
9/1969	"	30	15	3	9	1.1	5	13	4
6/1970	"	30	20	4	10	9.6	25	14	6
7/1970	"	30	19	3	10	22.3	31	14	6
8/1970	"	30	19	4	9	19.5	23	12	6

*Average values for 1963 and 1964 and June and July, 1966

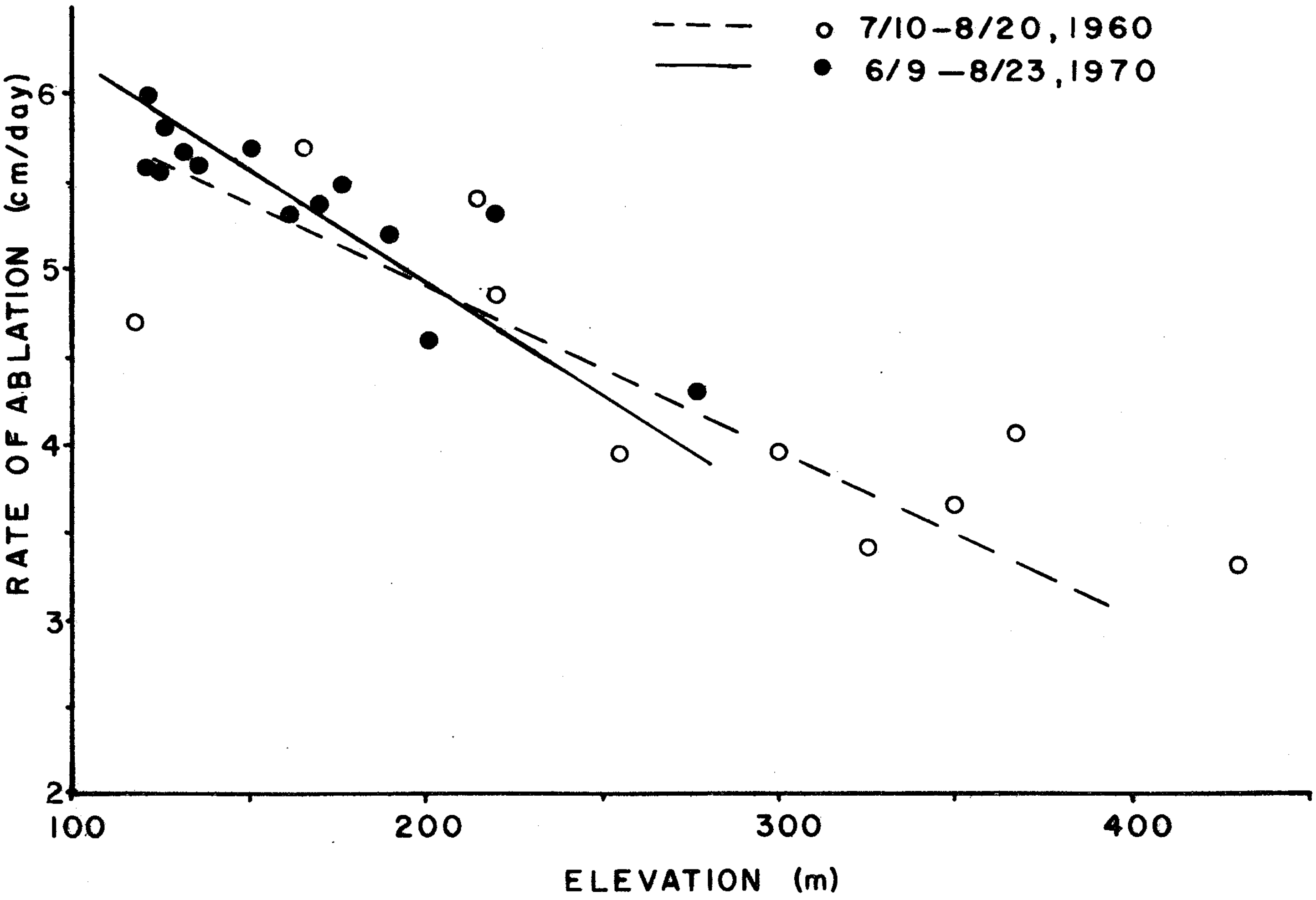


Figure 3. Relationship between elevation and ablation in 1960 and 1970.

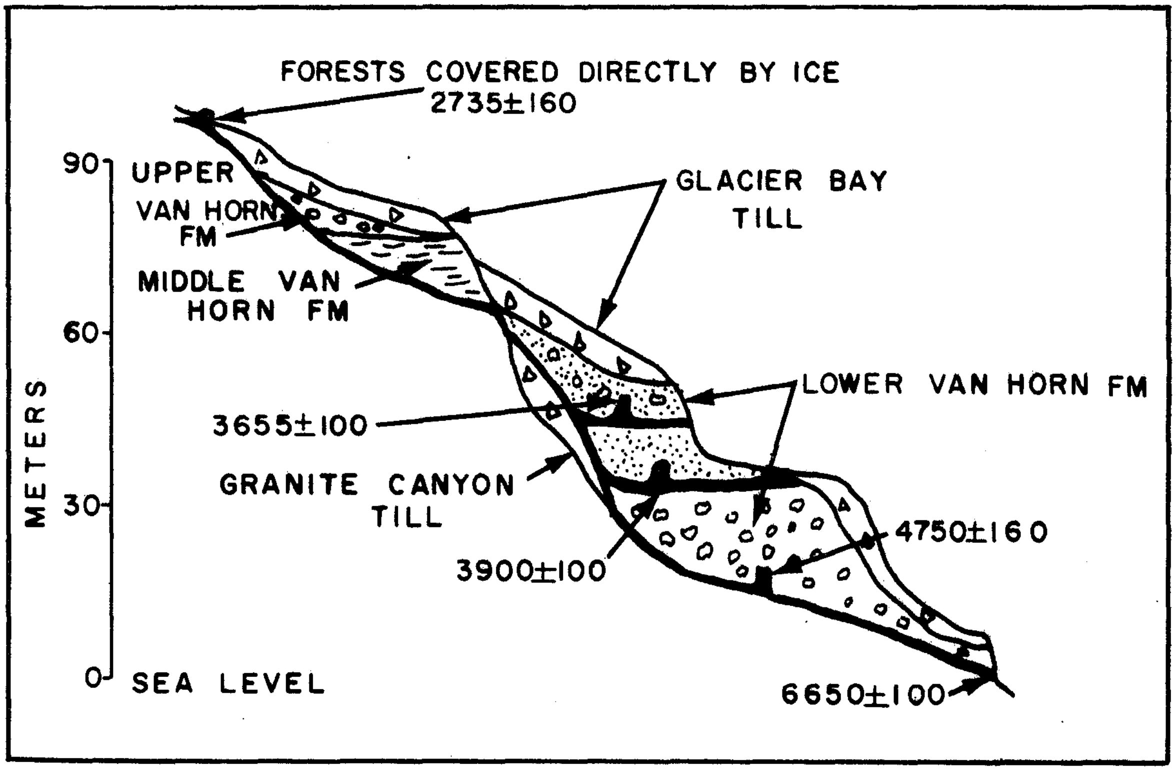


Figure 4. Glacial stratigraphy for the Burroughs Glacier area. (Modified from Goldthwait, 1963)

Granite Canyon Till

During the maximum extent of Wisconsin glaciation most of southeastern Alaska was covered by the Cordilleran Ice Sheet. The Granite Canyon Till (McKenzie, 1970) was deposited by this ice throughout the area. As used herein the Granite Canyon Till and the Muir Formation (Haselton, 1966) are considered to be identical and the term Granite Canyon Till is used because the stratigraphic position of the Muir Formation as defined by Haselton has been questioned (McKenzie, 1970). A minimum radiocarbon age for the formation on the east side of Muir Inlet was determined from an evergreen cone from the Forest Creek Formation (Haselton, 1966). The Forest Creek Formation, which lies directly above the Granite Canyon Till has been radiocarbon dated at 11,170±225 yr B.P. (I-2396). Peat above the Forest Creek Formation is dated at 10,400±260 yr B.P. (I-1615) and 10,940±155 yr B.P. (I-2395). Since deposition of the Forest Creek Formation probably took place in the flooding sea immediately after deglaciation, Wisconsin ice retreated shortly before 11,000 yr B.P. in the Muir Inlet area. No evidence of the Forest Creek Formation has been found in Wachusett Inlet although drowning probably occurred since it amounted to 60 m in an area only 15 km away.

The Granite Canyon Till in the Wachusett Inlet area is classified as a compact, gray loam where unweathered. In two of three exposures examined, a thin paleosol is present on the till and in one of these a thick layer (0.5 m) of forest duff and overturned rooted trees is present. Although the logs were not identified by species they probably represent a forest of hemlock and spruce with some cedar as was described by Goldthwait (1966a).

Granite Canyon Till in two fresh exposures in the area averages 49 percent sand, 41 percent silt, and 10 percent clay in the <2 mm fraction. This compares with 61 percent sand, 32 percent silt, and 7 percent clay found in the Muir Inlet area by Haselton (1967) and 41 percent sand, 39 percent silt and 20 percent clay found by McKenzie (1968) in the Adams Inlet area. The Granite Canyon Till cannot be distinguished from the more recent till of the Glacier Bay Formation on the basis of mechanical analysis in the Burroughs Glacier area (fig. 16). Clay minerals from two unoxidized samples of Granite Canyon Till show no significant differences from the overlying Glacier Bay till in the Burroughs Glacier area (table 2) or surrounding Adams Inlet (McKenzie, 1970). As noted by Haselton (1966) pebble counts of the Wisconsin age till surrounding Muir Inlet have more plutonic rocks than the overlying till of the Glacier Bay Formation. Fabrics in till surrounding Muir Inlet suggest that the Wisconsin age ice moved from a more westerly direction and this may account for differences in lithologies.

Table 2. Mean and extreme clay mineral percentages of Glacier Bay and Granite Canyon Tills.

	Glacier Bay Till N=17			Granite Canyon Till N=2		
	Mean* (%)	High (%)	Low (%)	Mean* (%)	High (%)	Low (%)
Illite	11	25	T	8	15	T
Vermiculite	11	40	T	20	20	20
Quartz	21	45	5	30	45	15
Chlorite	51	80	35	38	40	35
Expandable Clays	6	20	T	3	5	T

*Trace is considered 1% when calculating mean

Van Horn Formation

The Van Horn Formation (Haselton, 1966), deposited during Hypsithermal and early Neoglacial time, consists of washed valley-fill deposits nearly 100 m thick that are divided into three members. The lower member is composed of sand and gravel, usually rusty brown to yellowish in color in Muir and adjoining inlets and occurs from some depth below sea level to 50 m above sea level. The middle member is composed of lacustrine silt and is overlain by the upper member which is composed of sand and gravel.

The deposits were noted by early visitors to Glacier Bay and have been interpreted both as kame terraces (Cooper, 1937) and as outwash deposits (Goldthwait, 1963, 1966a, b; Haselton, 1966, 1967). Radiocarbon dates, internal structure, and composition suggest that the latter interpretation is correct and that the present inlets were once filled from side to side with outwash.

Deposition of the lower member of the Van Horn Formation began in Wachusett Inlet sometime before the gravels buried a stump at present sea level 6650±100 yr B.P. (I-59-13). Deposition continued at a rate estimated at 1.4 cm/yr (Goldthwait, 1963) until late in Hypsithermal time. Deposition in the Wachusett Inlet area was interrupted at least twice as evidenced by forest levels dated at 3900±100 yr B.P. (I-59-IC) and 3655±100 yr B.P. (I-59-ID) (fig. 4).

The middle member of the Van Horn Formation is composed of lacustrine silt. Evidently it was deposited in lakes dammed by advancing Neoglacial ice. A sample collected directly below these silts at Goose Cove, east of Muir Inlet is radiocarbon dated at 3290±55 yr B.P. (Y-303) (Goldthwait, 1963). In the Wachusett Inlet area, the silt deposits are 10-15 m thick and their highest exposures are at an elevation of about 75 m. Clay mineral analysis on unoxidized samples of this silt revealed no effect of weathering and is not significantly different from the Granite Canyon Till or the till of the Glacier Bay Formation (table 2). The upper member of the Van Horn Formation (Haselton, 1966) is outwash gravel, probably indicating an advance of glaciers from the northwest. The unit is poorly bedded, generally less oxidized than the lower member of the Van Horn Formation and contains some sand beds throughout its 75 m maximum thickness. Haselton (1966) reports that lithologies of the lower and upper members of the Van Horn Formation are very similar east of Muir Inlet. Two pebble counts on the upper member (table 3) near Wachusett Inlet were high in plutonic rocks probably indicating a source to the north or northwest (fig. 2). As would be expected for an outwash deposit, rounding is considerably higher than the till and kame terrace samples and slightly higher than the esker samples.

Glacier Bay Formation

The Glacier Bay Formation is Neoglacial in age, as defined by Haselton (1966). It includes the till which is identical to the "Little Ice Age Till" of Goldthwait (1963), Taylor (1962a, b) and Price (1964). The original definition of Glacier Bay Formation was expanded by McKenzie (1970) to include all ice-contact deposits formed with the retreat of Neoglacial ice, and this expanded definition is used herein. The advance of Neoglacial ice into the Burroughs Glacier area was dated by a radiocarbon date on a stump buried by gravel between the present Burroughs and Cushing Glacier margins. The date of 3090±250 yr B.P. (W-2017) gives a maximum age for the ice advance to this position. Two dates at the southeastern margin of the Burroughs Glacier indicate when stumps of trees were pushed over by the ice. One, at the base of station 7 (200 m above sea level), is 2735±160 yr B.P. (I-59-15) and the other, at the base of Nunatak A near the present ice margin, is dated at 2520±87 yr B.P. (OWU-489). If only these two dates are used and an ice profile similar to that in 1960 is assumed, the rate of thickening of the glacier was on the order of 0.5 m/yr.

Physical characteristics of the Glacier Bay Formation are discussed in detail in chapter III and its distribution southeast of Burroughs Glacier is shown in Figure 5. Although ice did retreat and readvance during the Neoglacial in Adams Inlet (McKenzie, 1970), there is no indication that the Burroughs Glacier area was ice free any time between 2700 yr B.P. and its deglaciation during the present century. The history and mode of deglaciation and its geologic record comprise the remainder of this paper.

CHAPTER II

HISTORY OF DEGLACIATION

The earliest maps of the Burroughs and Plateau Glacier area were published by H. P. Cushing (1891) and H. F. Reid (1892), as a result of surveying done in 1890. In 1896 a topographic map with contours drawn over much of the area (fig. 6) was published by Reid. Since that time numerous maps have been produced which, combined with photographic records, yield a detailed history of deglaciation. Since 1926, field parties sponsored by the American Geographical Society have photographed and surveyed ice margins in the area and these and other observations have been summarized by Field (1947, 1959).

Place names have changed somewhat from map to map. The area of the present Burroughs and Plateau Glaciers was first named the Northwestern Branch of the Muir Glacier and later called Cushing Plateau (fig. 6) by Reid (1896). The latter name was used on International Boundary Commission maps and U. S. Coast and Geodetic Survey maps until 1942. The name Plateau Glacier was first used by Cooper (1923). The terms Burroughs Glacier, Bruce Hills, Minnesota Ridge, and Curtis Hills were first used by Field (1947) and were used subsequently on U. S. Geological Survey topographic maps issued in 1948 and 1950 (fig. 7). The remainder of the terms used in the following descriptions are informal and are either new or are the same as those used by Taylor (1962a, b) or Field in unpublished manuscripts. The locations of these features are shown in Figure 10.

Deglaciation Before 1892

Neoglacial ice reached its maximum extent in the Bartlett Cove area, 60 km south of the study area in about 1700 A. D. (Goldthwait, 1966a, b). The thickness of glacier ice in the Burroughs Glacier area is not known but some estimates can be made.

Taylor (1962a, b) reported a lichen trim line at an elevation of 460 m on the north side of Idaho Ridge south of Wachusett Inlet, but it seems unlikely that it represents the highest extent of ice during Neoglacial time. Reid's 1896 map shows that the ice margin in 1892 was at least this high on Idaho Ridge. Air photographs taken in 1964 show depositional ridges and glacial erosion features suggestive of glaciation to an elevation of 850 m on Minnesota Ridge.

McKenzie (1968) suggests that Neoglacial ice remained at its maximum until sometime between 1817 and 1842. If it is assumed that the lowering of the surface of Cushing Plateau began by 1842 and that it was lowered at a rate equal to that between 1892 and 1942, then the ice surface

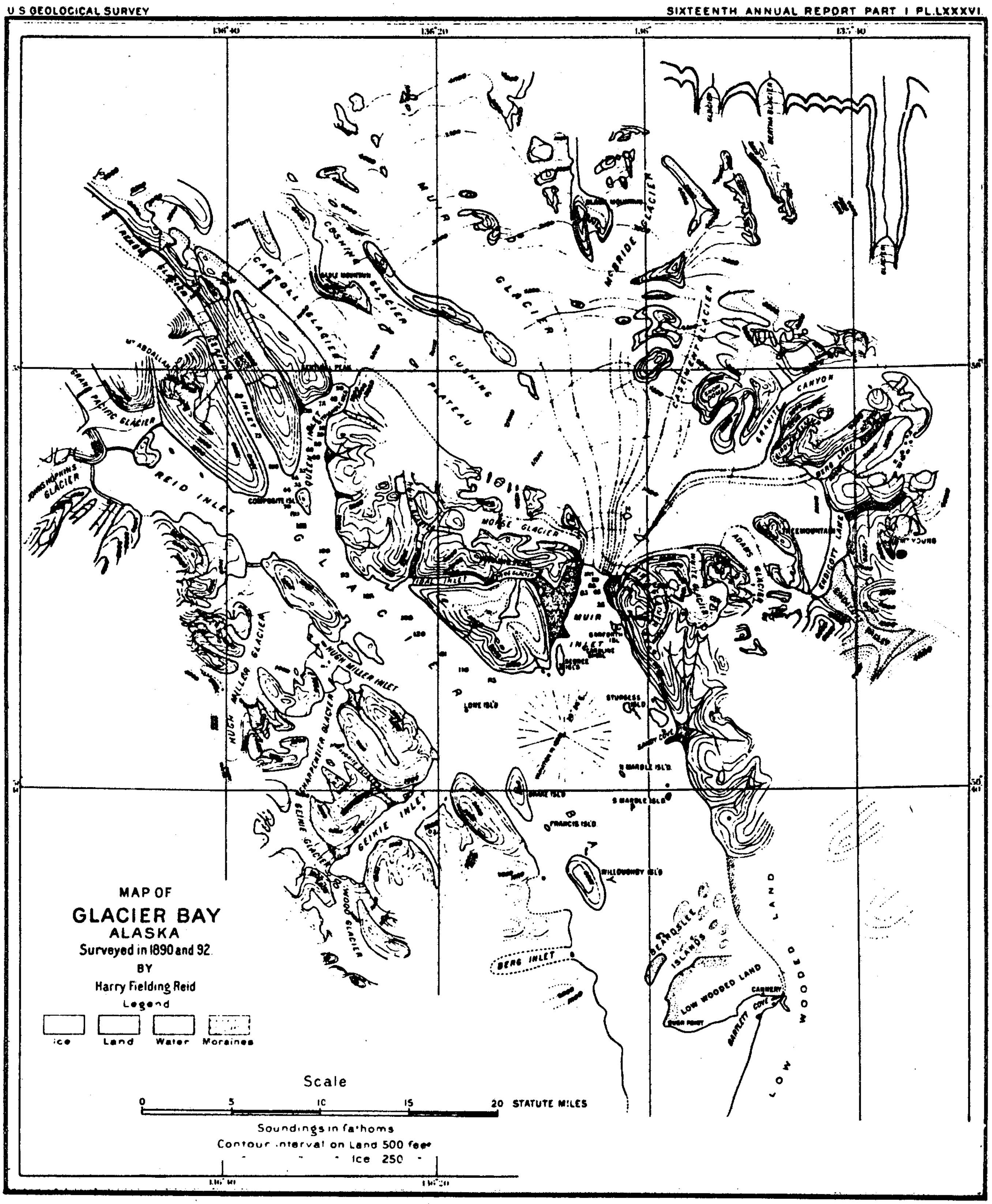


Figure 6. Map of Glacier Bay in 1890-1892 by H. F. Reid (1896).

must have been at least 700 m high along Idaho Ridge and at least 850 m high in the area of the present Burroughs Glacier. Thus, during the Neoglacial maximum, the Cushing Plateau probably was connected with the Muir Glacier throughout its length across Minnesota Ridge. Field (1959) suggested that the firn line at this time was probably no lower than 760 m.

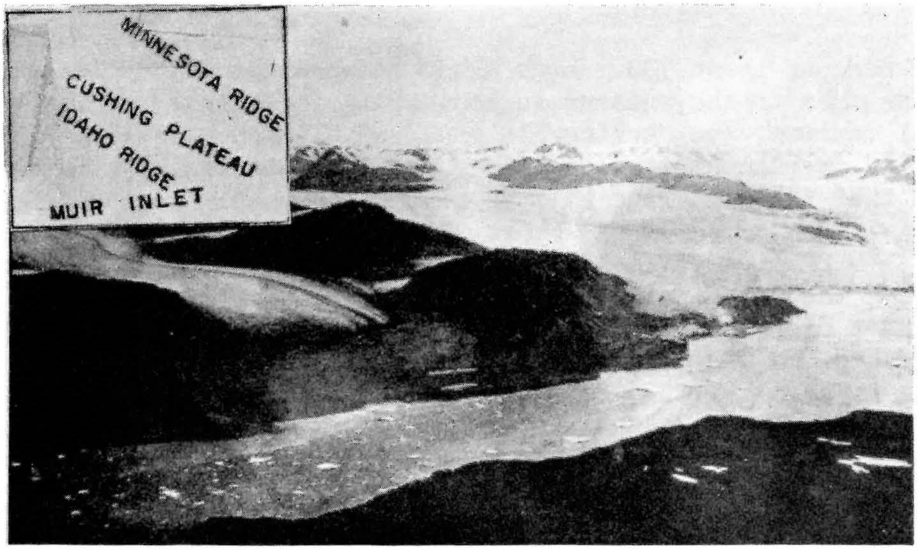
By 1892 the firn line was evidently above the crest of Cushing Plateau and a distinct ice divide had developed with ice draining both to the west toward Carroll Glacier and Queen Inlet and to the east toward Muir Inlet. Taylor (1962a, b) suggested that in early Neoglacial time this divide did not exist and that increased flow through Glacier Pass from Muir Glacier created a divide in the area of Burroughs and Plateau Glaciers which grew above the firn line and was enhanced by local accumulation. The presence of this local accumulation area is suggested by a series of concentric dirty-ice bands on the Burroughs Glacier; however, as pointed out by Taylor, these bands may have been derived from higher on the Cushing Glacier and the basin-like appearance of the bands may be a result of the present shape of the ice.

An alternative explanation is that the Cushing Plateau was never an isolated accumulation area and that the shape of the ice was determined by ice drainage patterns. An examination of Reid's map (fig. 6) shows that by 1892 the ice margin had retreated 35 km up the west arm of Glacier Bay to a position in Queen Inlet, while the Muir terminus had retreated only 12 km. This more rapid retreat of the western ice stream, probably due both to its wider calving terminus and to a smaller supply of ice than the Muir System, would have caused a marked lowering of the Carroll Glacier relative to the Cushing Plateau and initiated a reversal of flow at the west end of the Cushing Plateau.

Deglaciation During the Period 1892 - 1948

Retreat of Ice Margins

The pattern of deglaciation since 1892 can be traced on Figure 7. By 1907 the calving margin of Muir Glacier had retreated nearly 10 km from its 1892 position and the highest peaks of the Bruce Hills and Curtis Hills had emerged. A photo taken in 1907 from Mt. Wright, southeast of the area, and an oblique air photo taken in 1965 are shown in Plate I. The Muir terminus retreated past the mouth of Wachusett Inlet in about 1915, and the retreat of Plateau Glacier after that time was about 150 m/yr to its 1929 position (Field, 1959). Burroughs Glacier remained connected to Muir Glacier by an ice tongue just north of the Curtis Hills until about 1920. By 1929, peak 2040 of the Bruce Hills had emerged and became an important factor in determining ice flow directions in the southeastern part of the Burroughs Glacier. Station 7 emerged from the ice just before 1929 (fig. 7).



(a)

(b)

Plate I. Photos taken (a) in 1907 (by the International Boundary Commission) and (b) in 1965 (by A.Post).

The tidal terminus of Plateau Glacier retreated at an average rate of 145 m/yr between 1929 and 1940. Retreat of the ice tongue on land north of station 7 was approximately 21 m/yr between 1926 and 1941 (Field, 1959); the col between peak 2221 and peak 1960 on the Bruce Hills became ice free during this time and ice in the Curtis Hills had withdrawn to form two ice tongues north and south of station 7 (fig. 7).

Subsequent ice retreat during the period between 1940 and 1948 resulted in the deglaciation of the col between peaks 2221 and 2040 (fig. 7) on the Bruce Hills and the hill on which stations 14 and 15 (fig. 10) are located. The margin of the Plateau Glacier retreated about 221 m/yr between 1941 and 1948. The ice tongue north of station 7 retreated about 915 m between 1941 and 1948 (Field, 1959).

Ice Surface Lowering

In 1892 the crest of the Cushing Plateau was the same elevation over both the present Burroughs and Plateau Glaciers. At this time the Muir Glacier was higher than the Cushing Plateau and flow took place across Minnesota Ridge. By 1907, however, ice on both sides of Minnesota Ridge was at nearly the same elevation and ice probably was not flowing across the ridge. Since that time ice surface over the Burroughs Glacier has dropped significantly less than that of Plateau Glacier, probably because of increased ice loss through calving and higher ablation at the lower elevations of Plateau Glacier.

Although early contour maps are probably somewhat in error, an impression of the relationships between downwasting and backwasting can be obtained from long profiles of the Burroughs (fig. 8) and Plateau (fig. 9) Glaciers. Position of the lines was determined by triangulation from prominent landmarks on the map and some error may be due to the location of the profiles. The 1907 map of the International Boundary Commission is not used except in the eastern end of the Plateau Glacier profile because in the rest of the profiles it is below the 1941 and 1948 levels. The elevations of some bedrock hills in 1907 do not agree with later maps, suggesting that the map is in error. The 1958 ice divide on Plateau Glacier is estimated from data given by Field (1959). Although calculating annual rates of lowering from the profiles is probably unrealistic because of errors in elevations and the large time span between the 1892 and 1941 profiles, several interesting points can be noted.

Since sometime not long before 1892, ice has flowed in both directions from the crest of the Cushing Plateau. This flow, however, has not been enough to offset ice loss near the margin due to calving (on the Plateau Glacier and on the Burroughs Glacier before about 1920) and ablation. Thus ice surface lowering has been greater near the margins than at the crest with the exception of the period between 1941 and 1948. The elevation of the crest in 1941 is probably in error as other measurements by Field (1959) indicate that surface lowering between 1935 and 1948 was 2 to 5 m/yr. Map values yield a rate of lowering of nearly 8 m/yr.

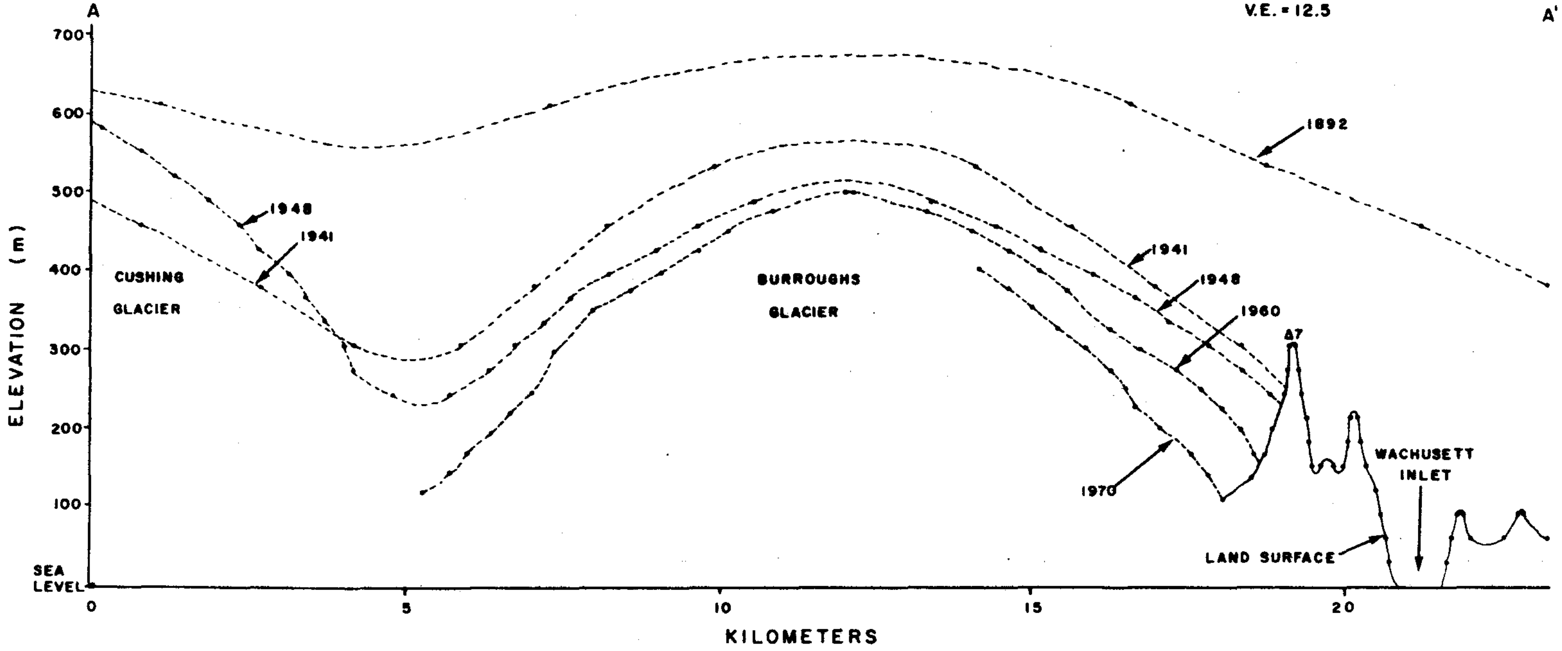


Figure 8. Ice-surface profiles along axis of Burroughs Glacier. Position of profile is shown in Figure 7.

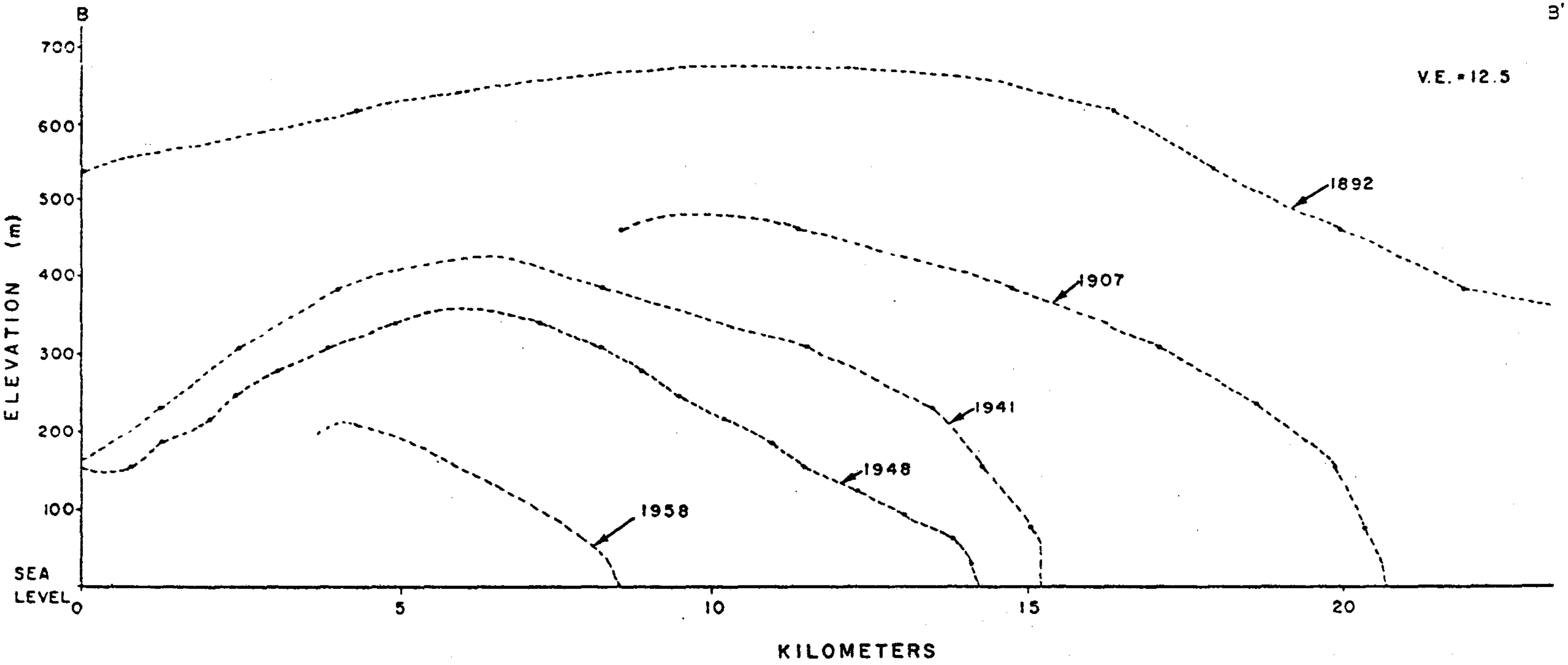


Figure 9. Ice-surface profiles along axis of Plateau Glacier. Position of profiles is shown in Figure 7.

It can be seen that the 1941 profile is lower than the 1948 profile on the Cushing Glacier (fig. 8). This could be an error in mapping but bedrock points in the area agree fairly well. Field (1947) reported that the Carroll Glacier, parallel to and sharing an accumulation area with the Cushing Glacier, surged shortly after 1941. The Cushing Glacier may have also surged at this time producing a steeper ice front on the Cushing in 1948. As would be expected, no effect of the surge is seen on the profiles of the Burroughs and Plateau Glaciers.

The crest of the Plateau Glacier shifted toward the northwest as deglaciation proceeded. This may be the result of more rapid flow toward the calving terminus in Wachusett Inlet than toward the Carroll Glacier and Queen Inlet to the west.

Deglaciation During the Period 1948 - 1970

Retreat of Ice Margins

The tidal terminus of Plateau Glacier retreated about 530 m/yr between 1948 and 1960 and 350 m/yr between 1960 and 1970. An ice mass, herein called Plateau Remnant, was stranded on the north side of Wachusett Inlet. Its eastern terminus retreated about 140 m/yr over gently rolling topography between 1960 and 1970 (plate II).

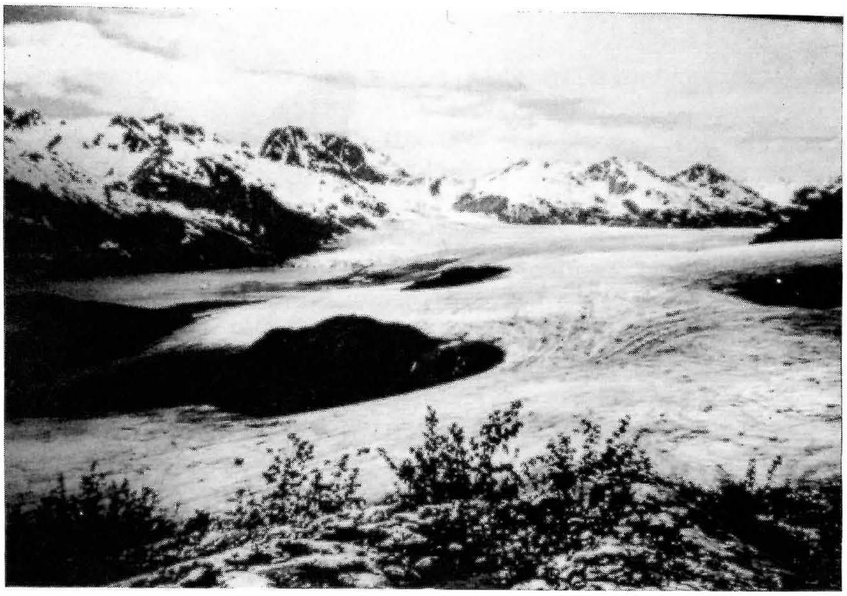
Sometime between 1950 and 1958, the northwest terminus of Burroughs Glacier separated from Cushing Glacier. Retreat of this margin of Burroughs Glacier between 1960 and 1970 averaged 50 m/yr. This much lower rate of retreat than that of Plateau Remnant is probably due to the addition of ice from the high part of the Burroughs Glacier and to its northwest exposure. Ice was still present, although separated from both the Muir and Burroughs Glaciers, in Glacier Pass in 1970. This small ice mass will probably disappear within a few years.

Ice margin positions for 1948, 1960, 1964, and 1970 have been compiled from air photographs of the southeast terminus of the Burroughs Glacier (fig. 10). Rates of margin retreat range from 9 m/yr to over 60 m/yr between 1960 and 1970 depending on land slope, ice surface slope, and exposure. In 1948 no nunataks showed through the ice near the southeast terminus of Burroughs Glacier although their effect on ice surface slope is obvious on the 1948 topographic map. In 1950 Nunataks A, B and D were just showing through the ice (plate II). By 1958 six nunataks (A, B, C, D, E and F) were exposed. After the nunataks emerged, increased ablation caused a rapid retreat of the margin surrounding the nunataks.

Ice Surface Lowering

Ice surface lowering between 1960 and 1970 along a profile northwest of station 7 (fig. 11) ranges from 4.6 m/yr where ice was at an

a. Taken in 1950 by
W. O. Field



b. Taken in 1960 by
L. D. Taylor.

c. Taken in 1969. Note
emergence of nunataks.

Plate II. Photographs of Plateau Glacier and the southeast terminus of Burroughs Glacier.

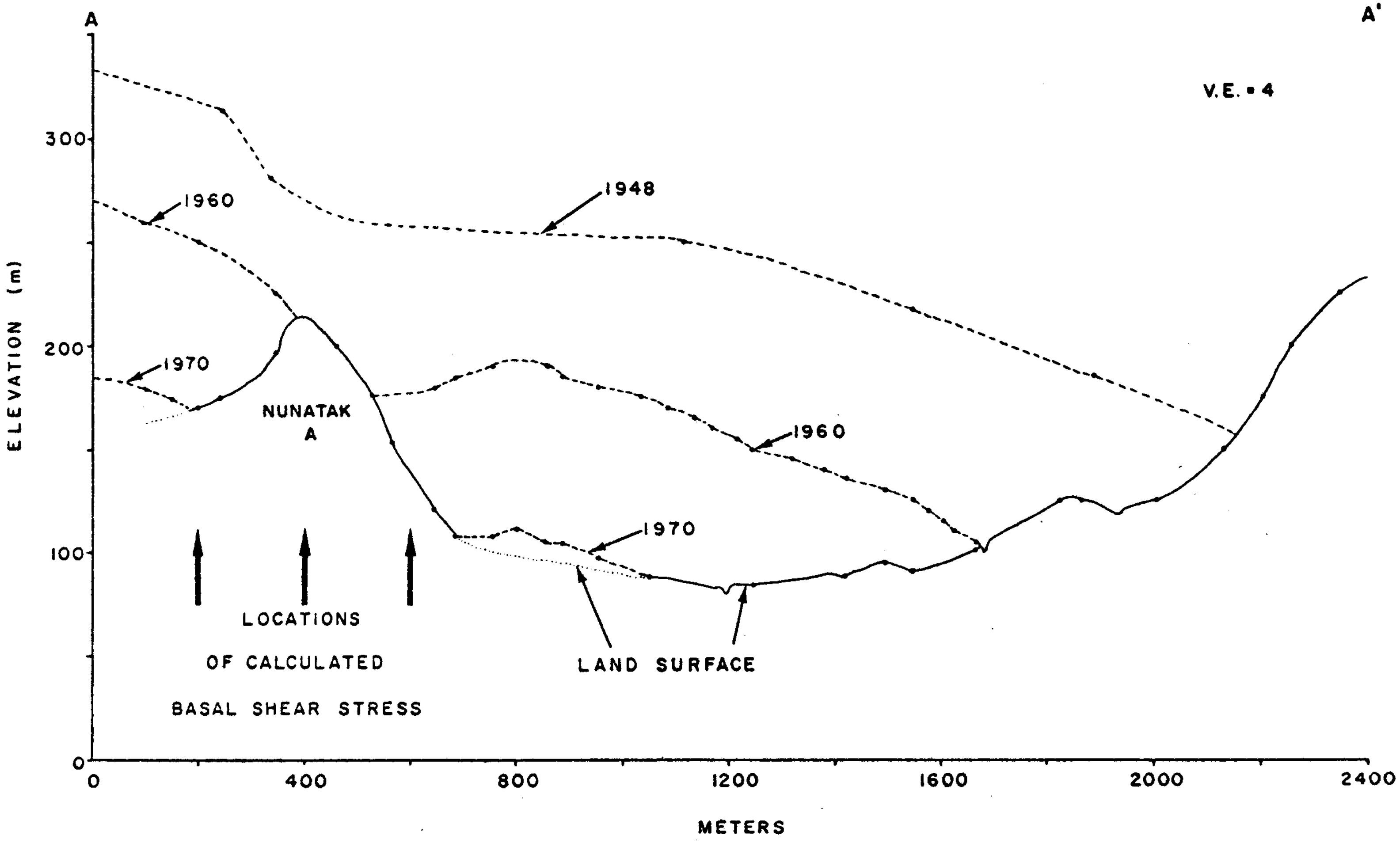


Figure 11. Ice-surface profiles over Nunatak A. Position of profiles is shown in Figure 10.

elevation of 440 m in 1960 to 9.5 m/yr where ice was at an elevation of 205 m in 1960. Similar values are obtained from other profiles. There is a relationship between elevation and ice surface lowering; however, slope also affects the rate of lowering considerably (fig. 12). The greatest deviation from a straight line relationship is where there is a flattening of the ice surface which reduces the ablation rate. Although no aerial photographs of the crest area are available, Taylor (1962a, b) reported a rate of lowering of 0.8 m/yr between 1948 and 1960. This relationship between elevation and rate of ice lowering indicates that ice flow is insufficient to offset differences in ablation rate with change in elevation. Thus the ice surface has had a continually increasing surface gradient. When the curve is projected to zero melting an elevation of nearly 700 m is obtained for an estimate of the firn line elevation.

A marked steepening of the ice on the stoss side of nunataks in 1948 is evident (fig. 11, 13). This steepening may have been present in 1941 but it is not evident on a map with 250 ft contours. The steepening occurs on the stoss side of the nunataks rather than directly above them as would be expected, and is probably an effect of the high ablation rate. If ablation on the stoss side of Nunatak A averaged 6 m/yr between 1948 and 1960, the slope retreat in the horizontal direction would be an order of magitude larger than surface movement velocities measured by Taylor (1962a, b). Thus the steep slope has retreated from its expected position over the nunatak.

The appearance of the nunataks causes increased ablation of ice surrounding the nunatak due to heat gained from long wave radiation reflected from the nunataks. This gives rise to a dome-shaped, nearly stagnant ice mass on the lee side of each nunatak which eventually separates from the main ice mass. Taylor (1962a, b) measured surface velocities of 0.1 m/yr on the ice mass south of Nunatak D and 0.7 m/yr on the ice mass south of Nunatak A.

Changes in Ice Flow Direction

Because of differences in rates of retreat between Plateau and Burroughs Glaciers and because of the emergence of the Bruce Hills, major changes in ice flow direction have taken place at the southeast end of the Burroughs Glacier. If the horizontal component of ice flow is assumed to have been in the direction of the ice-surface slope, ice flow in 1892 over the whole southeastern Burroughs Glacier was from the west-northwest (fig. 14). In 1907 direction of ice flow had shifted slightly and by 1929 (observation from oblique aerial photo), ice was flowing from the northwest. This general change has continued until the present time (fig. 14).

As nunataks emerged after 1948, flow direction evidently changed considerably around the nunataks but surface velocities in 1960 were only in the range of 0.1 - 3.0 m/yr. The geologic record of changes in ice flow direction is discussed in Chapter III.

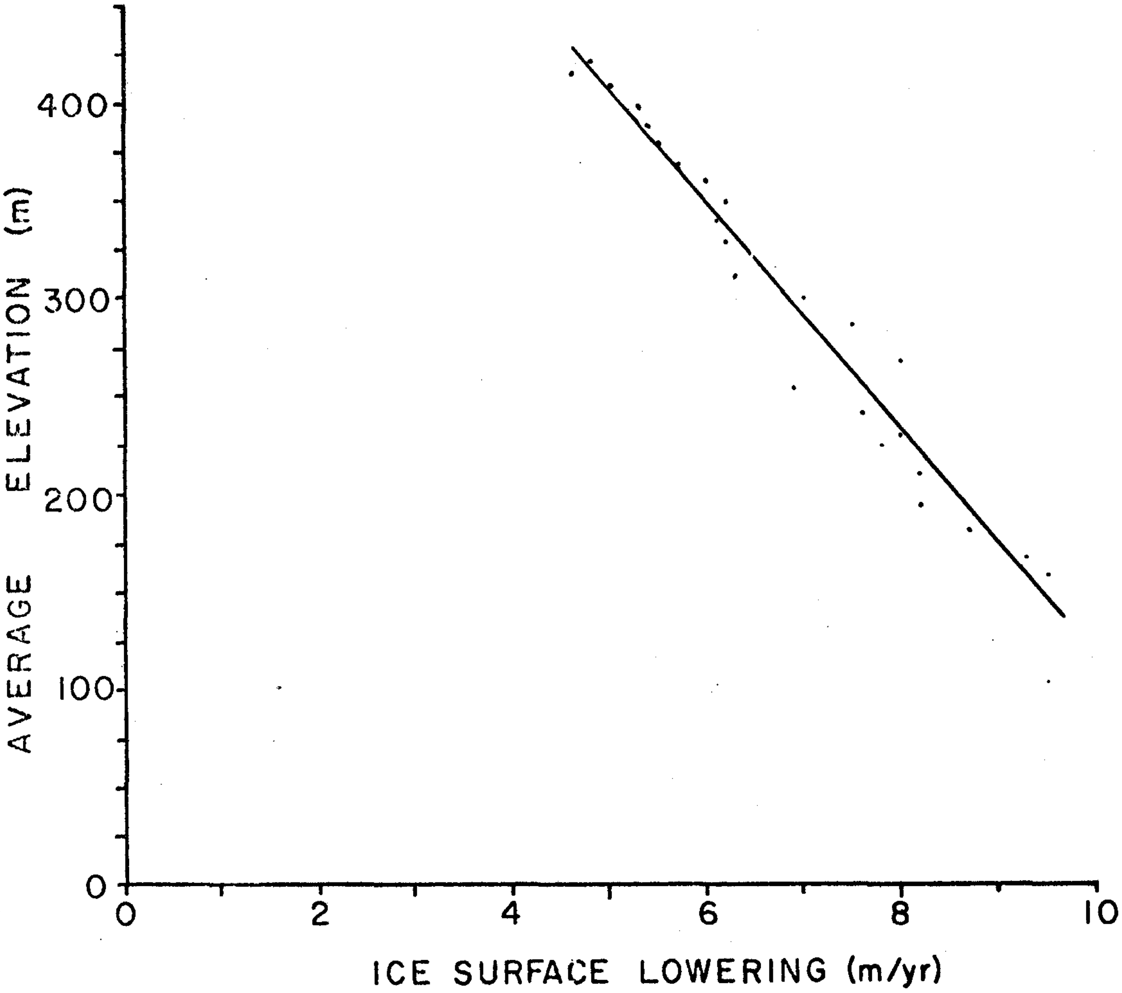


Figure 12. Relationship between rate of ice-surface lowering, 1960 to 1970, and mean of 1960 and 1970 elevations.

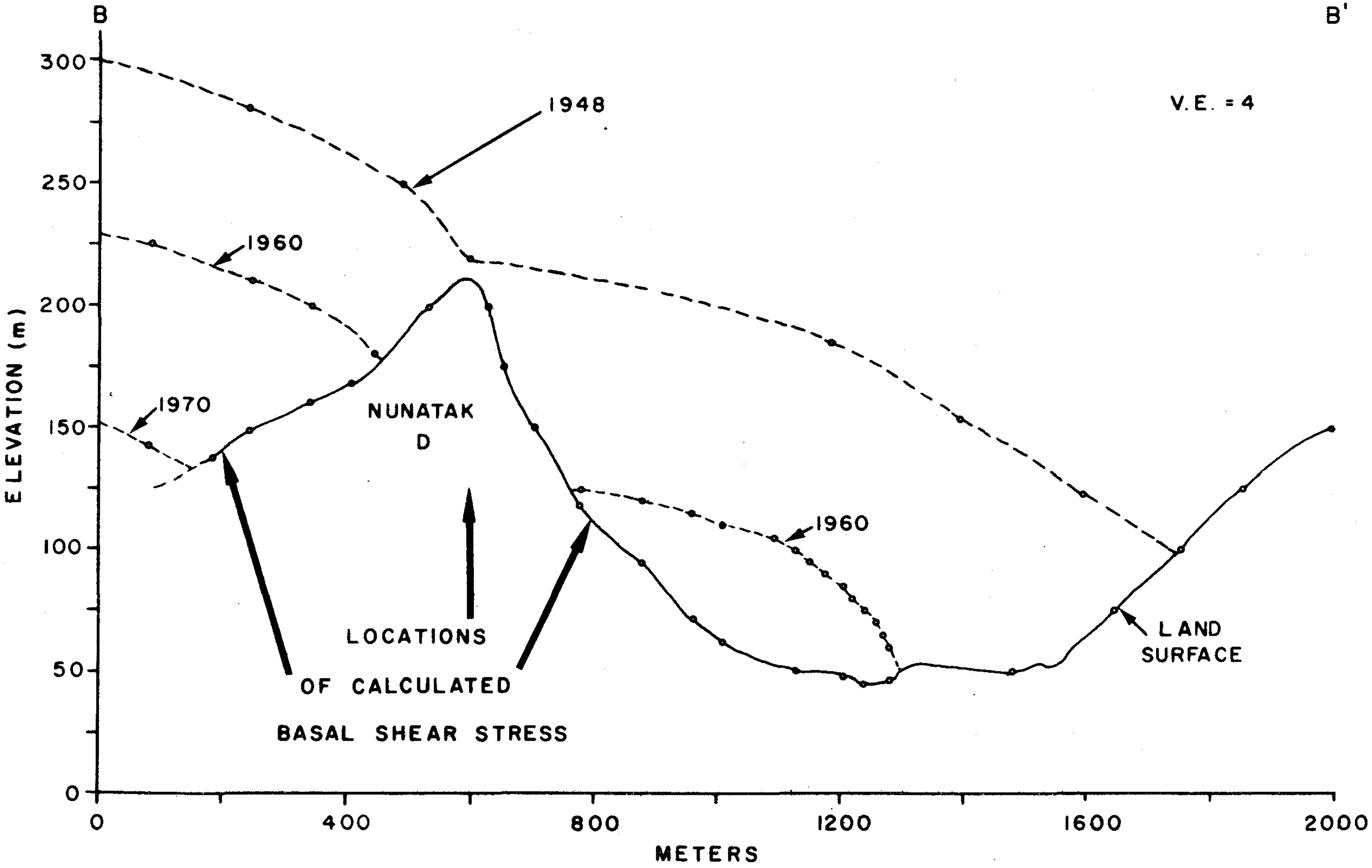


Figure 13. Ice-surface profiles over Nunatak D. Position of profiles is shown in Figure 10.

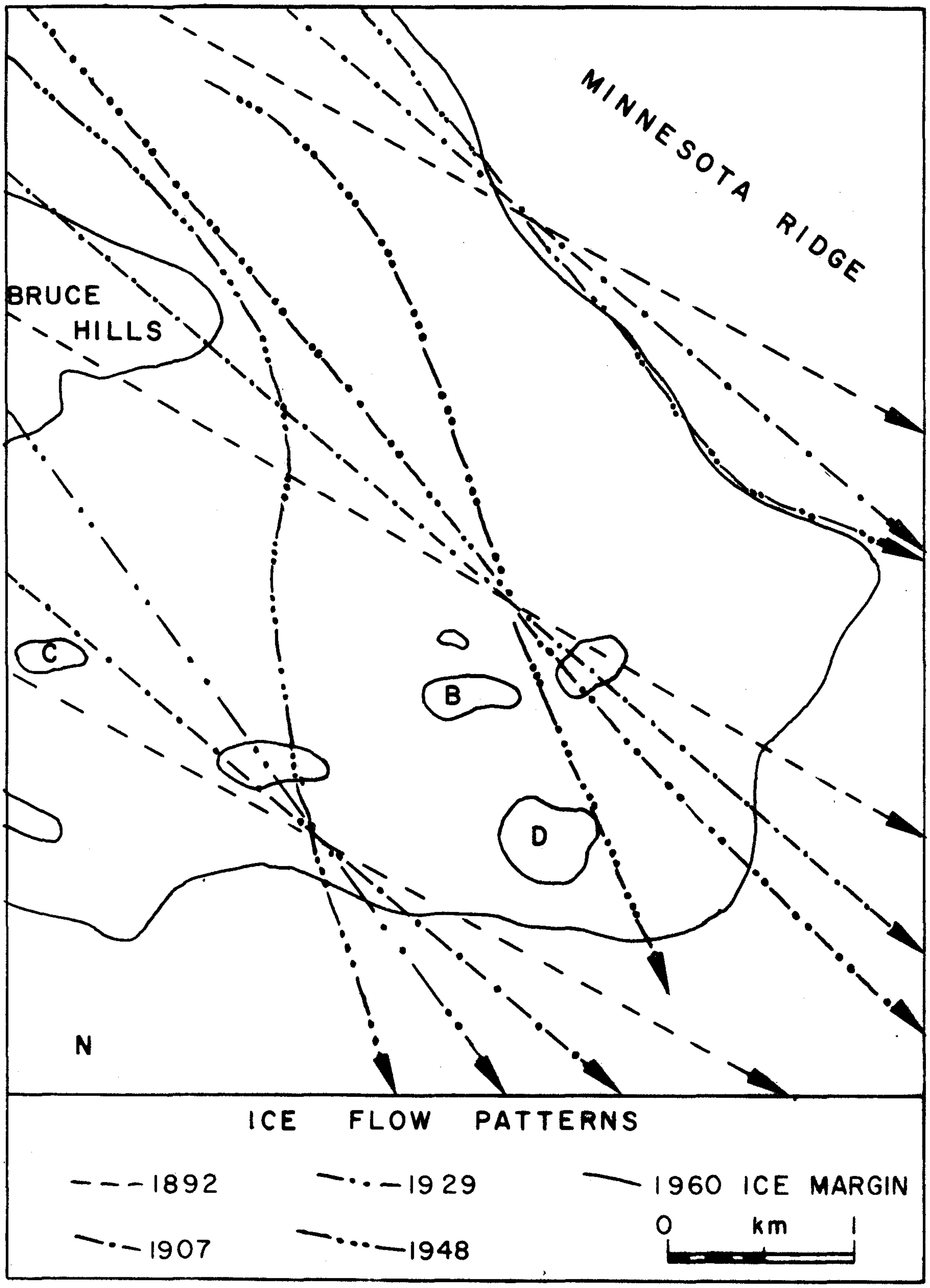


Figure 14. Changes in the direction of ice flow between 1892 and 1948.

CHAPTER III

TILL

Basal Till

Nature of Till and Rate of Till Deposition

The till of the Glacier Bay Formation has a maximum thickness of more than 10 m and is unleached and unoxidized in the study area. Moist colors range from gray (7.5YR 6/2) to olive gray (5Y 4/2) and a platy or blocky structure is generally present. As previously described, ice flow directions changed as deglaciation proceeded in the study area. Changes in fabric and composition accompanied this change and these are described in the following sections. The location of sample collection sites is shown in Figure 15. Because the rate of change of ice-flow direction is known till fabrics at different depths provide the first known rates of basal till deposition.

Particle Size Distribution

Analyses of 98 samples, 38 of which are till from the Glacier Bay Formation are tabulated in Appendix A. Standard hydrometer procedures were used to determine the percent less than 2 microns (American Society for Testing and Materials, 1958). Sand fractions were segregated by dry sieving after the total sand fraction had been separated by wet sieving.

In the Glacier Bay till, sand percentages range from 32 to 81, silt from 16 to 62 and clay from 3 to 15 with a mean of 57 percent sand, 36 percent silt and 7 percent clay (fig. 16). As discussed in a later section, particle size varies considerably with depth at any given location. Although only two samples of Granite Canyon Till were analyzed, both fall well within the range for the till of the Glacier Bay Formation, suggesting that the two cannot be distinguished on the basis of particle size in this area.

Clay Minerals

Clay minerals (Appendix B) were identified by X-ray diffraction of (a) air-dried, (b) ethylene-glycolated, and (c) heated (at 400°C) and (d) heated (at 550°C) samples of the <2 micron fraction of 19 till samples. The clays were first saturated with magnesium, resuspended,

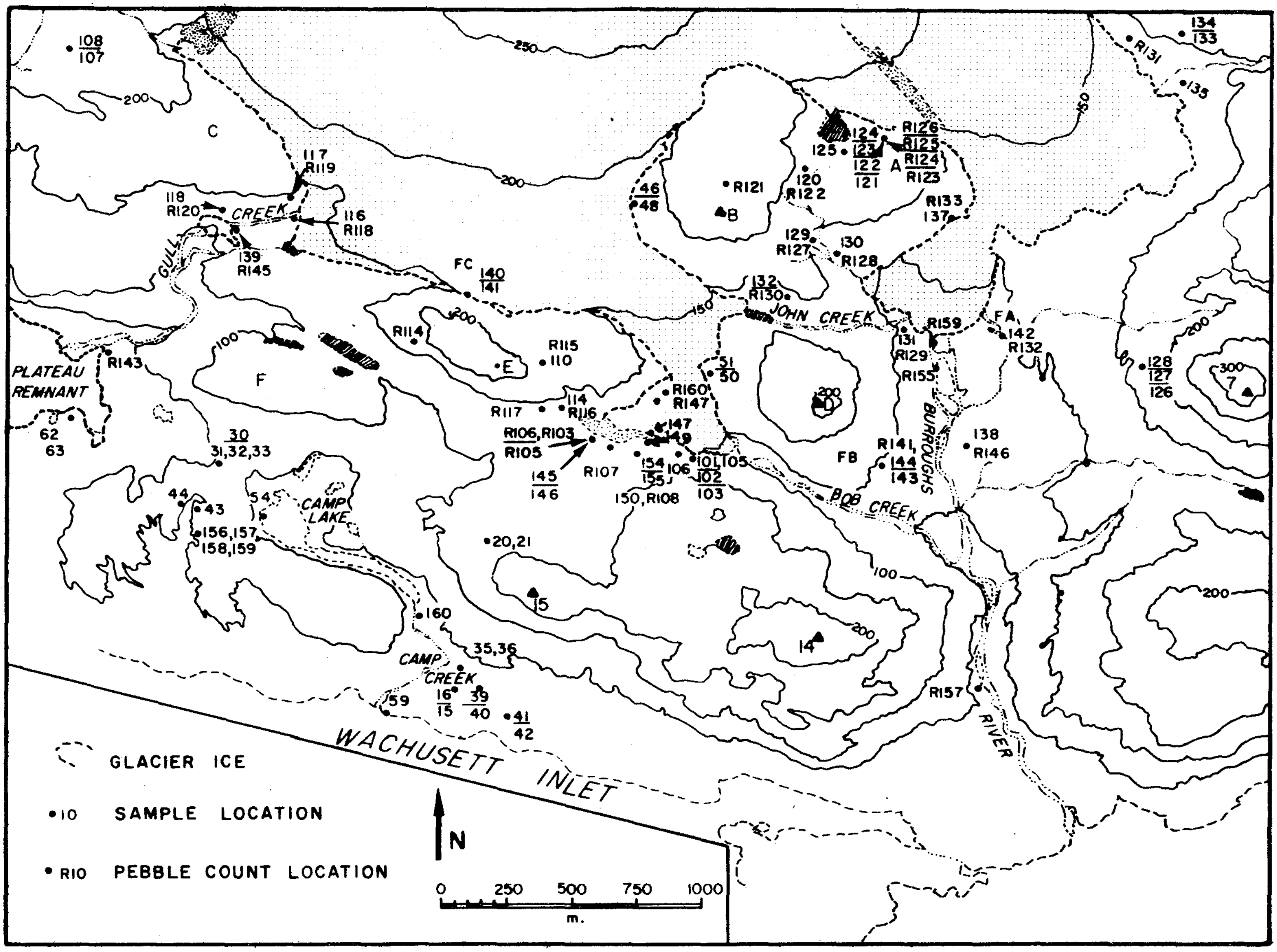


Figure 15. Sample collection sites and location of three till fabric sites (FA, FB, FC). Lines between numbers indicate samples collected one above the other.

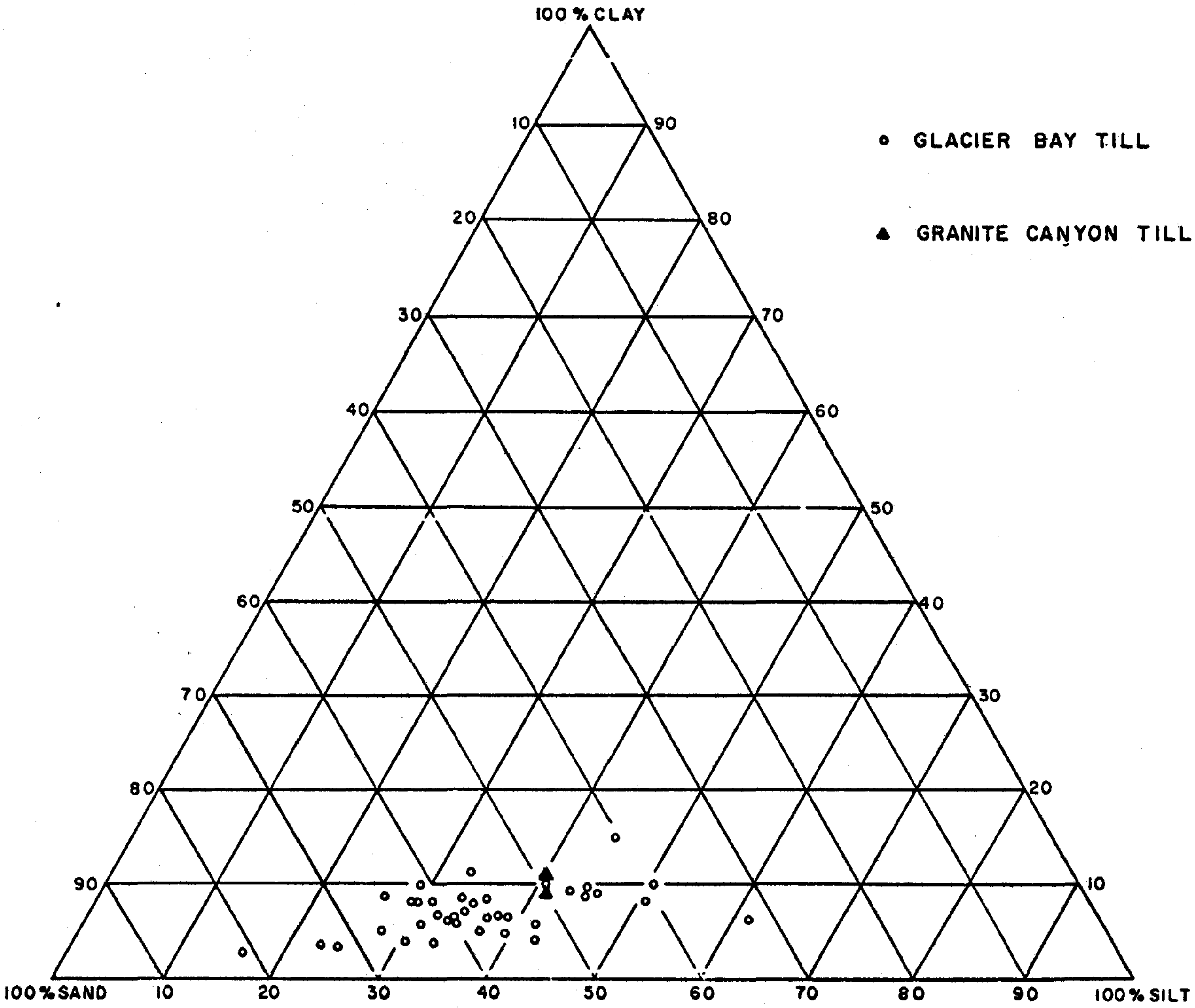


Figure 16. Sand-silt-clay diagram for Granite Canyon Till and till of Glacier Bay Formation.

and then plated in approximately 1 mm thick layers on ceramic plates by applying a vacuum to the base of the plate and allowing water from the suspension to be drawn through. Semiquantitative extimates of abundances are based on peak areas after the method of Johns et al. (1954) modified by Wilding and Drees (1966, unpublished report).

The following relationships were used in determining relative abundances:

Illite = Area of 10 Å peak (E.G.) X 4

Vermiculite = 2 (Area of 14 Å peak (400°) - Area of 19 Å peak (E.G.))

Chlorite = Area of 14 Å peak (400°) X 2

Quartz = Area of 3.3 Å peak (A.D.) - 0.75 (Area of 10 Å peak (A.D.))

Expandable Clays = Area of 14 Å peak (E.G.) - Area of 14 Å peak (A.D.)

where (A.D.) is the air-dried sample, (E.G.) is the ethylene-glycol-treated sample and (400°) is the heat-treated sample with temperature in degrees centigrade. These abundances were then converted to percent of the total assemblage considered.

McKenzie (1968) determined that most of the 3.5 Å peak in tills in Adams Inlet could be accounted for by iron-rich chlorite and this is also the case at Burroughs Glacier. In calculating abundances, montmorillonite was included with other expandable clays. Since glycolation produced a broadening of the 14 Å peak toward 16 Å the montmorillonite may be interlayered with chlorite. Fairly large feldspar peaks were present for the clay fraction, however the feldspars, calcite, dolomite and amphiboles were not considered in the calculations of relative abundance. Mean and extreme abundances are given in Table 2.

The samples are all characterized by large amounts of chlorite and in some cases, quartz, relative to the other minerals considered. Relative abundances vary from those of McKenzie (1968) in that tills in Adams Inlet average 50% illite. O'Brien and Burrell (1970) report similar differences between the Muir Inlet area and areas to the east in relative amounts of illite and chlorite.

One sample of metasiltstone, the predominant rock type at the southeast end of the Burroughs Glacier (see fig. 2), was ground and analyzed by X-ray diffraction. A relatively high mica peak (10 Å) was present but the amount of illite in the till samples shows no relation to the amount of metasediment pebble counts from the same localities. This suggests that in such a short distance, little of the mica was ground to the clay fraction, but a large number of bedrock samples would have to be analyzed before this could be demonstrated.

Of the clay minerals whose relative abundances were considered, only quartz showed a correlation with lithologies determined in pebble counts. This correlates well (fig. 17) with the amount of granite in the pebble counts. No granite outcrop was found by the writer and it may be beneath the Burroughs Glacier or somewhere to the northwest.

Pebble Lithology and Rounding

Pebble lithology and rounding data are presented in Appendix C. Counts were made on 100 pebbles ranging in size from 1 to 10 cm at each location and lithologies were determined by hand sample identification. Plutonic rocks containing more than 10 percent quartz were grouped as granitic rocks; those with less than 10 percent quartz and more plagioclase than orthoclase were considered to be diorite; those with more orthoclase than plagioclase were considered to be syenite. Since distinction of plagioclase and orthoclase can be difficult in some cases there may be error in the estimates of percentages. Andesite-dacite porphyries were distinguished from basalt. Limestones and high grade metamorphic rocks were not differentiated but white and black marble was counted separately. Rounding was determined by comparison to the rounding scale of Krumbein (1941).

Mean and extreme percentages of the various types are given in Table 3 for samples taken at the southeast terminus of the Burroughs Glacier and on Plateau Remnant. Fairly large percentages of local bedrock (metasiltstone) occur in the tills, presumably because the tills are thin and the ice is on bedrock in most places. This does not agree with the findings of Harrison (1960) and Anderson (1955) in the midwest where tills are much thicker and most of the till is derived from previously deposited till.

Since the position of bedrock contacts under the ice can only be postulated, no attempt was made to relate till lithologies with precise bedrock sources. In general, however, the percentage of siltstone increases down-ice from its contact with plutonic rocks across the Bruce Hills and Minnesota Ridge (fig. 2).

Heavy Minerals

The heavy mineral fraction (specific gravity > 2.96) of the very fine and fine sand fraction (0.0625-0.25 mm) of a number of till samples was examined. In the grain mounts only fine-grained rock fragments

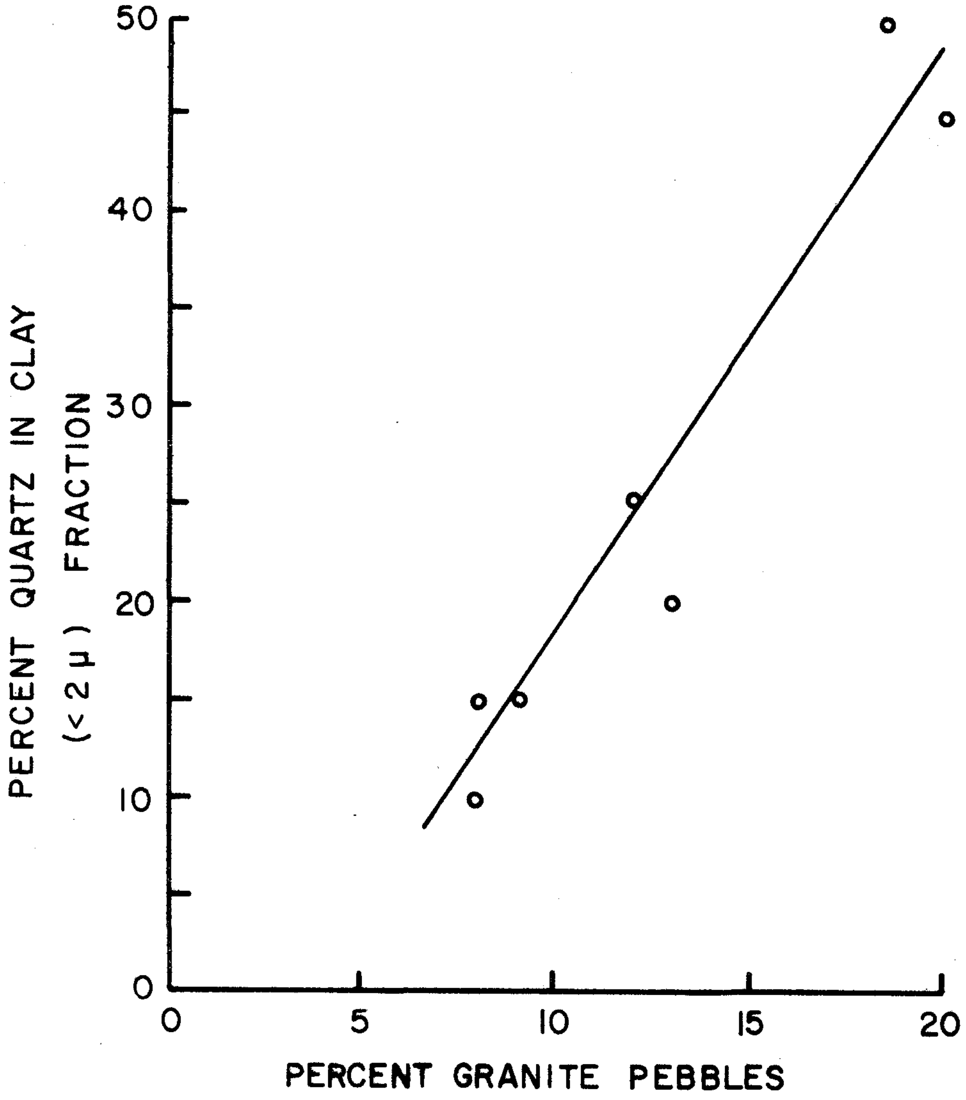


Figure 17. Relationship between amount of quartz in clay fraction and percent granite in pebble counts.

Table 3. Mean and extreme percentages of pebble lithologies, rounding, pebbles with fresh breaks, and pebbles with striations in tills.

	% Granite	% Syenite	% Diorite	% Basic Dike Rocks	% Acidic Dike Rocks	% Marble	% Metasiltstone	% Other	Plutonic/ Metasiltstone Ratio	Rounding (Krumbein Scale)	% Pebbles with Fresh Breaks	% Pebbles with Striations
Glacier Bay Till* N = 17												
Mean	10.8	1.3	10.6	22.1	16.5	0.3	35.0	3.3	1.03	0.31	15.1	8.3
High	21	9	25	31	29	2	51	21	5.50	0.42	32	18
Low	5	0	1	9	8	0	10	0	0.24	0.26	7	2
Granite Canyon Till N = 1												
	20	0	4	24	29	0	23	0	1.04	0.33	22	6
Ablation Till* N = 6												
Mean	25.7	2.7	18.8	19.2	17.8	1.7	10.8	3.2	6.25	0.33	23.7	4.0
High	34	4	25	28	20	5	20	4	14.50	0.36	32	9
Low	18	1	13	13	13	1	4	2	2.05	0.29	17	1
Ablation Debris N = 2												
Mean	1.0	0	1.0	10.5	3.0	38.5	38.5	7.0	0.06	0.17	3.5	0
High	2	0	1	14	4	71	59	13	0.06	0.19	7	0
Low	0	0	1	7	2	6	18	1	0.05	0.14	0	0

* Does not include those south of Wachusett Inlet

similar in appearance to the metasiltstones examined in this section were counted for most samples and the percentages out of a total of 100 grains is given in Appendix A. Hypersthene, hornblende, augite and diopside, chlorite, and opaque minerals are also present. Very few uniaxial or isotropic minerals were noted.

Although it is likely that some grains derived from fine-grained dike rocks were included in the counts, there is a correlation between the percent of metasiltstone in the pebble counts and the percent of rock fragments in the heavy mineral fraction (fig. 18). Scatter is great but large changes in the percentages of metasedimentary rock fragments do represent changes in the percentage of metasiltstone in the till. This allows the percent of rock fragments to be used as an indicator of the amount of local bedrock in areas where no pebble counts were done.

Carbonate Minerals

Although limestone was found in only one small outcrop, small percentages of carbonates were found in all till samples. The Chittick apparatus (Dreimanis, 1962) was used to analyze carbonate content in samples of the <2 mm fraction of the till of the Glacier Bay Formation. Results of all samples are given in Appendix D. Leaching is probably negligible except right at the surface because of the age of the tills (Ugolini, 1966); therefore, differences in carbonate content probably represent differences in till provenance. The highest percentages found are located down-ice from Nunatak E, where the small limestone outcrop was observed.

Till Fabrics

The orientation of pebbles with a:b axis ratios of 1.7:1 and long axes between 1 and 10 cm long was measured in the field. Strike and plunge of the long axis was measured on 50 pebbles at most locations. These were plotted and contoured on an equal area projection using a computer program developed by C. E. Corbató (personal communication) after Kamb (1959). The computer program is given by McKenzie (1968). The lowest contour is one standard deviation and the contour interval above this is given for each diagram.

The fabrics indicate ice movement from the northwest and reflect the ice-movement direction changes that were discussed in Chapter II (fig. 19). In order to demonstrate the relationship of fabric to late ice-flow direction, fabrics were measured on near surface tills as close as possible to the positions of ice-movement stakes of Taylor (1962a, b). In location A (fig. 15) the ice was approximately 65 m thick in 1960. The fabric and 1960 ice flow direction are shown in Plate III, a. Rate of flow was 0.7 ± 0.2 m/yr and the standard deviation

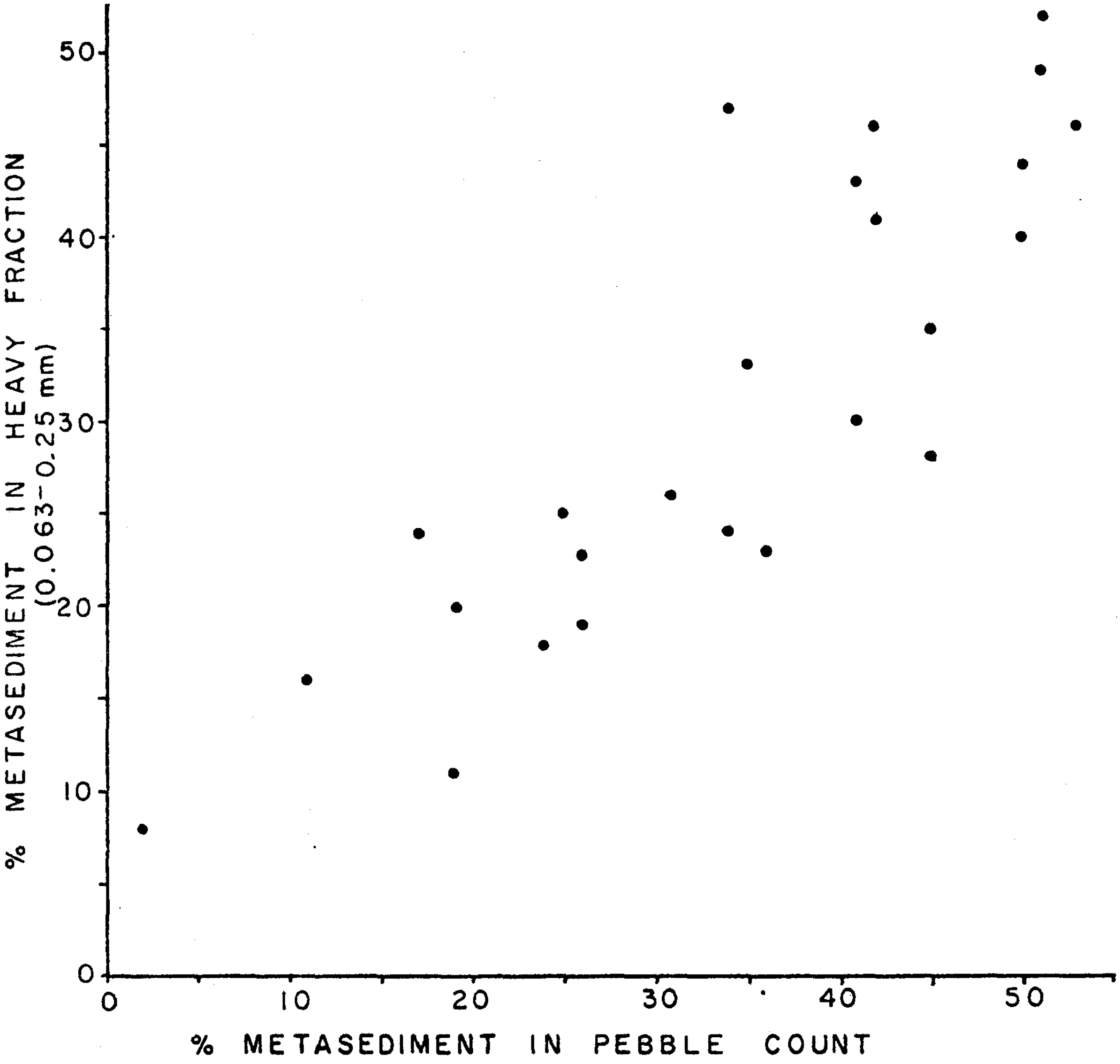


Figure 18. Relationship between percent metasedimentary rocks in pebble counts and percent metasedimentary rock fragments in heavy mineral fraction.

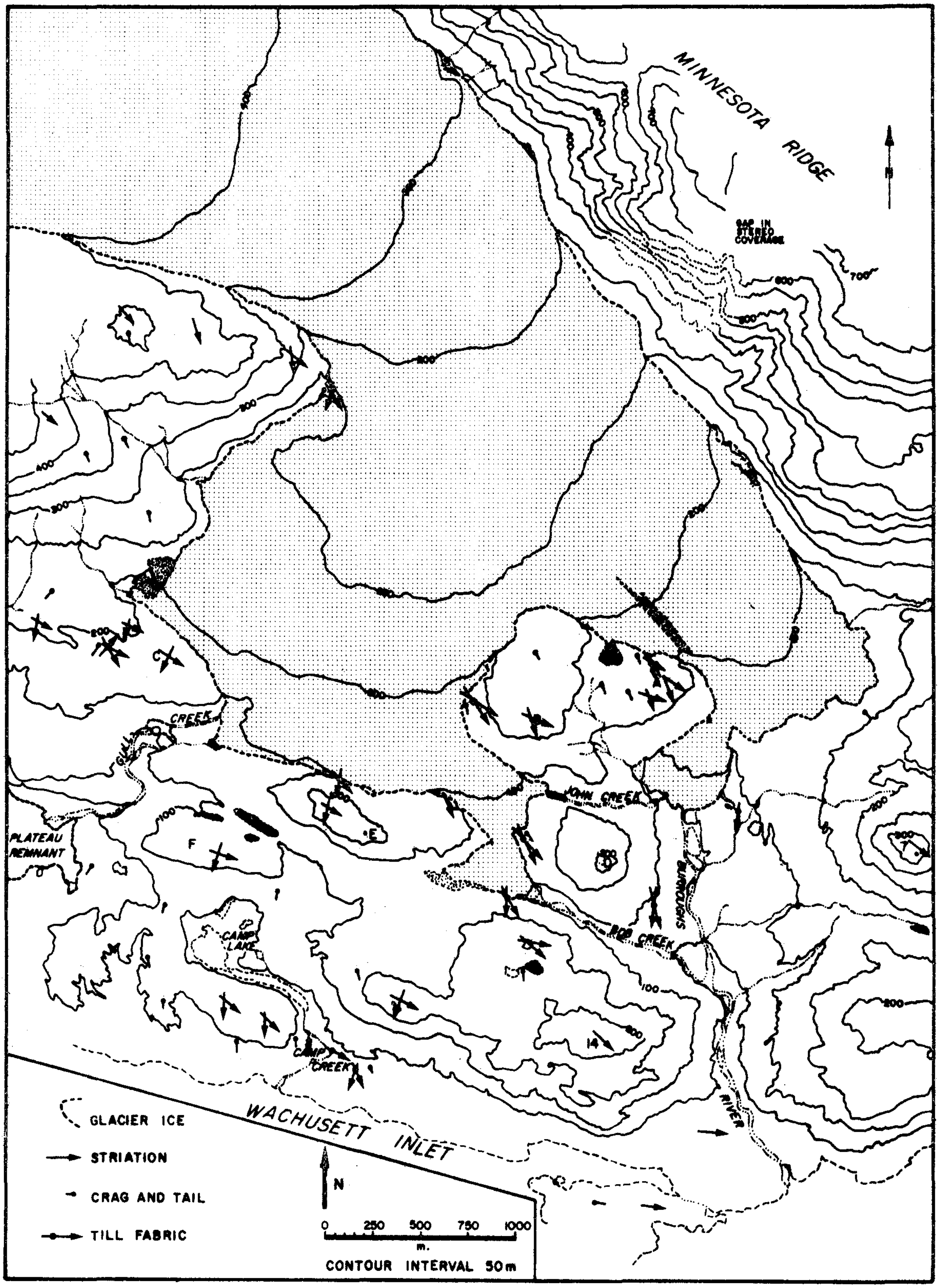
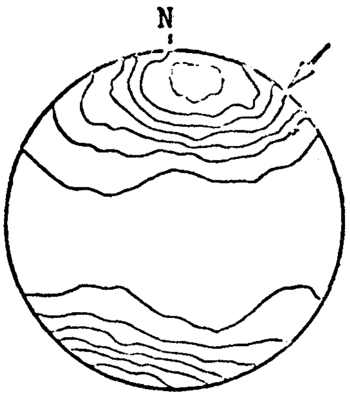
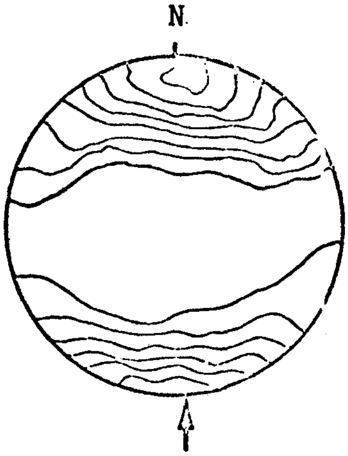


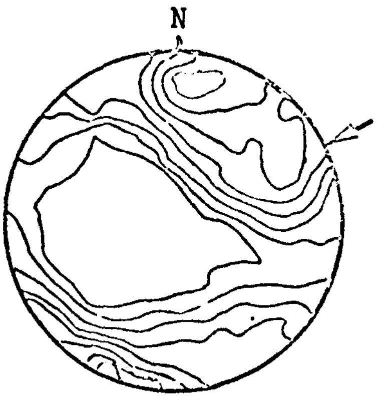
Figure 19. Directional features map of southeast terminus of Burroughs Glacier. Multiple till fabric measurements made in one location are discussed in text. Fabric measurements made in crag and tail features and till ridges not shown.



a. Fabric at Location A



b. Fabric at Location B



c. Fabric at Location C

Plate III. Fabric diagrams of surface till beneath 1960 ice movement stakes. Arrow indicates flow direction measured by Taylor (1962) in 1960. For location see Figure 15. Contour interval is two sigma.

of the direction of movement was estimated at 15 to 20 degrees. The direction of flow that would be estimated by measuring the direction normal to ice surface contours is toward 180±10°*. Thus the fabric of pebbles in the top 15 cm of till at this location does, in general terms, reflect the ice movement direction.

At location B (fig. 15) ice flow was measured at 0.1 m/yr toward the north. Since this is essentially stagnant the fabric direction (Plate III, b) probably represents ice flow at some time previous to 1960.

In location C only 25 pebbles in the upper 15 cm of till were measured (Plate III, c). Here the ice flow direction in 1960 was 3.5 m/yr from N 60° E. The ice-flow direction estimated from contours is N 30° E. Thus the direction of flow as calculated by Taylor might be in error or the fabric of the surface till may not record the very last motion of the ice.

The basal till fabric diagrams are similar in form to many of those reported by Lindsay (1970) from the Pagoda Formation in Antarctica, Drake (1968) for basal tills in New Hampshire, and others. They have one strong symmetrical mode plunging up-glacier and also are similar to an englacial fabric from Casement Glacier (Lindsay, 1970). They differ slightly from those in moraine ridges and ablation tills as discussed later in this chapter.

Till Deposition During the Change of Ice-Flow Direction

Although fabrics are related to ice-movement direction and the fabrics show a change in ice-flow direction parallel to that determined from early maps, it must be shown that till was actually being deposited while the flow direction change was taking place if a rate of till deposition is to be calculated. Tills reoriented by flowing ice have been reported by MacClintock and Dreimanis (1964) in the St. Lawrence Valley and they no doubt occur elsewhere.

In seven localities tills were sampled and fabric diagrams constructed one above the other in stratigraphic succession. The fabric diagrams for these locations are shown in Plate IV. Composition of samples related to these fabrics is given in Tables 4 and 5. In one section, on the north side of Nunatak A, three thin tills were differentiated on the basis of color (samples 122, 123, 124). Although changes in most parameters cannot be related to specific areas of bedrock upstream, significant changes in till composition do take place. There is not, however, a progressive change in composition that parallels

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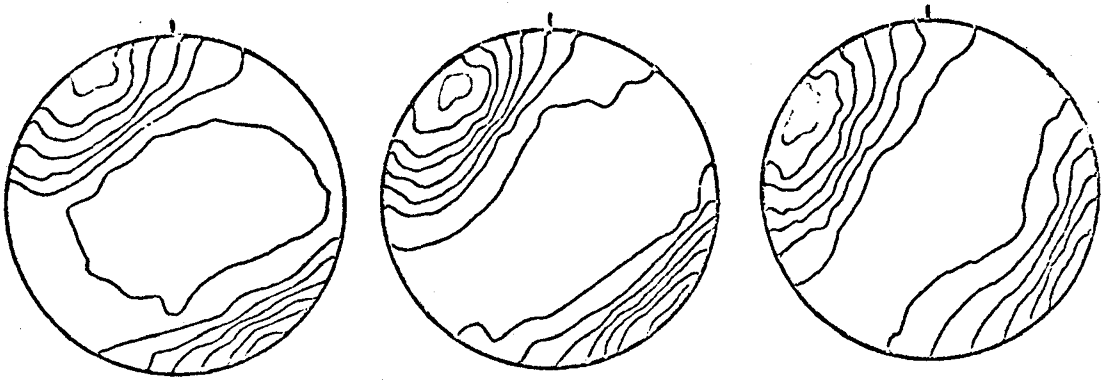
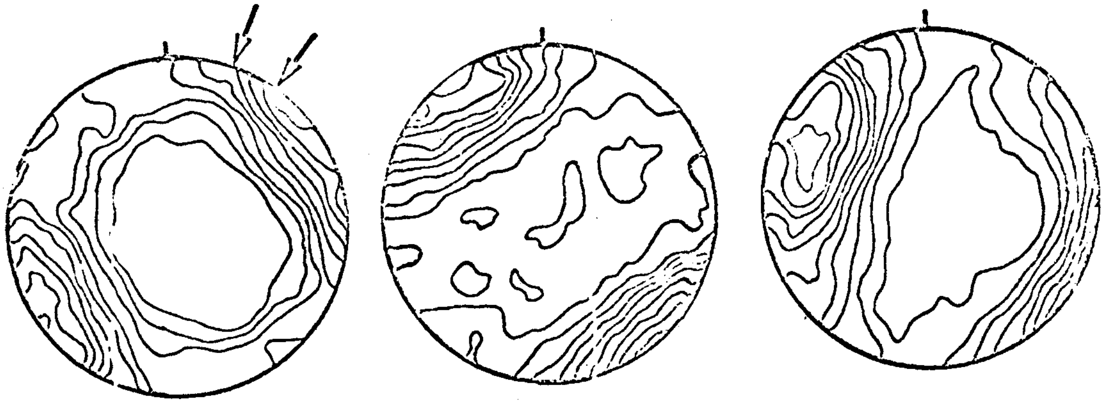
*Azimuths measured on 360° circle.



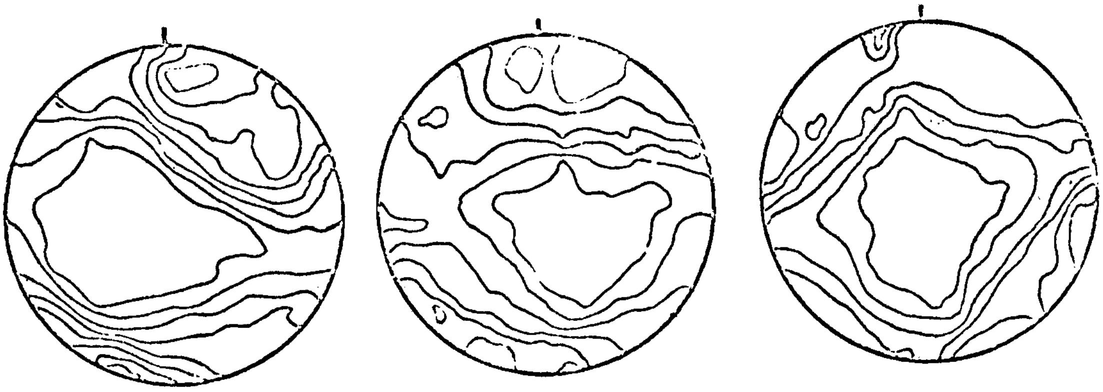
a. 0-25 cm b. 25-65 cm c. 65-80 cm

NORTH SIDE NUNATAK A (2σ)



d. 0-10 cm e. 10-30 cm f. 45-70 cm

WEST SIDE NUNATAK C (1σ)

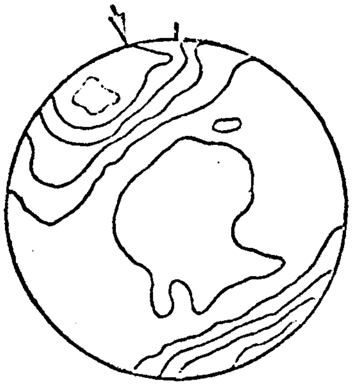


g. 5-10 cm (2σ) h. 10-20 cm (1σ) i. 25-40 cm (1σ)

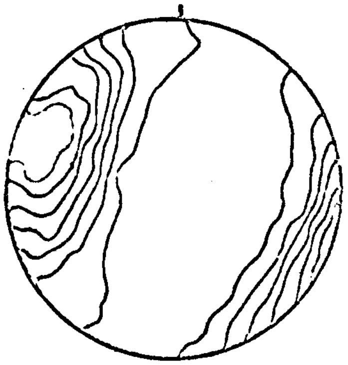
NORTH SIDE NUNATAK E

Plate IV. Fabric diagrams one above the other at same locations. North at top of diagrams. Arrows show orientation of crag and tail feature at surface. Contour interval and depth below surface is given. For location see Table 4.

WEST SIDE
NUNATAK B
(2σ)

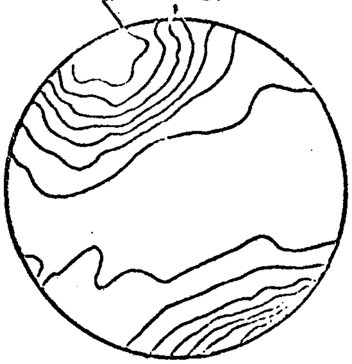


j. 0-35 cm

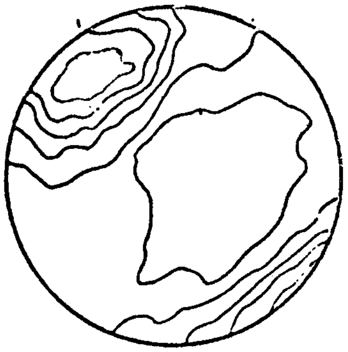


k. 50-65 cm

WEST SIDE
NUNATAK D
(2σ)

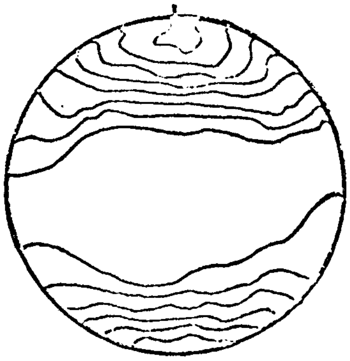


l. 0-40 cm

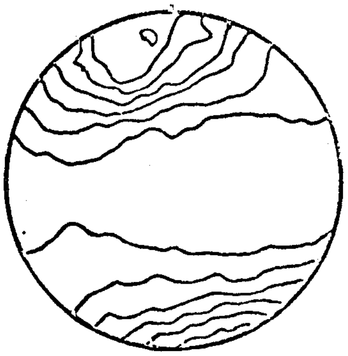


m. 40-60 cm

EAST SIDE
NUNATAK D
(2σ)

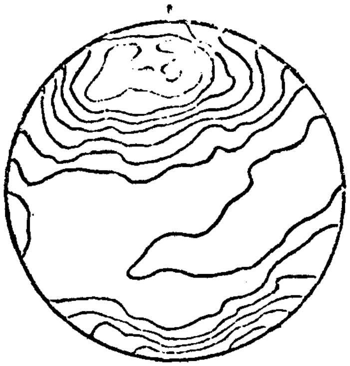


n. 0-10 cm

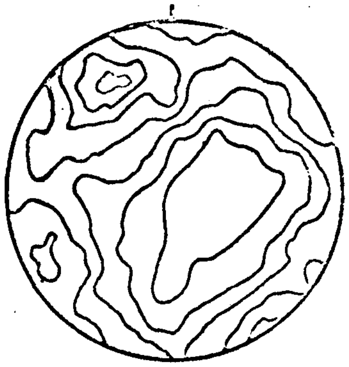


o. 10-25 cm

BETWEEN
NUNATAKS D
AND E (1σ)



p. 0-10 cm



q. 10-100 cm

Plate IV. Continued.

Table 4. Composition and fabric data from seven sections of till in the Glacier Bay Formation. See Figure 15 for exact location.

Location	Sample Number	Depth of Till (cm)	% of Total Sample < 2 mm	< 2 mm Fraction % Sand	% Silt	% Clay	% Metasiltstone Rock Fragments	% Calcite	% Dolomite	% Total Carbonate	Fabric Reference (Plate, Figure)	Depth of Fabric (cm)	Mode of Azimuths (Degrees)	Vector Mean (Azimuth)	Standard Deviation of Vector Mean
North Side Nunatak A	124	0-25	83	46	44	9	33	1.6	0.6	2.2	IV,a	0-25	330	330	5.1
	123	25-65	70	40	45	15	46	1.1	1.2	2.3	IV,b	25-65	321	319	3.9
	122	65-80	83	40	50	10	41	0.8	0.2	1.0	IV,c	65-80	301	302	4.4
North Side Nunatak E	140	0-20	67	57	35	8	25	1.6	0.5	2.1	IV,g	5-10	9	18	8.1
											IV,h	10-20	9	15	8.9
	141	20- ?	28	48	43	9	24	1.4	0.4	1.8	IV,i	25-40	319	315	10.4
West Side Nunatak C	108	0-30	61	55	39	7	20	0.9	0.3	1.2	IV,d	0-10	45	240	6.7
											IV,e	10-30	320	326	6.4
	107	30- ?	87	52	42	6	13	0.9	0.3	1.2	IV,f	45-70	290	289	5.6
West Side Nunatak B	46	0-35	88	41	51	9	40	2.0	0.3	2.3	IV,j	0-35	322	329	6.0
	48	35- ?	74	46	45	9	38	2.0	0.8	2.8	IV,k	50-65	292	293	4.2
East Side Nunatak D	144	0-15	70	53	43	5	29	0.9	0.3	1.2	IV,n	0-10	7	3	4.4
	143	15-40	68	60	34	6	44	2.3	0.8	3.1	IV,o	10-25	350	350	4.6
West Side Nunatak D	50	0-40	61	57	37	6	33	1.8	0.8	2.6	IV,l	0-40	335	341	4.3
	51	40- ?	74	56	39	5	22	2.6	0.3	2.9	IV,m	40-60	320	324	5.4
Between Nunataks D and E	101	0-10	58	63	31	6	-	5.5	0.8	6.3	IV,p	0-10	0	350	6.0
	102	10-80	40	73	23	5	-	7.5	1.4	8.9	IV,q	10-100	330	328	7.6

Table 5. Relative clay mineral abundances in three sections of Glacier Bay till. For depth and relation to fabric diagrams see Table 4.

Location	Sample Number	% Illite	% Vermiculite	% Quartz	% Chlorite	% Expandable
North Side Nunatak A	124	T	40	25	35	T
	123	T	T	10	65	20
	122	10	35	15	35	T
West Side Nunatak B	46	25	10	20	35	10
	48	T	T	40	60	T
West Side Nunatak D	50	15	15	30	35	5
	51	10	T	5	75	10

the change in fabric direction suggesting that bedrock contacts beneath the ice and up-ice from this location are more complex than is shown in Figure 2 or the change in till composition occurs suddenly after crossing a bedrock contact.

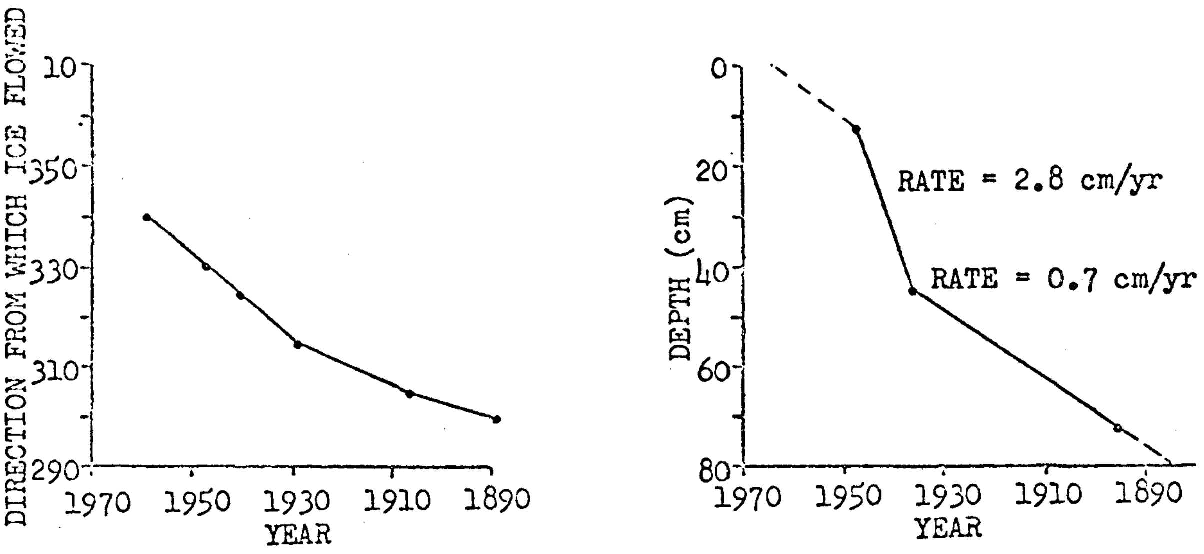
On the east side of Nunatak D (samples 143, 144) a 14 percent decrease in metasiltstone fragments occurs in the upper till. This may reflect a more eastward origin (over Nunatak A which is composed of diorite) of the upper till than the lower till, which would have been carried over metasiltstone after leaving the Bruce Hills. Till west of Nunatak C (samples 107, 108) shows a decrease in metasiltstone fragments with depth. This might be expected because the ice that deposited the lower till presumably flowed over less metasiltstone than the ice that deposited the upper till. Changes in the amount of chlorite also seem significant but this has not been related to bedrock. Although the changes in compositional parameters are not progressive with depth in the till, they must indicate that till was being actively carried and deposited while the change in ice flow direction was taking place.

Rate and Mode of Till Deposition

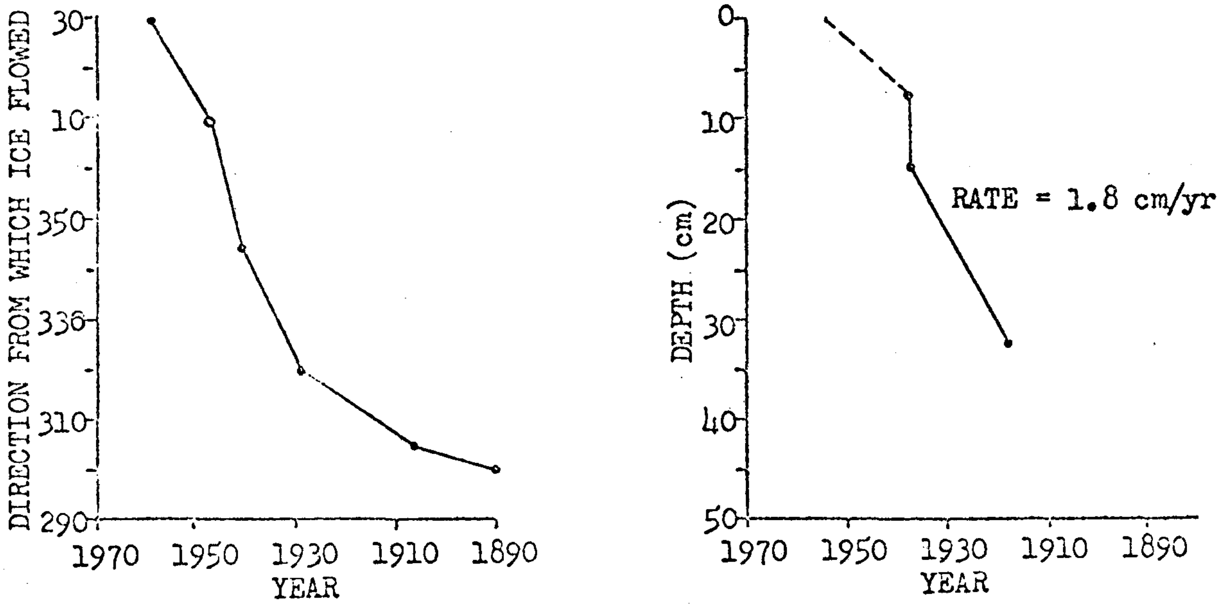
If it is accepted that the till sequences listed in Table 4 and 5 do represent real changes in till provenance accompanying the change in ice flow direction, then rates of basal till deposition can be estimated. Ice flow directions for 1960, 1948, 1941, 1907 and 1892 were estimated by measuring the direction normal to the ice surface contour over each location on published maps. Oblique aerial photographs were used to estimate the direction of ice-surface slope in 1929. These values were then plotted against time for each location (plate V). The directions are estimated to be correct to within ±10°.

Since fabrics were done on 5- to 40-cm-thick till units, the fabric direction (mode) is taken to be the fabric (thus the ice-flow direction) at the midpoint of each till unit. Since the ice-movement direction for any year is known, the "year" of deposition can be plotted against depth for each location (plate V). The slopes of straight lines through these points yield rates of deposition ranging from 0.5 to 2.8 cm/yr. The number of points is probably not sufficient to conclusively demonstrate changes in rate during this short time span but the curve suggests that till deposition began quite late in Neoglacial time and had ceased or slowed down when the ice was about 70-100 m thick. Since the tills can be differentiated as units, deposition of the till may not have been continuous.

The significance of the rate of deposition is not readily apparent. Basal melting due to geothermal heat is sufficient to melt 0.5 cm of ice a year. Heat added by friction is probably negligible at flow rates measured in 1960 (<5 m/yr in this area) but if higher flow rates are assumed to have been prevalent in the past, the total basal melting of

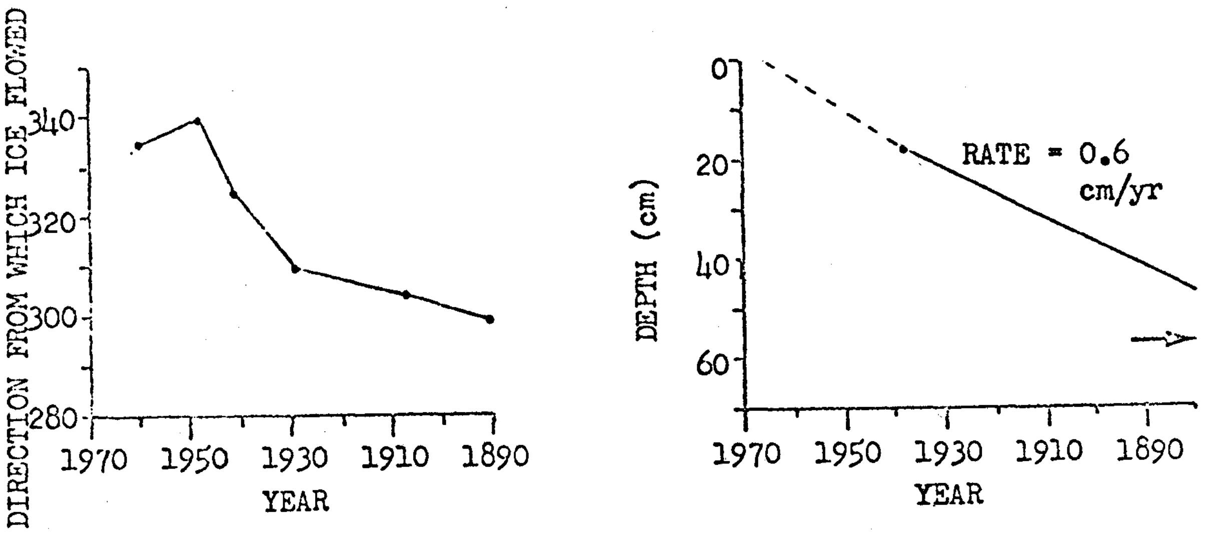


a. North Side Nunatak A

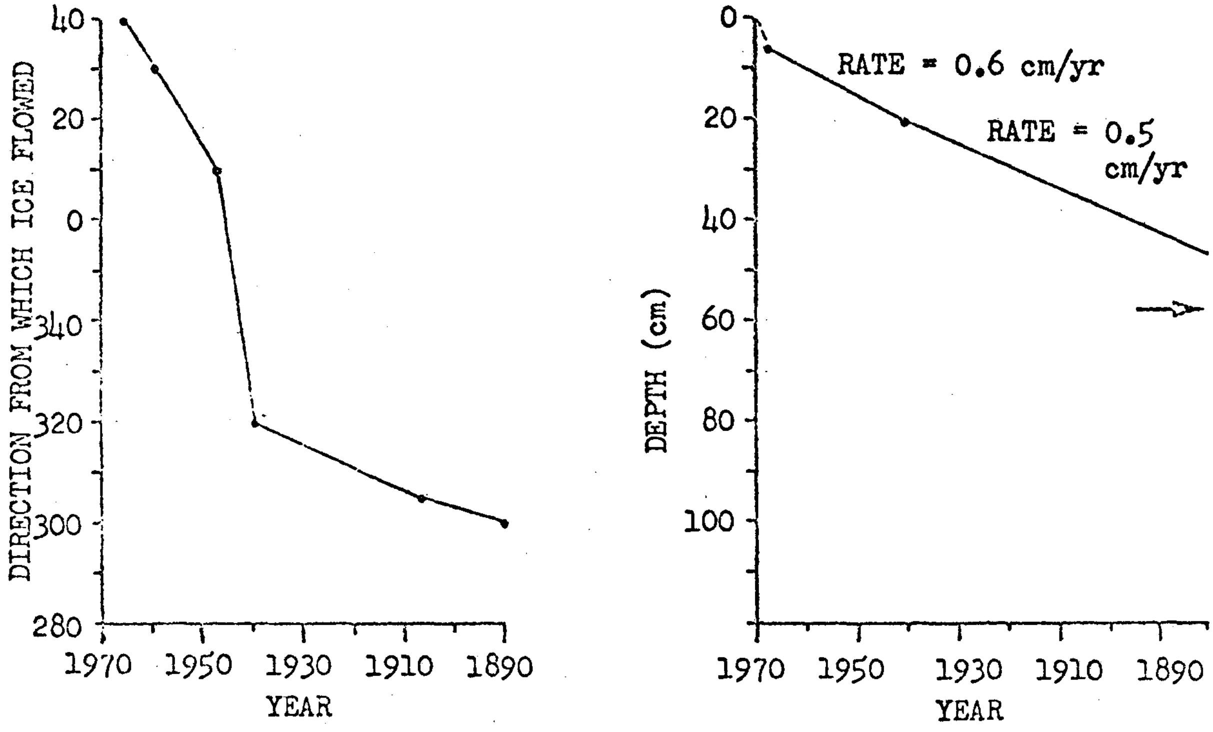


b. North Side Nunatak E

Plate V. Graphs showing change in ice-flow directions with time and thickness of till deposited between known times at seven locations.

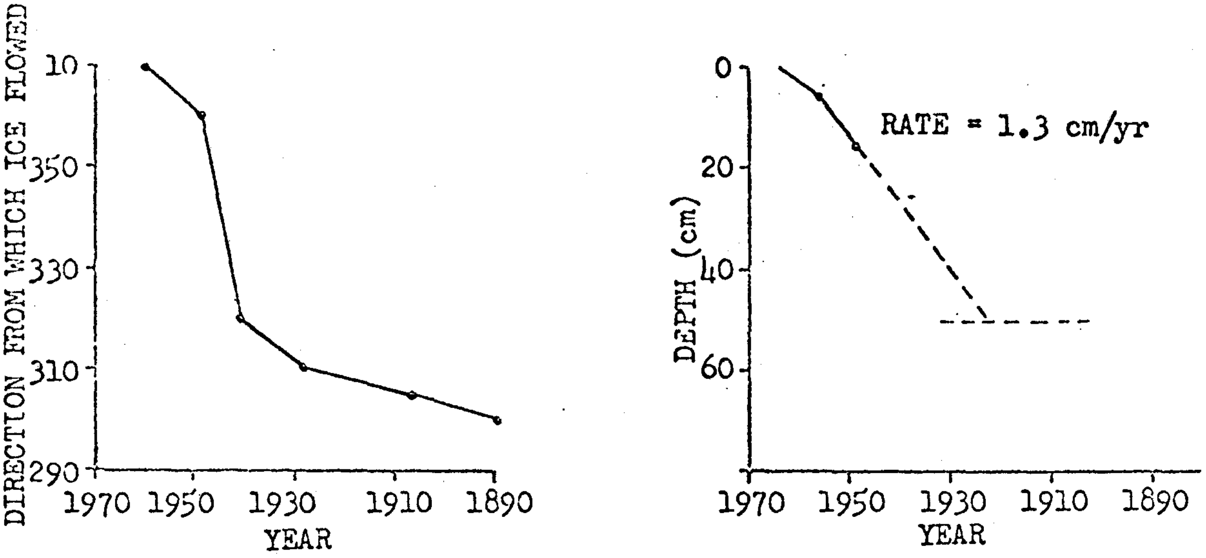


c. West Side Nunatak B

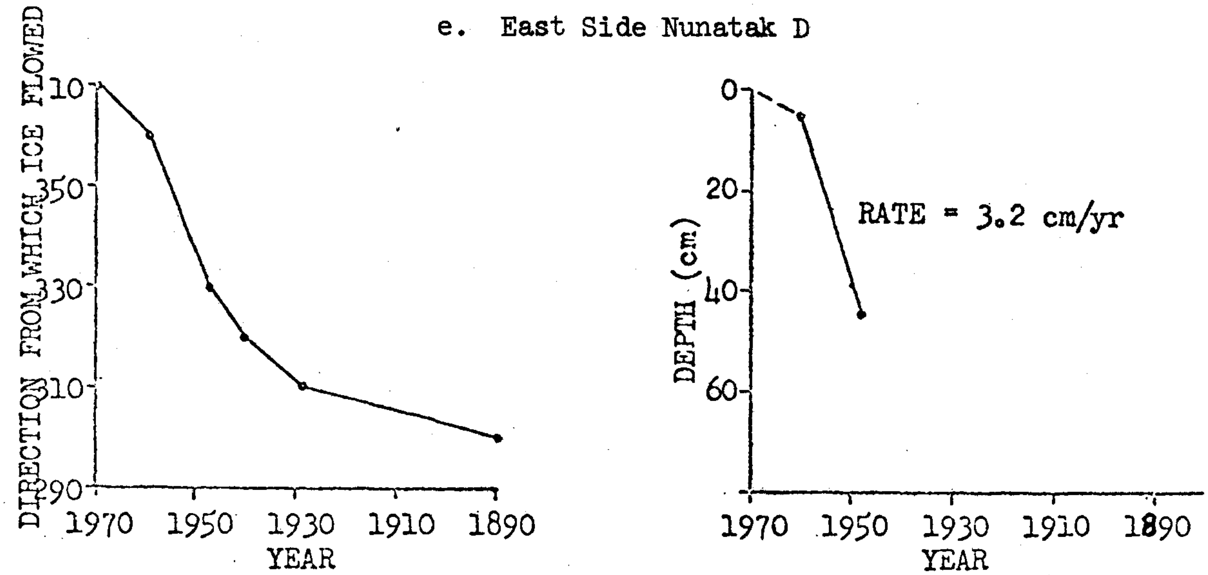


d. West Side Nunatak C

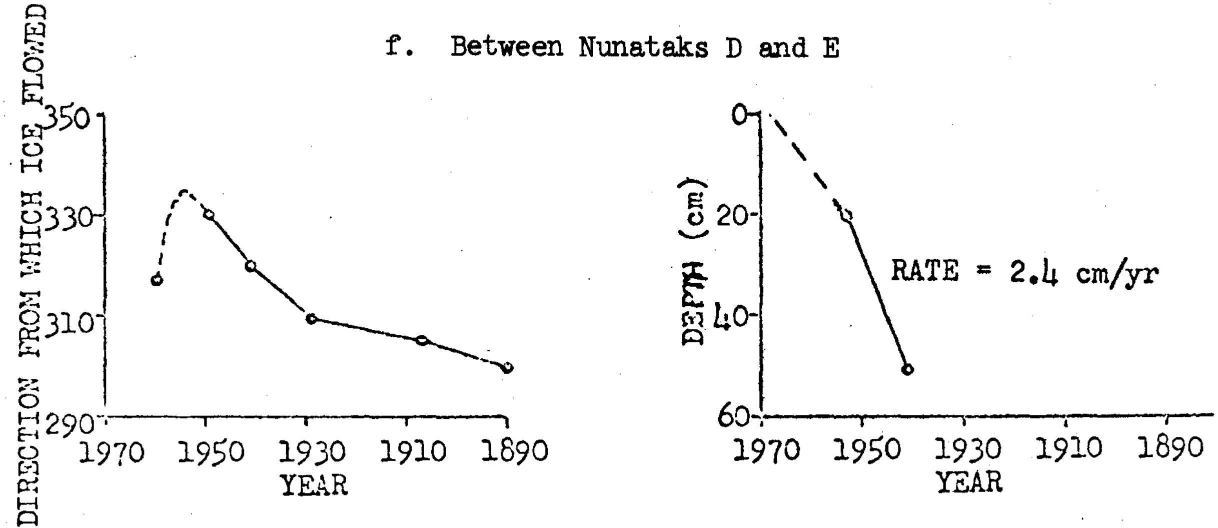
Plate V. Continued.



e. East Side Nunatak D



f. Between Nunataks D and E



g. West Side Nunatak D

Plate V. Continued.

ice would be on the order of 1 cm/yr. If the ice-till mixture at the base of the ice was as little as 25 percent ice, then 2 to 4 cm of ice-till mixture or 1.5 to 3 cm of till would be expected to have melted out each year. Thus basal melting might account for the till deposition.

Depending on the amount of water present in the till at the time of deposition, some reorientation of the till which had melted out might be expected. This does not seem to have happened except west of Nunatak C and on Nunatak E (plate IV).

A second possibility is that the calculated rate of till deposition represents the stagnation of a basal till-ice mixture under flowing ice as suggested by Boulton (1970). In this case a thin, stagnant ice-till mass would separate the overlying active ice from the till-water mixture being formed by basal melting, preventing reorientation of fabric.

Although till was observed in only one location under the Burroughs Glacier (fig. 20) it tends to support the above hypothesis. One meter of frozen till (estimated 25-40 percent ice) was examined between Nunataks D and E under 5 m of relatively clear ice. Fabric diagrams (plate IV, p and q) indicate that the top 20 cm of till which is slightly lighter gray does show a different flow direction than the till below. Thus the lower till is considered to be in place in determining the rate of deposition even though it is still frozen. If the frozen till is 25 percent ice, then the actual rate of deposition calculated after melting would be 2.3 cm/yr (plate V, f).

Ice thickness and ice slope in 1892, 1941, 1948, and 1960 can be determined over Nunataks A and D. The basal shear stress (τ) can be estimated by the equation

$$\tau = \rho g h \sin\alpha$$

where ρ is the ice density, g is acceleration due to gravity, h is ice thickness, and α is the surface slope of the ice. Basal shear stress at points on the stoss sides, lee sides, and tops of Nunataks A and D range from 1.4 to zero bars (table 6). Values for 1892 are nearly the same for both stoss and lee sides of the nunataks. This suggests that difference in the time of till deposition is not directly related to difference in basal shear stress. Because of a decrease in ice-surface slope accompanying ice-surface lowering, values on the sides of both nunataks steadily decrease through time. The apparent increase in shear stress between 1941 and 1948 on the stoss sides of the nunataks may be a function of the estimation of 1941 ice-surface slopes from the map with a 250-ft contour interval.

If this stagnation of frozen till took place earlier or at a greater rate on the stoss side of bedrock knobs as described by Boulton (1970), then it might explain the apparently thick Neoglacial till in these positions at both ends of the Burroughs Glacier (fig. 21). At the southeast end, these till deposits indicate flow from the west-northwest. A similar flow direction is indicated by the till buildup

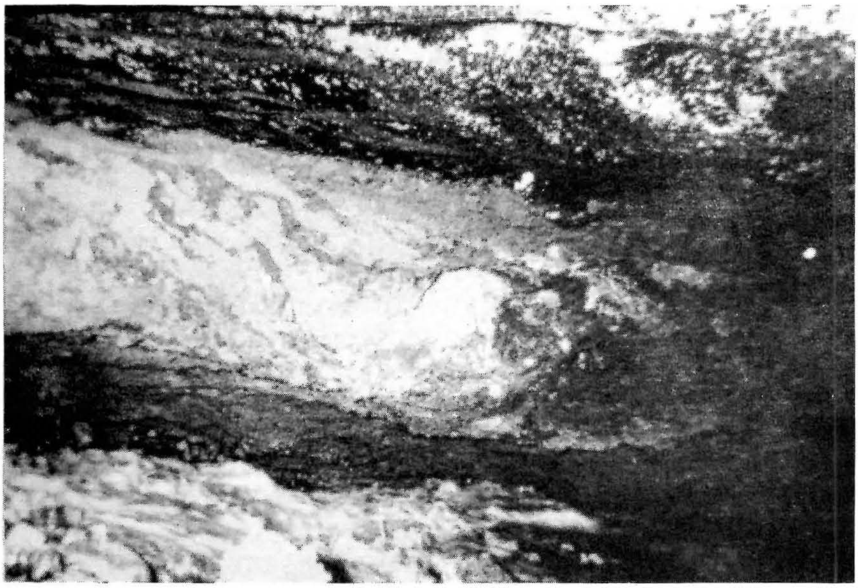


Figure 20. Photo of 1 m of frozen till beneath ice. Note meltwater stream at base. (Photo by G. Larson)

Figure 21. Photo from Nunatak A showing shape of hills with till deposits built up on stoss side.

Table 6. Calculated basal shear stress at six locations.

Location	Year	Basal Shear Stress (bars) Stoss Side	 Nunatak Top	 Lee Side
Nunatak A	1892	1.3	1.1	1.4
	1941	0.8	0.6	0.9
	1948	1.1	0.6	0.2
	1960	0.7	-	0.1
Nunatak D	1892	1.4	1.1	1.4
	1941	1.2	0.4	1.0
	1948	1.3	0.2	0.4
	1960	0.6	-	0.0

on one hill at the terminus of the Cushing Glacier. However, at the western terminus of the Burroughs Glacier, in the same valley only 1 km southeast of Cushing Glacier, the till deposits suggest flow from the east. Thus, most till deposition must have taken place after the Burroughs reversed its flow direction (shortly before 1892).

The following history of till deposition is proposed for the Burroughs Glacier:

1. Erosion and non-final deposition through most of Neoglacial time.
2. Progressive stagnation of frozen till (as described by Boulton, 1970) on the stoss sides of bedrock knobs sometime after the western half of Burroughs Glacier began to flow toward Queen Inlet.
3. Continued deposition in these areas and deposition beginning in areas not directly up-ice of nunataks as late as 1940. Basal melting of the till-ice mixture and deposition is consistent with the hypothesis of Harrison (1957) and, in general, with that proposed for basal melt-out tills by Boulton (in press).

Surface Features

Moraine Ridges

Small ridges, usually composed of till and trending roughly parallel to a retreating ice margin, have been discussed by numerous writers using differing terminology (e.g., annual moraines - DeGeer, 1897; Washboard moraines - Mawdsley, 1936; sub-lacustrine moraines - Goldthwait, 1951; DeGeer moraines - Hoppe, 1957, 1959; cross-valley moraines - Andrews, 1963a, b; Andrews and Smithson, 1966). The till ridges that occur near Burroughs Glacier differ only slightly from all of these.

The origin of the till ridges has been debated and, in fact, several origins may be correct. Hoppe (1952) suggested that wet till might be squeezed into basal crevasses and in later discussions this mode of formation was extended to squeezing at the margin as well. Andrews (1963a, b) and Andrews and Smithson (1966) have also accepted this mechanism for the formation of cross-valley moraines on Baffin Island. They suggest that an influx of meltwater might make the till very mobile. Haselton (1967) observed till evidently squeezed into fractures in the ice at the margin of Plateau Glacier. Others (e.g., Price, 1970) have suggested that ridges form at the ice front by squeezing of till from beneath the margin. Another possibility is that the ridges are composed of material dropped from above into a crevasse.

The ridges (fig. 22), 0.5 - 2.5 m high, occur in a band approximately 1 km wide (fig. 23) along the north and south shores of Wachusett Inlet and are found at least as high as 70 m above sea level. Most ridges are not continuous for more than 25-75 m and are straight or arcuate with their concave sides downglacier. They are usually asymmetrical in cross section, the stoss side usually being more gentle (15°-30°) than the lee (> 25°). Several recently uncovered ridges had slumped distal sides. No ridges were noted on or below the present beach although soundings might demonstrate their presence in Wachusett Inlet.

To demonstrate the squeezing hypothesis, it is necessary to show that the till in the ridges is in fact basal till and that crevasses were present in the correct orientation. The distribution of ridges along the northern side of Wachusett Inlet is shown in Figure 23, and Figure 24 is an aerial photograph of the same area in 1948. Note the similarity in alignment of the till ridges and the crevasses that existed in 1948. In these air photos and ground photos that were taken in 1950 the ice appears to be clean. It is doubtful that enough ablation till was at the surface of the ice to create the numerous ridges now present. This is supported by observation of crevasses at the present terminus of Plateau Glacier where, although the ice is thin, very little ablation debris is present.

Samples from three till ridges and the till 20 cm beneath them were analyzed (table 7). Although one ridge shows some washing, the analyses are nearly the same for samples from and below each till ridge (compare with the range of percentages in Glacier Bay till, Appendices A, B, C, D). A large enough sample is not available for statistical analysis but these data do suggest that the till in the ridges is the same composition as the tills beneath them. In one locality, a till ridge overlies the lower member of the Van Horn Formation. Stringers and pods of clean sand (fig. 25) are present both in the ground moraine and in the ridge, suggesting that the ridge is composed of squeezed basal till.

In general, fabrics in the till ridges show more dispersion about the mean azimuth than other fabrics in the Glacier Bay till. Fabric diagrams were made for pebbles on the stoss and lee flanks and beneath five ridges (plates VI, VII). The fabric modes trend nearly normal, oblique, and nearly parallel to the ridge axis and show fairly good agreement with the late ice-flow direction determined from nearby crag and tail features, grooved till, and from ice surface contours on 1948 and 1960 maps. In one case (plate VI, d) the fabric on the lee side is much less distinct than that on the stoss side, perhaps because of slumping. Andrews and Smithson (1966) also report weaker fabrics on the steeper stoss slopes.

The relationship between fabric and ridge orientation does not agree with that determined by Hoppe (1957), Andrews and Smithson (1966)

Figure 22. Photo of till ridges north of Wachusett Inlet. Note man in foreground of ridge on left. Distance between ridges is about 30 m.

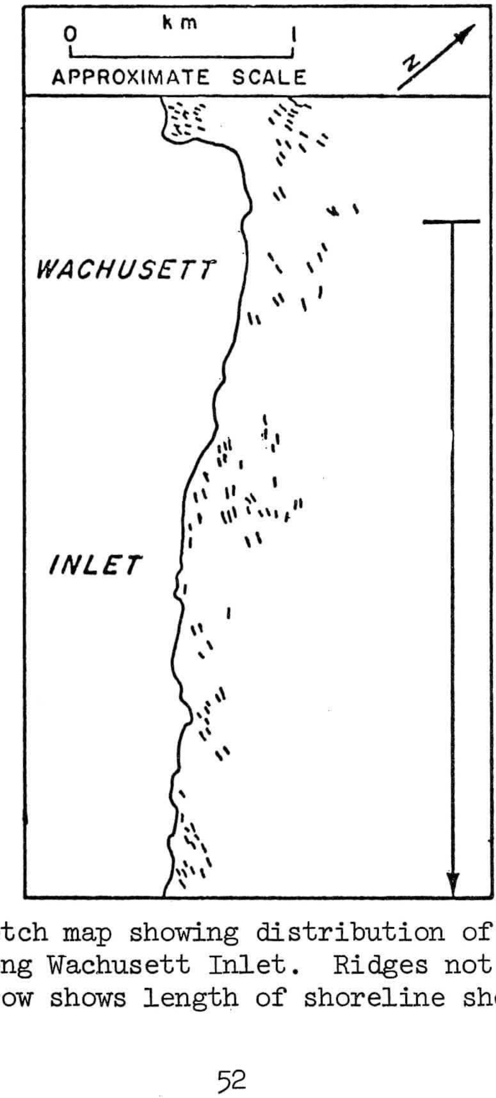


Figure 23. Sketch map showing distribution of till ridges along Wachusett Inlet. Ridges not drawn to scale. Arrow shows length of shoreline shown in Figure 24.

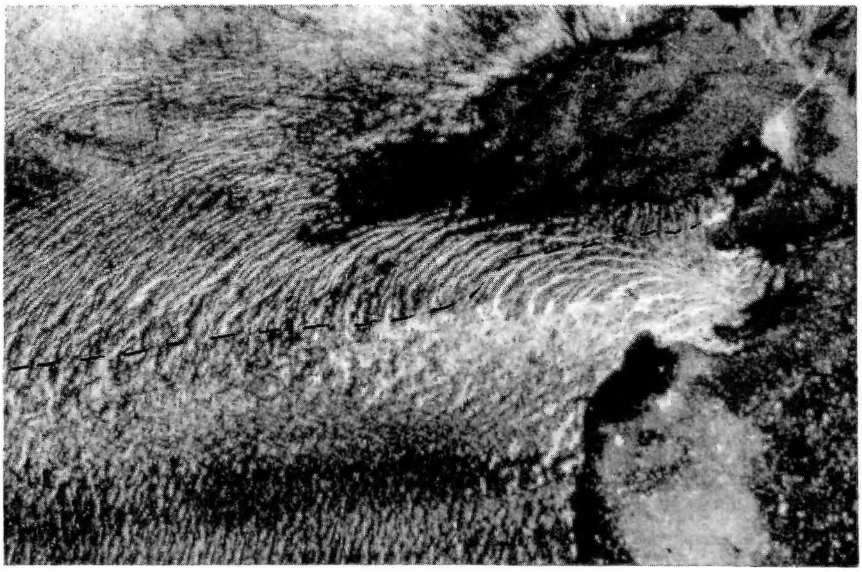
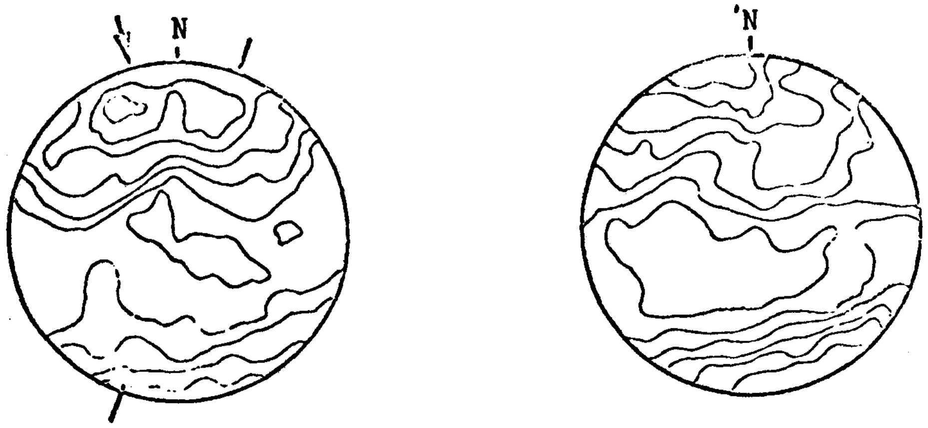


Figure 24. Aerial photo (1948) of crevasse distribution in relation to edge of Wachusett Inlet (---). For approximate scale see Figure 23.

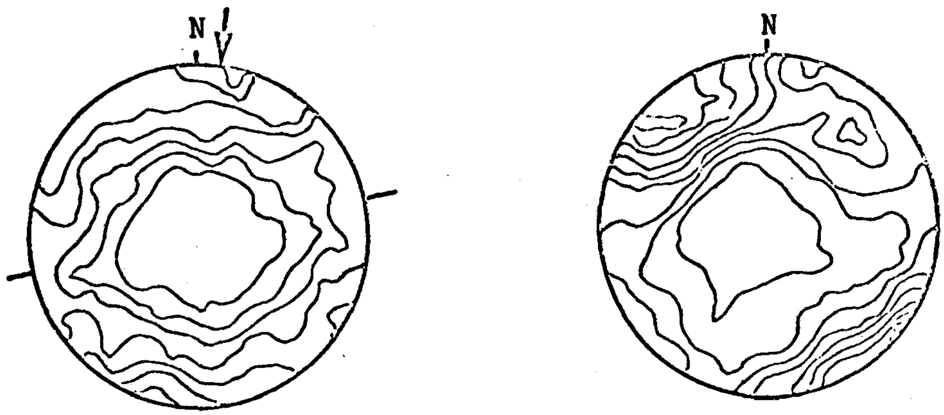
Table 7. Composition of samples collected in and below till ridges.

				< 2 mm Fraction											
Location	Sample Number	Fabric Reference (Plate, Figure)	% of Total Sample < 2 mm	% Sand	% Silt	% Clay	% Metasiltstone Fragments	% Illite	% Vermiculite	% Quartz	% Chlorite	% Expandable	% Calcite	% Dolomite	% Total Carbonate
In Till Ridge	40	VII,a	65	58	34	9	25	25	T	10	55	5	2.6	1.7	4.3
Below Till Ridge	39	VII, b	65	65	27	9	28	20	15	10	50	T	2.4	1.1	3.5
In Till Ridge	41	VII,c,d	72	62	30	8	29	15	10	15	55	5	2.7	0.6	3.3
Below Till Ridge	42	VII,e	42	63	29	8	31	25	5	5	60	5	2.6	0.8	3.4
In Till Ridge	62	VIII,b	63	81	16	3	26	T	20	45	35	T	1.5	0.7	2.2
Below Till Ridge	63	-	76	63	34	3	25	T	5	50	40	5	1.6	0.2	1.8

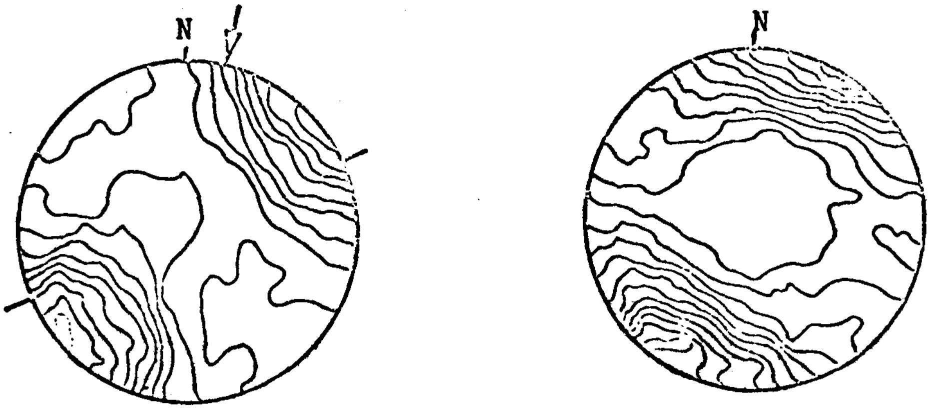
Figure 25. Cross section of small till ridge over sand. Note sand pods and stringers in till.



a. Stoss Side b. Lee Side

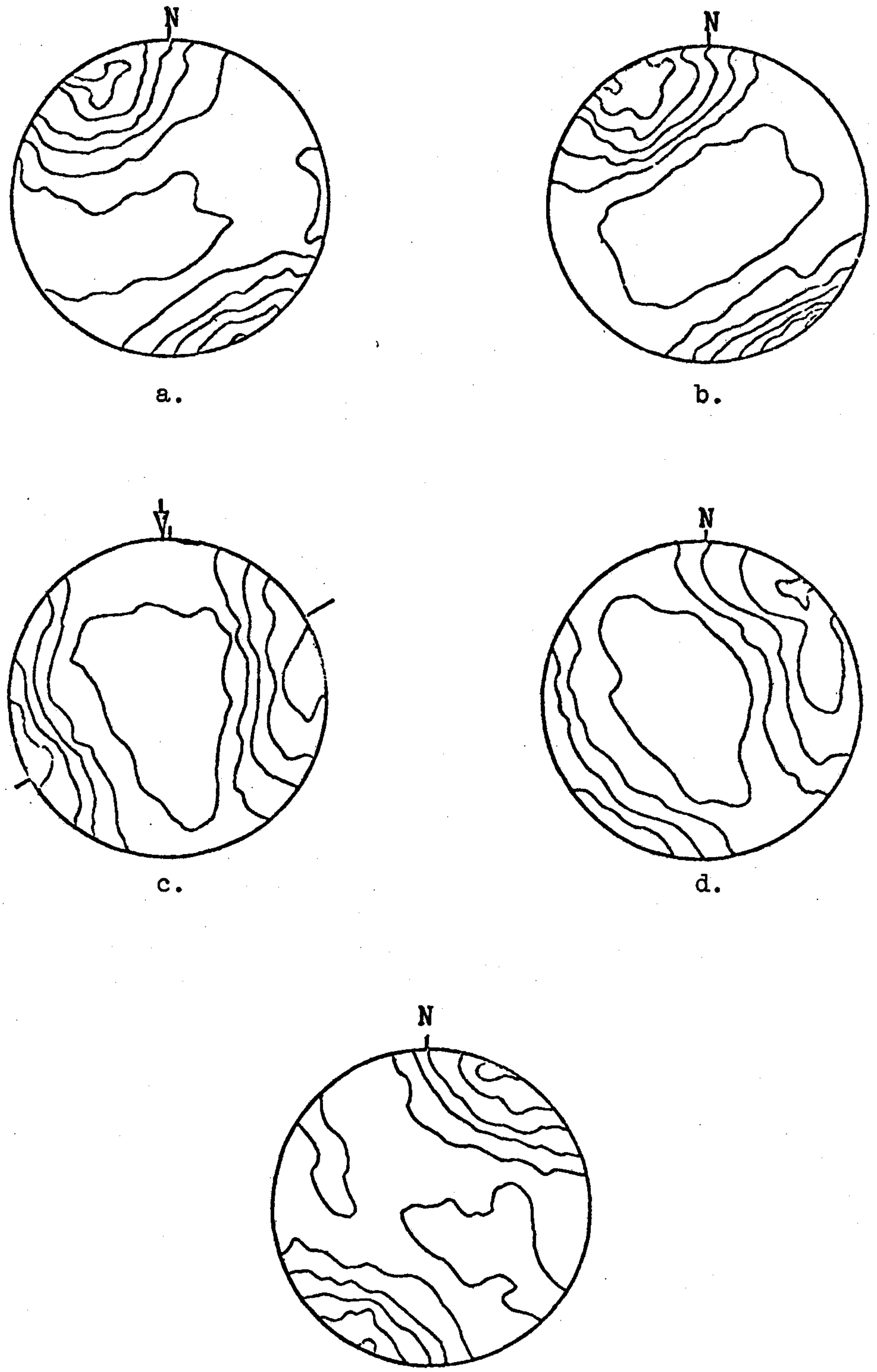


c. Stoss Side d. Lee Side



e. Stoss Side f. Lee Side

Plate VI. Fabric diagrams on opposite flanks of three till ridges. Bar denotes orientation of ridge axis, arrow is late ice flow direction (crag and tail orientation). All have one sigma contour interval.



a.
b.
c.
d.
e.

Plate VII. Fabric diagram from within ridge (a) and 20 cm beneath it (b) and proximal (c) and distal (d) sides and below (e). Arrow is late ice-flow direction. Bar is ridge orientation. All have contour interval of two sigma.

or Price (1970). They all report fabrics normal to the ridge axis independent of ridge orientation. These differences are not irreconcilable, however. As the ice edge retreated up Wachusett Inlet the last ice left on either side moved very slowly toward the inlet from the Burroughs Glacier on the north and from a stranded ice mass on the south side of the inlet. This was the last ice movement and the fabrics may have been reoriented after squeezing.

One ridge, located at the south edge of Plateau Remnant, was observed partly in ice that was about 2 m thick (fig. 26). The frozen till was confined to a 35-cm-thick layer dipping 55-70° up-ice. The fabric (plate VIII, b) is oblique to the ridge axis and the mean azimuth plunges considerably less than the dip of the till layer. Approximately 50 m away from this site, a fabric was measured in the same ridge off the ice (plate VIII, a) and it has a fabric mode nearly normal to the ridge axis and also nearly parallel to several nearby crag and tail features. The steep dip of the till unit in the ice and its thickness suggest that it is a till-filled crevasse rather than a shear plane. Since the crevasses forming along the edge of the inlet would be expected to be vertical some rotation of the till unit due to ice flow may have taken place.

Another till ridge still in ice (fig. 27) shows till evidently squeezed upward 1 m high into a nearly vertical crevasse close to the present terminus of Plateau Glacier. It has a nearly horizontal fabric mode (plate VIII, d) that is nearly normal to the crevasse and has a wide dispersion. The crevasse is closed above the till. A fabric in a ridge, nearly parallel to and 20 m from this ridge, but off the ice shows nearly the same orientation. Since the margin of Plateau Glacier did not retreat past this location until 1966 or 1967, there has been very little time for reorientation of pebbles by ice flowing from the Plateau Remnant directly toward the inlet. The horizontal component of any flow that has taken place, however, was probably perpendicular to the ridge axis.

Any hypothesis for the formation of these ridges must take into account the following observations:

1. Till ridges are abundant in the area that was covered by crevassed ice as the Plateau Glacier receded and their orientation is parallel to the orientations of the crevasses.

2. The till ridges were formed above sea level.

3. Before deposition some till ridges occur at the base of nearly vertical closed fractures in the ice which have the same orientation as the crevasses of less than a decade ago.

Figure 26. Photo of till ridge in ice on Plateau Remnant.

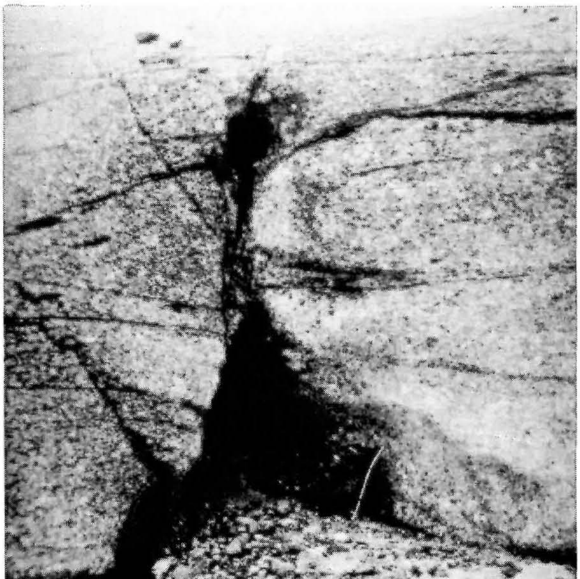
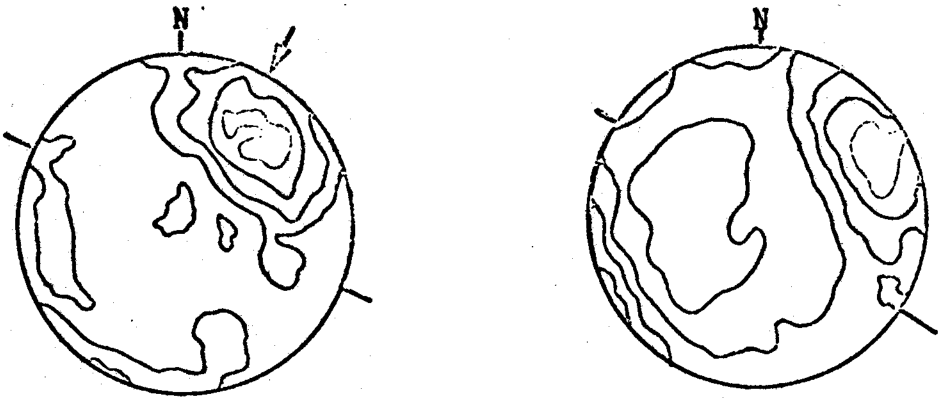
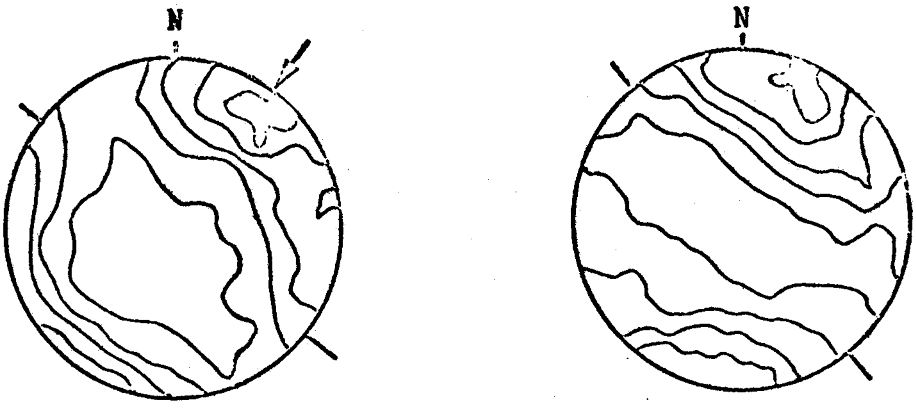


Figure 27. Photo of till in crevasse at south margin of Plateau Remnant. Circular tube above is empty. Note pack on right of crevasse.



a. In Ridge Near Edge of Ice

b. In Same Ridge as a. in Ice



c. In Ridge Near Edge of Ice

d. In Ridge in Ice Near c.

Plate VIII. Fabric diagrams from till in closed crevasse (b,d) and in ridge off ice (a,c). Arrow is late ice-flow direction. Bar is crevasse or ridge orientation. Contour interval is two sigma.

4. The till in the crevasses is frozen.

5. The ridges have gentle stoss sides and steeper lee sides.

6. Fabrics in ridges affected by late ice movement reflect this late ice movement; those in areas close to the present Plateau Glacier may not show reorientation and are normal to the ridge axis.

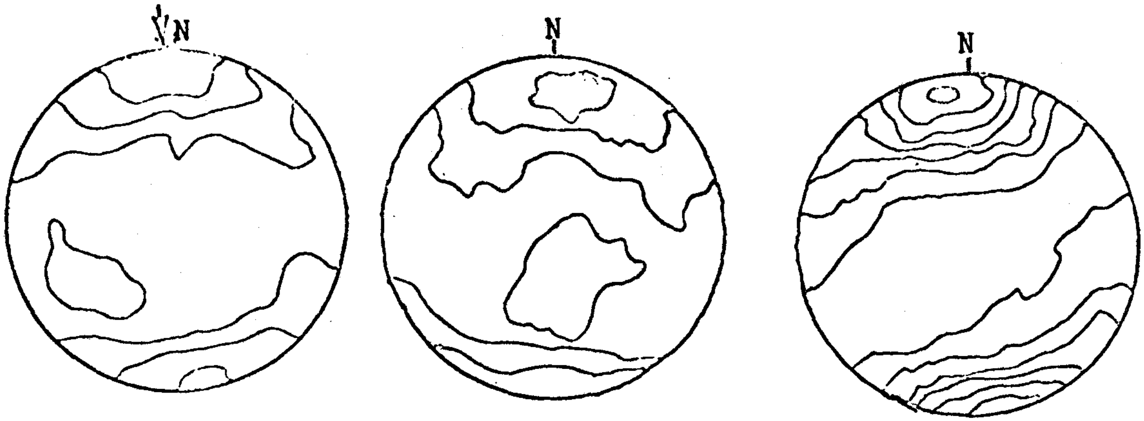
Therefore I propose that wet, basal till was squeezed into the base of crevasses that formed in the late stages of deglaciation. Subsequent ice flow reoriented particles within the till that had been squeezed up. This flow may also have been responsible for the rotation of crevasses and the formation of the more gentle stoss slopes on the ridges. The till that was observed to be frozen during the summers of 1969 and 1970 may have become frozen during the previous winter.

Crag and Tail Features

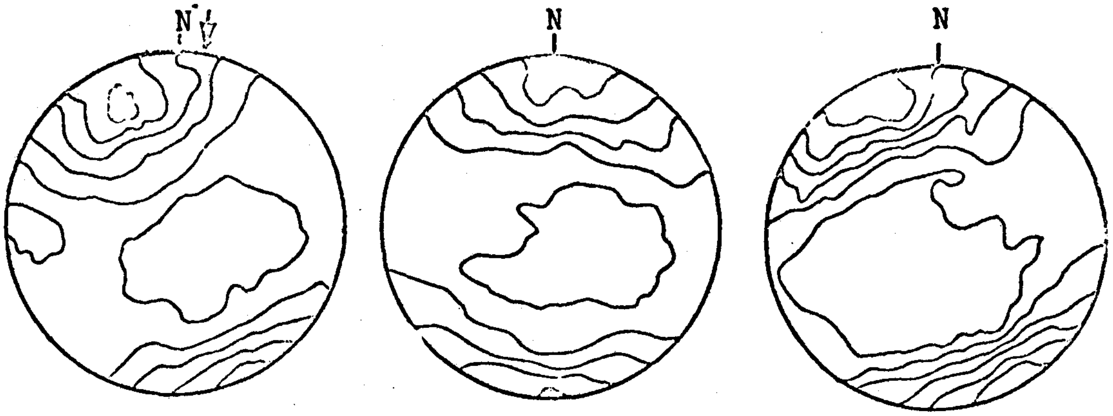
Although small crag and tail features composed of till are not commonly found in areas deglaciated during the late Pleistocene, and thus cannot be used, they seem to be excellent indicators of late ice-flow directions at the Burroughs Glacier. The orientations of some were measured (fig. 19). They consist of till ridges as much as 3 m long and 0.5 m high that extend from the lee side of boulders. They are presumably composed of basal till that filled cavities at the base of the ice similar to those described by Peterson (1970) and others.

Fabrics were measured on the east and west sides and beneath two crag and tail features (plate IX). In one case the fabric modes (plate IX, a, b, c) on both sides of the ridge parallel the axis of the feature and show the change in ice flow direction recorded by maps and photographs. In the other case (plate IX, d, e, f), the east flank shows a more westerly fabric indicating the pebbles plunging upglacier and toward the ridge axis.

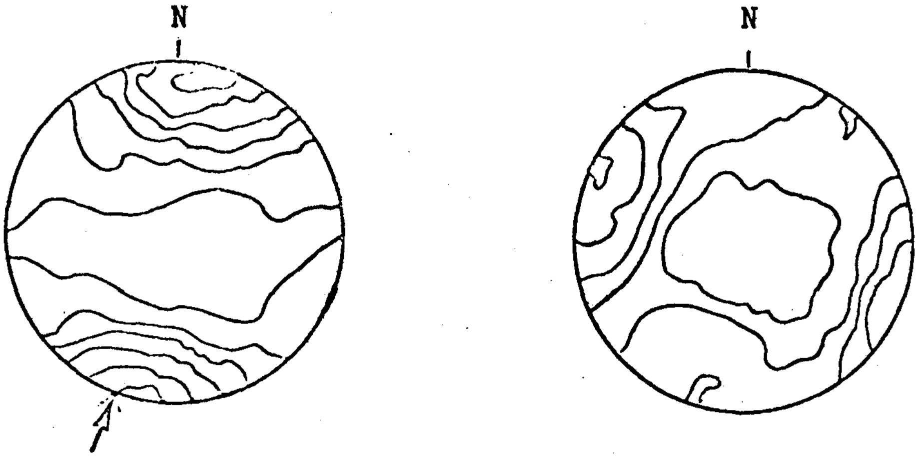
In all cases where fabric diagrams were made below crag and tail features the change in ice-flow direction is consistent with known changes. An ice mass was stranded along the south side of Wachusett Inlet just before 1948 (fig. 7) as the Plateau Glacier margin retreated. Although the ice was probably not more than 30 to 40 m thick, crag and tail features show ice movement directly downslope (northward) toward the inlet. The fabric mode (plate IX, g) is nearly horizontal and parallel with the crag and tail axis. The surface till (top 20 cm) on the hillside (plate IX, h) shows ice flowing from the west-northwest (down Wachusett Inlet) but it does have a slight concentration of pebbles with long axes in the north-south direction suggesting that some reorientation of the pebbles took place in the 1940-1950 decade.



a. West Flank b. East Flank c. Beneath



d. West Flank e. East Flank f. Beneath



g. In Crag and Tail h. Till Nearby at Surface

Plate IX. Fabric diagrams from and beneath crag and tail features at three localities. Arrow is orientation of crag and tail feature. Contour interval is two sigma.

Ablation Till

There is very little ablation till on the Burroughs Glacier today. A debris layer as much as 0.5 m thick occurs around the margin of the ice north and east of Nunatak E. This is evidently derived by shear planes carrying material toward the ice surface. These planes dip at about 30°, and the frozen till layers in them are up to 5 cm thick. Details of their distribution and attitude are given by Larson (in preparation). Ablation has created a blanket of till as much as 1 m thick over the lower part of the ice, thus locally reducing ablation, and causing slumping and mudflows as described by Sharp (1949). Progressive development of the area over a one year period is shown in Figures 28 and 29. The mean percentages of sand, silt, and clay for seven samples of ablation till (Appendix A) indicates some loss of fines by washing.

Till fabrics were measured in four locations where ablation till had been deposited stratigraphically above basal till (plate X). In one case (plate X, c) the ablation till was in a ridge approximately 0.5 m high which had evidently formed by flow or slump of till into a fracture between ice blocks. The fabric is well defined and the mode is oriented at about 60° to the ridge axis. The three other fabrics show dispersed orientations probably not related to ice movement direction. One fabric on till which dropped about 0.5 m from a roof of melting basal till beneath an overhang at the ice margin (plate X, g) shows no strong preferred orientation.

Although the small areal extent of ablation till, and the small number of samples make it difficult to reach statistically valid generalizations from rounding and lithology data (Appendix C), several observations can be made. Because rounding values vary for different lithologies (in one case average rounding was 0.41 for plutonic rocks, 0.36 for dike rocks, and 0.21 for metasedimentary rocks), it is difficult to deduce much from the average rounding values except that pebbles in the ablation till are probably slightly less rounded than pebbles in the basal till within a given lithology. The average number of striated pebbles (4.0 percent) in the ablation till is only half as large as the average for basal tills (8.3 percent), although the percentage of basalt pebbles present (on which most striations are found) remains nearly the same. Likewise the percentage of pebbles with fresh broken surfaces is considerably higher (23.7) in the ablation tills than in the basal tills (15.1). These data tend to support the findings of Drake (1968) on distinguishing between ablation and basal tills in New Hampshire.

More widespread than the ablation till described above are scattered cobbles and boulders that occur on the thinner parts of the ice, especially on the lee sides of nunataks (fig. 30). Evidently carried englacially, these rocks are very angular (average rounding of 2 random samples of 100 pebbles is 0.17) and some have evidently shattered in

Figure 28. Photo of ice margin between Nunataks D and E on July 27, 1969. Note white boulder (about 2 m in diameter) in upper center and ablation till in foreground.

Figure 29. Photo of ice margin between Nunataks D and E on August 12, 1970. Compare with Figure 28.

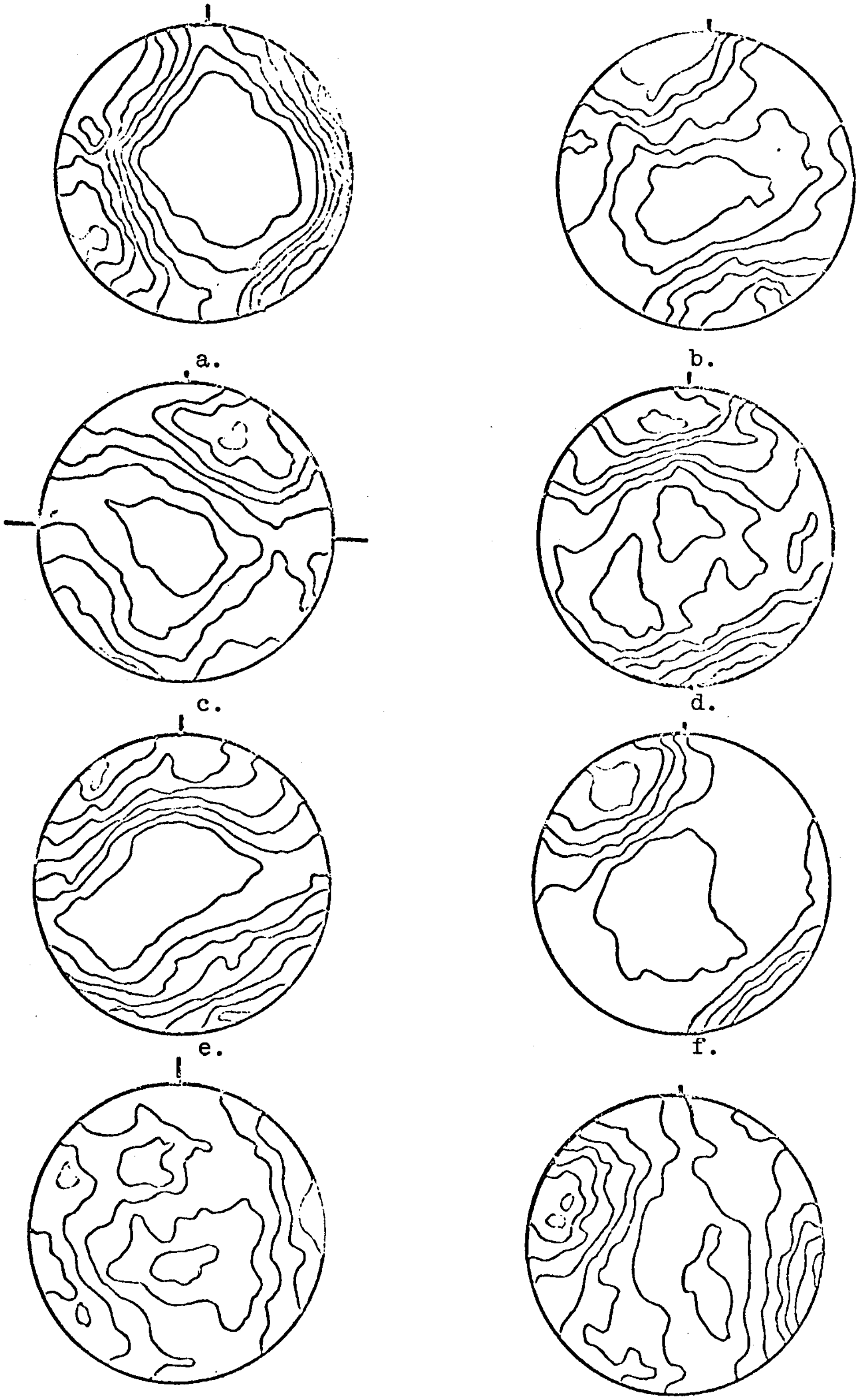

a. b. c. d. e. f. g. h.

Plate X. **Fabric diagrams for ablation tills (left column) found above basal tills (right column) at four localities. Bar on c** shows orientation of ablation till ridge. All have contour interval one sigma except f which has interval two sigma.

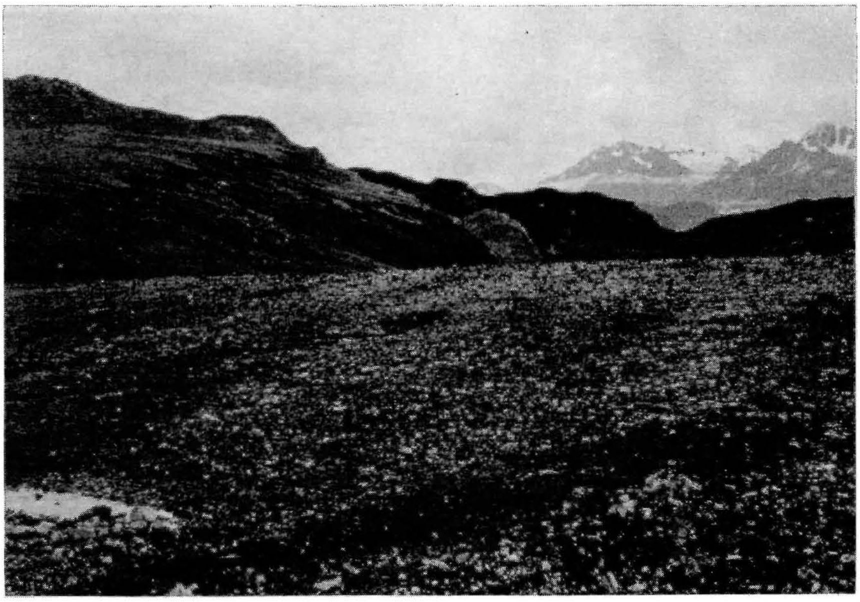


Figure 30. Photo of ablation debris on ice surface on lee side of Nunatak E. Note light colored boulder train in center.

place for pieces lie scattered in groups. No striations were found on rocks in the two samples counted. The lithology data (Appendix C, samples 147, 160) suggest that the rocks were derived locally, possibly by plucking from bedrock knobs. The amount of debris contributed by boulders and cobbles rolling down hillsides onto the ice surface is probably negligible in most areas because of the steep ice margin that developed around the nunataks as they were exposed. This is occurring now, however, on the north side of Plateau Glacier at the base of Nunatak F and has most likely taken place locally in the past.

A marble-boulder train can be traced on stagnant ice and land from the depression at the head of Gull Stream across the east end of Nunatak E and onto the ice to the east. The source of the marble is evidently still beneath the ice as no trace of it was found at the east end of the Bruce Hills. The orientation of the boulder train records early ice movement and seems to show little effect of late ice movement except that it may have been offset to the south and rotated slightly.

On the ice the boulder train is approximately 13 m wide at the base of the ice depression just north of Gull Stream. It is 25 m wide at the southeast base of Nunatak E and 29 m wide about 400 m farther to the east where the elevation of the ice surface is lower and the ice is probably thinner. Seventy-one percent of all pebbles in one area at the center of the train were marble (Appendix C, sample 160) and the percentage decreases to 6 percent marble outside the boulder train in the same area.

CHAPTER IV

ICE-CONTACT STRATIFIED DRIFT

Small eskers and kame terraces formed locally with the retreat of the Burroughs Glacier. Most are on the lee side of nunataks where they formed under and against what must have been nearly stagnant ice. Long lasting detached ice masses are noted in these lee locations in early photos. Because they have nearly all formed englacially or subglacially, the precise date of formation is difficult to determine. However, minimum dates for their formation are available from photographs and certain limits can be placed on the beginning of deposition by relating eskers to moulins or collapsed areas on the ice in early photographs. Kame terraces can be related to the age of channels that carried water to them.

In this paper an esker is considered "a long narrow ice-contact ridge commonly sinuous and composed chiefly of stratified drift" (Flint, 1971, p. 214). Kame terraces are considered to be those stratified deposits "laid down chiefly by streams between a glacier and the side of a valley" (Flint, 1971, p. 209). Also included in this category are stratified deposits that are more extensive laterally than an esker and are deposited on ice near a valley wall.

Eskers

Eskers and kame terraces have been described from many areas and are summarized by Charlesworth (1957), Flint (1971), and Embleton and King (1968). Two main hypotheses have been developed to explain esker formation. One, suggested by Shaler (1884) and adopted and developed by DeGeer (1897), proposes that eskers, especially the segmented type, form when an englacial or subglacial stream enter ponded water at the margin. Thus the ridge consists of a series of small deltas built at the ice margin as it retreats. The other possibility, evidently suggested by Hummel in 1874 (Flint, 1971) is that eskers are deposited by subglacial streams. This second hypothesis best explains most eskers in the study area.

Very few eskers have actually been seen forming. Lewis (1949) reported a recently exposed esker that had formed near an ice margin. It was composed of sand and gravel with a thin silt cap that had filled a subglacial, and in some places englacial, tube. An esker-like ridge leading into a subglacial tube in Norway has been described by Stokes (1958). The ridge was composed of unstratified drift evidently deposited by material melting out of dirty ice. The deposits occurred only on the sides of the tunnel and presumably were not deposited by running water.

Price (1966) described a series of eskers at the margin of Casement Glacier, which is 12 km east of the study area. Here several systems of eskers 3 to 18 m high formed in supraglacial or englacial and subglacial tubes both on land sloping toward and away from the ice. He noted that complex esker systems formed only in the lee of bedrock knobs.

Small esker complexes on the lee side of nunataks occur in three localities at the southeast end of the Burroughs Glacier (fig. 5). Solitary eskers occur just south of Nunatak F and at the terminus of the Cushing Glacier and smaller ones (< 1 m high) occur throughout the area.

No eskers were actually seen forming during the summers of 1969 and 1970. Since all of the esker complexes occur in the lee of nunataks it seems likely that eskers are forming or will form under the nearly stagnant ice mass south of Nunatak A. The subglacial or englacial portion of the Burroughs River now flows beneath this ice mass. Entrances to the tunnel were too small to afford access to the subglacial portion of the stream but where observed from the ice edge the eroded channel was floored by cobbles and boulders and there was no sign of deposition.

The final deposition of two recently accumulated eskers was observed over the course of two seasons. One esker which completely filled an englacial tube about 2 m above the base of the ice was nearly destroyed by subsequent melting of the underlying ice. Another, deposited in a subglacial tube with nearly vertical sides (fig. 31) remained a distinct ridge after the sides of the tube melted. In this case the sharp crest and steep sides are the result of the sand and gravel slope adjusting to its angle of repose.

Esker Complexes

Eskers in the esker complexes are composed of poorly stratified sand and gravel with cobbles as much as 15 cm in diameter. Mechanical analyses of 10 samples are given in Appendix A and are summarized in Table 8. There is no significant difference between the mean sand and silt-clay percentages of the kame terraces and eskers that were sampled. Most of the eskers head at the break in slope between the valley wall and valley bottom. Imbrication of cobbles and some cross bedding indicate that the water, which deposited the eskers, was flowing in the same direction as the ice surface gradient. On the lee sides of nunataks D and F the eskers are about 2 m high and are separated by water-filled kettle holes (fig. 32).

The eskers in the lee of Nunatak D are typical of this type of esker complex. They evidently were deposited by subglacial Burroughs River. The mean plutonic rock/metasiltstone ratio of four samples (Appendix C, samples 135, 136, 137, 140) in the eskers is 1.78, whereas the ratio of till below and on the hillslope above the eskers (samples

Figure 31. Photo of esker with nearly vertical ice walls in 1969.

Figure 32. Photo of esker on lee side of Nunatak D. Esker is 2 to 3 m high. Note kettle hole at right of esker.

Table 8. Summary of mechanical analyses of eskers, kame terraces, and outwash.

Landform	Statistic	% < 2 mm	% Sand in < 2 mm Fraction	% Silt-Clay in < 2 mm Fraction
Eskers* N = 10	Mean	68	89	11
	High	100	97	42
	Low	19	58	3
Kame Terraces N = 10	Mean	53	90	11
	High	95	98	35
	Low	33	65	2
Outwash N = 4	Mean	51	97	4
	High	69	98	5
	Low	35	95	2

* Does not include fine-grained esker south of Nunatak F

138 and 141) averages 0.43. One sample of Van Horn gravel taken across the Burroughs River in the channel wall has a ratio of 2.8. This suggests that the eskers were deposited by a stream eroding both the upper Van Horn Formation and the Glacier Bay till somewhere to the northeast. The mean roundness of the 4 esker samples is 0.41 and the mean for 10 kame terrace samples is 0.29, considerably lower than the eskers and about the same as the Glacier Bay till. The difference between these values is probably mainly due to the different lithologies of the source material rather than to differences in the duration of transport. The eskers formed beneath an ice mass where Taylor (1962a, b) measured 0.1 m of movement between 1960 and 1961. Thus the ice was nearly stagnant. The appearance of the ice mass in 1962 and the position of the present esker system is shown in Figures 33 and 34. The eskers are nearly all sharp crested and have an undulating long profile descending in elevation toward the valley floor (fig. 35). Flat crests up to 0.5 m wide are present locally and presumably are the only remnants of the original esker surface. In one case a channel cuts a ridge crest to a depth of 1 m. This may have been cut by a supraglacial stream superimposed on the esker surface as the ice surface lowered.

A small kame terrace, collapsed along its ice contact face, lies stratigraphically above the heads of the Nunatak D eskers and has a markedly different composition. It has a plutonic rock/metasiltstone ratio of 0.14 (Appendix C, sample 139) and is similar in composition to the till of the Glacier Bay formation on the hillside above. Since a dry channel leads down the south slope of Nunatak D to the kame terrace (fig. 36), it is likely that the terrace was deposited while this stream was active. The col between Nunatak D and the small knob to the south became ice free between 1958 and 1960. In 1960 water flowed from ice, across this col and under the ice in the position of the channel that leads to the kame terrace. By 1963 and perhaps as early as 1961 the water had stopped flowing across the divide and the kame terrace had presumably been deposited. If it is assumed that flow was not initiated in the channel until just before the col was deglaciated and that the kame terrace was deposited by water flowing down that channel, then there is a 2-4 year period available for kame terrace deposition. No maximum age can be placed on the eskers because the Burroughs River was flowing subglacially by 1948.

Solitary Esker

One esker, just south of Nunatak F, is approximately 275 m long (fig. 37). A longitudinal profile and representative cross sections are shown in Figures 38 and 39. The esker ranges in height from 0.5 m to 3 m. At the north end, below the moulin shown in Figure 40, the esker is composed of ice-cored sand and gravel. In its low, central portion it is composed of laminated sand and silt that is not ice cored. At its south end, where the esker is a channel 0.5 m deep in till, it is composed of sand to 0.5 m deep and gravel at the base, overlain by 30-40 cm of laminated silt which is deformed near the base. This silt

Figure 33. Photo of isolated ice mass south of Nunatak D in 1962 (Photo by R. Price). This smooth ice surface covered the eskers shown in Figure 34.

Figure 34. Photo of eskers south of Nunatak D which were ice covered in 1962 (see Figure 33).

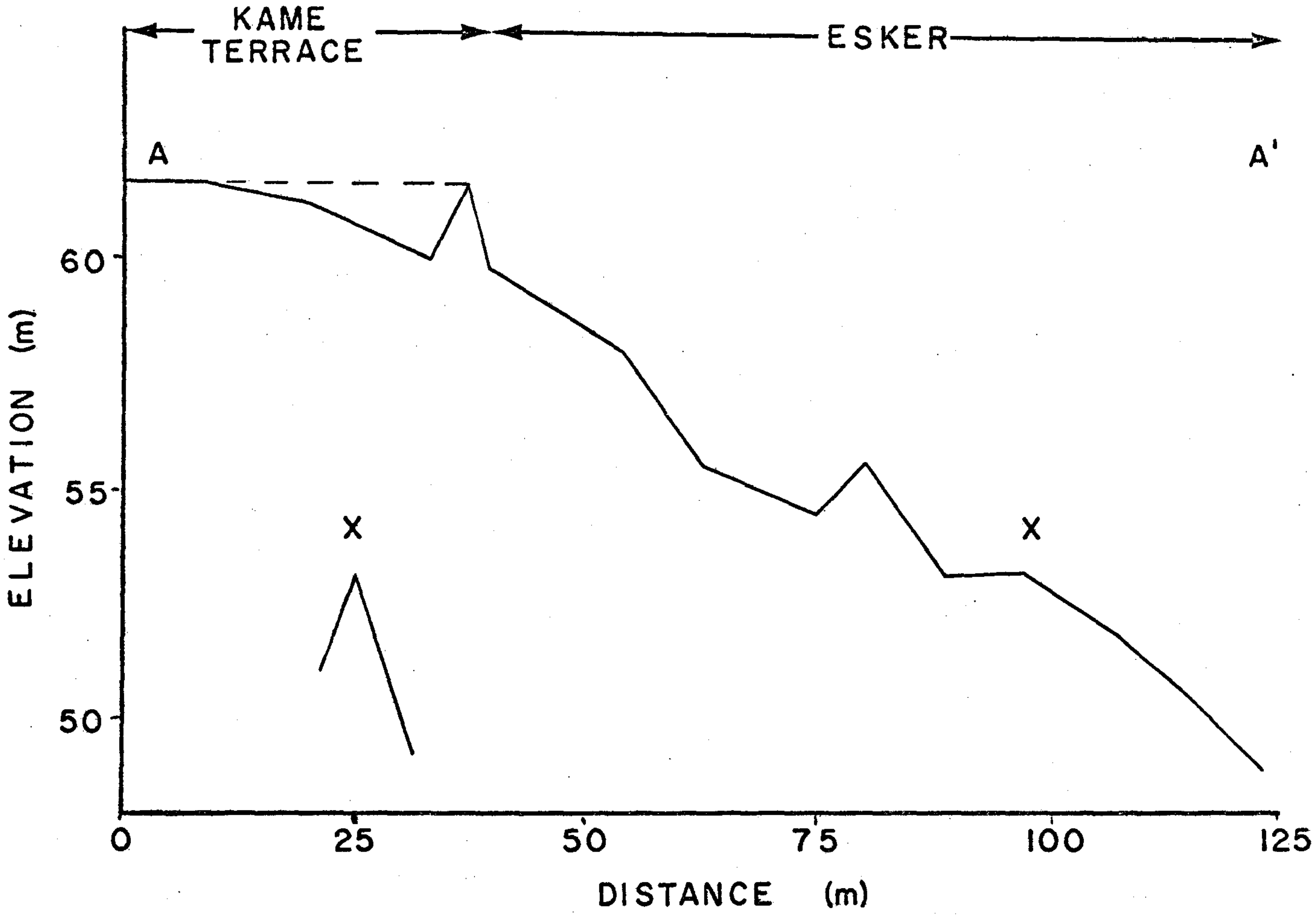


Figure 35. Typical longitudinal profile and cross section of esker south of Nunatak D.

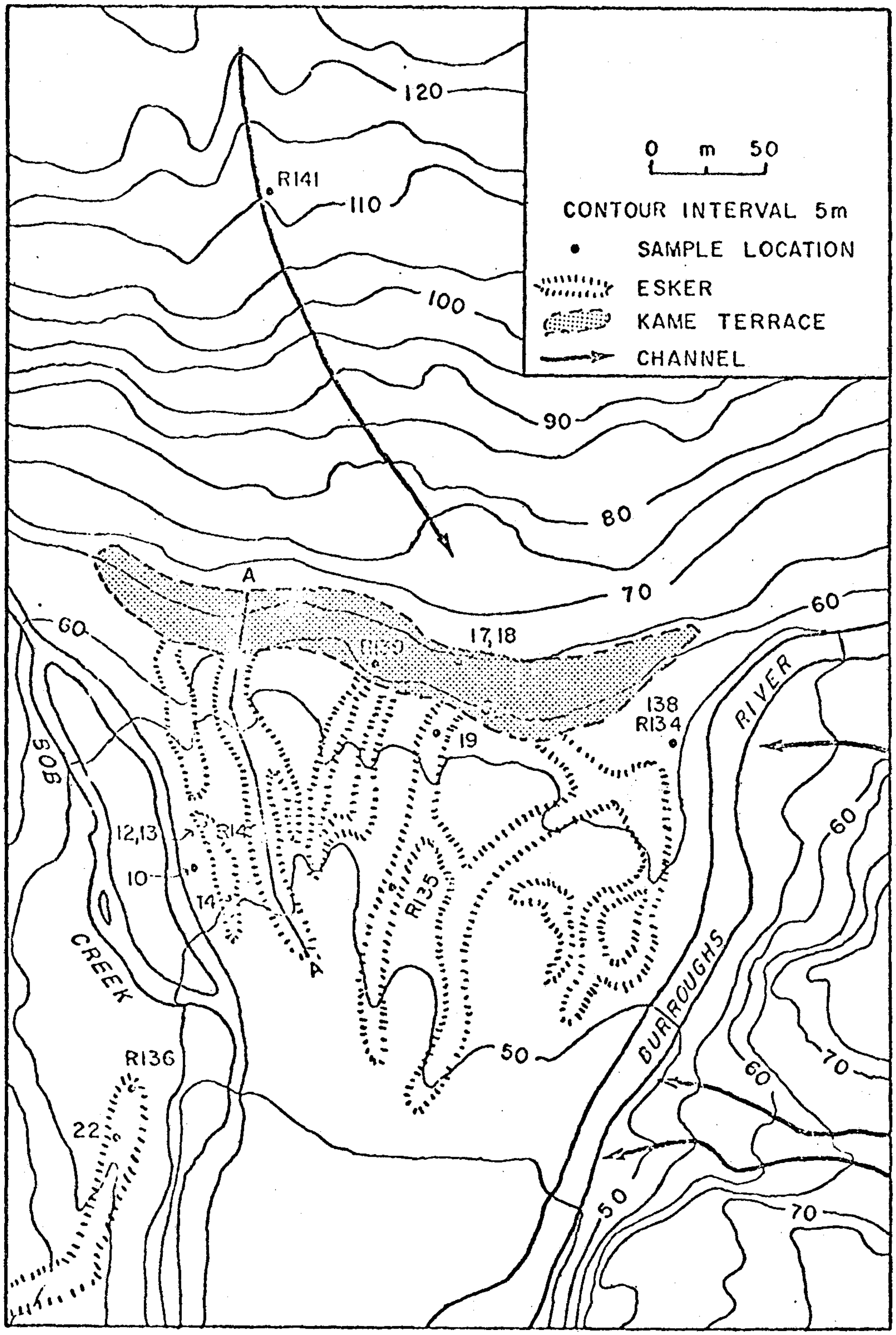


Figure 36. Map of area south of Nunatak D.

Figure 37. Photo of solitary esker south of Nunatak F looking north. Esker is composed of sand and gravel at north end (arrow).

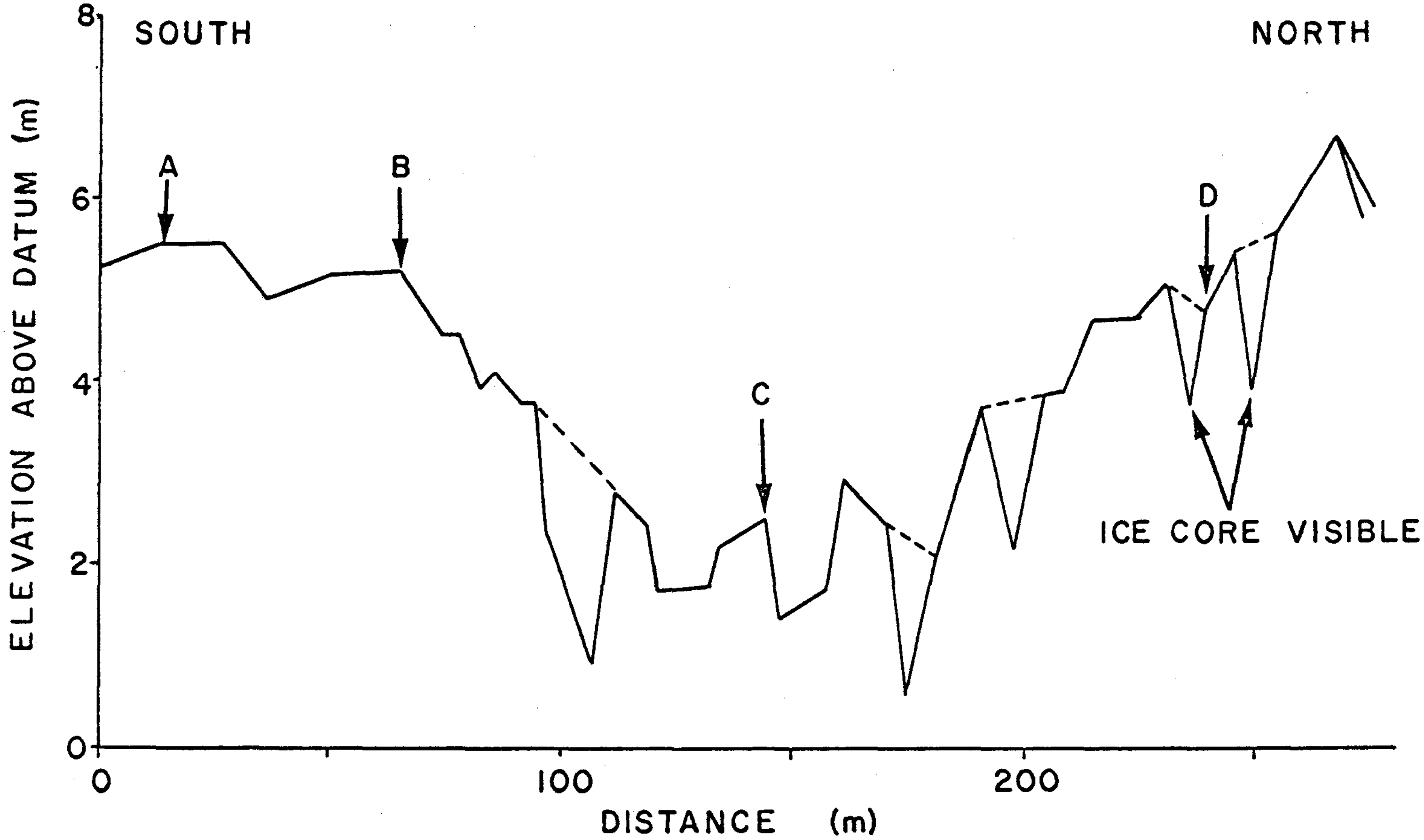


Figure 38. Longitudinal profile of esker south of Nunatak F showing positions of cross sections shown in Figure 39.

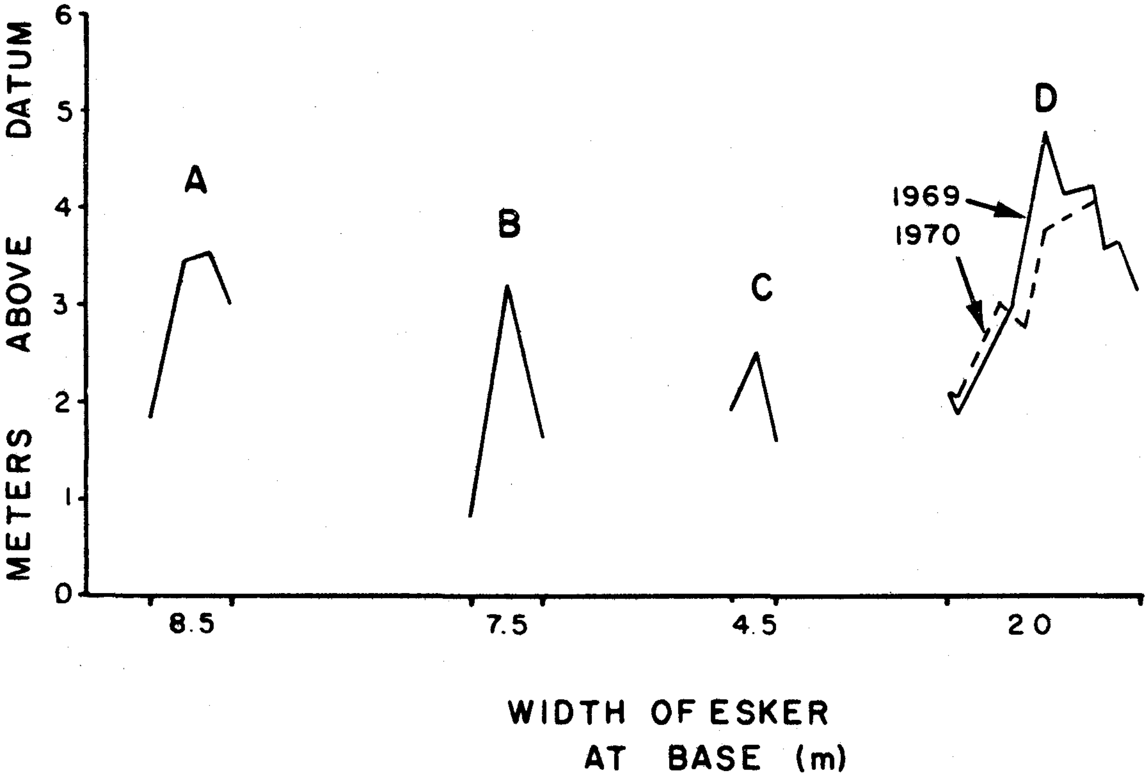


Figure 39. Cross sections of esker south of Nunatak F. A, B, and C had identical profiles in 1969 and 1970. D had changed between 1969 (solid) and 1970 (dashed)

grades upward into fine cross-bedded sand at the surface. The laminae range in thickness from 1 mm to 5 mm thick in the lower section. The thickest ones are graded and several counts indicate that there are 3-8 laminae/cm. Thus the lower section has 150-320 graded laminae. Clay mineral composition is the same in clay fractions of the dark and light laminae (Appendix B, samples 31 and 32) and they are not significantly different from the nearby till. Their size distributions are different; the dark laminae have 2.3 percent sand, 95.7 percent silt and 2.0 percent clay and the light laminae have 51.6 percent sand, 47.3 percent silt and 1.2 percent clay. Cross bedding in the upper, coarser part of the esker indicates flow from north to south.

In 1960 photographs a small moulin is present in the same position as the moulin shown in Figures 40 and 41. In 1960 a stream emerged from the ice at the same location as in 1963, thus water was evidently flowing englacially and subglacially in the present position of the esker at that time. A photograph taken by R. P. Goldthwait in 1959 shows what is evidently the same channel being cut near the ice margin. In 1961 and 1963 water still flowed out of the channel but by 1964 and more so in 1965, ponding had occurred at the edge of the ice at the channel head (fig. 41). Although no photographs are available, rates of retreat indicate that the ice cover was removed from the esker in late 1967 or 1968. The following history of formation seems likely:

1. Opening of a tube which was englacial at its head and subglacial near its mouth sometime in or before the summer of 1959. Water in the channel was under hydrostatic head because the central part of the esker is lower than either end and there is no buried ice or suggestion of collapse to show that the central portion was built on ice.

2. Erosion of a channel floor first and then deposition of gravel at the base of the esker between summer 1959 and late summer 1964.

3. At this time the hydrostatic head in the tube was insufficient to force water over the southern channel threshold and ponding occurred. This presumably happened because the ice surface lowering in the area of the moulin was faster than channel cutting at the downstream end of the channel. Deposition of laminated silt began and continued while a fluctuating head (probably daily) caused velocity fluctuations and the deposition of laminated silt.

4. As the ice surface lowered after 1964 a nearly constant difference in hydrostatic head may have been maintained but the filling of the tube probably caused an increased velocity due to the decrease in tube diameter and the deposition of fine to medium cross-bedded sand in the upper sections took place.

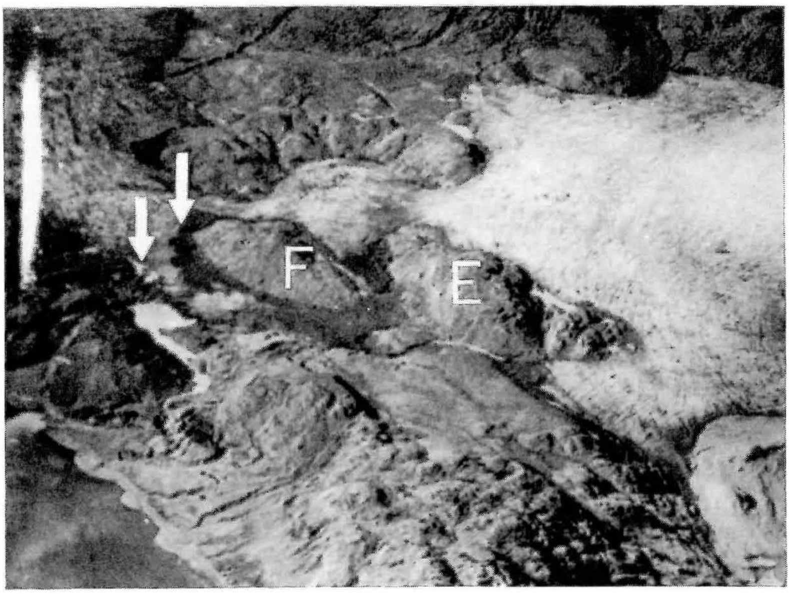


Figure 40. High oblique aerial photo (1963) showing position of esker. Note moulin at head of esker tube and lack of ponding at mouth. Nunataks F and E are labelled. Camp Creek is at lower left. (Photo by A. Post)

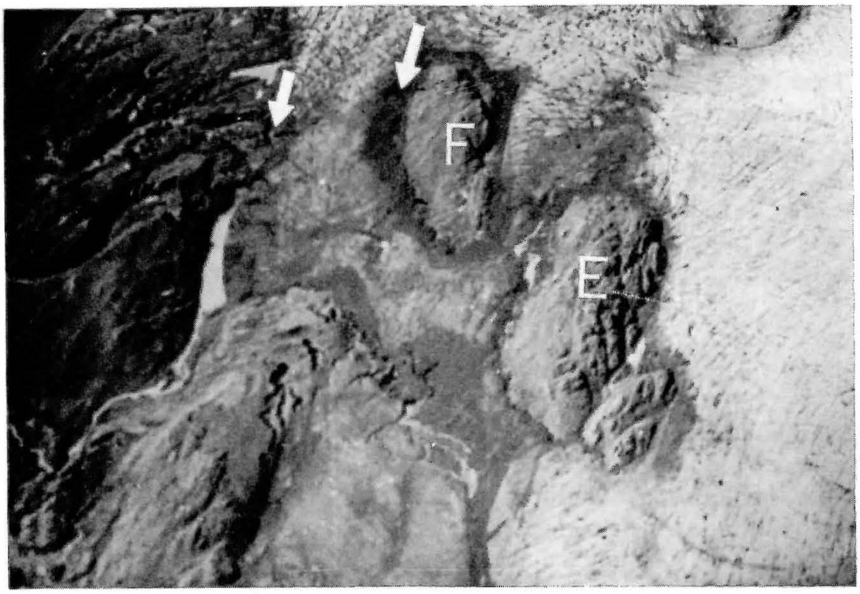


Figure 41. Low oblique aerial photo (1965) showing position of esker tube. Compare with Figure 40 and note ponding at mouth of tube. (Photo by A. Post)

5. The tunnel may either have filled completely with sediment, or water that was in the moulin may have begun draining another way in 1966 or 1967. The esker was uncovered in 1967 or 1968 and the north end (where the water flowed englacially) remained ice cored. Collapse of the esker in this ice cored section is producing a multi-crested ridge (fig. 39).

Kame Terraces

The kame terraces, which have formed and are forming around the Burroughs Glacier, are all much less extensive than many described in New England (e.g., Goldthwait, 1968) and in other areas deglaciated during the Pleistocene. Most around the Burroughs Glacier were deposited on ice and many are actually englacial. Their usefulness in determining ice margin gradients is lessened because after collapse, little or none of the original surface remains. In many locations only a hummocky blanket of sand and gravel covers till.

Dating the formation of kame terraces is difficult because the englacial terraces are not observed until their deposition is completed and they become exposed by ablation of the overlying ice. If it can be assumed that channels that lead to the kame terraces carried the water that deposited the terraces, then some limits can be placed on their age.

As is the case with esker complexes, ice-cored kame terraces occur on the lee side of most of the nunataks. They also occur along Minnesota Ridge and along the sides of Nunataks A and B (fig. 5). The kame terrace on the southeast flank of Nunatak A is composed of poorly stratified sand and gravel 2-3 m thick (fig. 42). In July 1969 no kame terrace was evident at the margin (fig. 43). By August 1970, however, the kame terrace was exposed and drainage evidently was at the base of the ice (fig. 44). This suggests that the terrace was actually deposited in ice probably before 1969 and subsequent deglaciation exposed it. At no time was a subaerial stream flowing on the terrace surface while it was being formed.

A channel leads to the east end of the terrace (fig. 62) and the channel crossing the col between Nunataks A and B ends close to the west end of the terrace. Pebble counts suggest that the terrace was built primarily by water that flowed around the east flank of Nunatak A, or by the subglacial Burroughs River. A sample from the kame terrace has a plutonic rock/metasiltstone ratio of 1.40 (Appendix C, sample 133). Till in the col channel between Nunataks A and B has a ratio of 0.24 (Appendix C, sample 122) and a small gravel terrace along this channel at the 1969 ice margin has a ratio of 0.25 (Appendix C, sample 128). Some ablation debris exists on the terrace surface, as it does on the eskers and kame terrace southeast of Nunatak D.

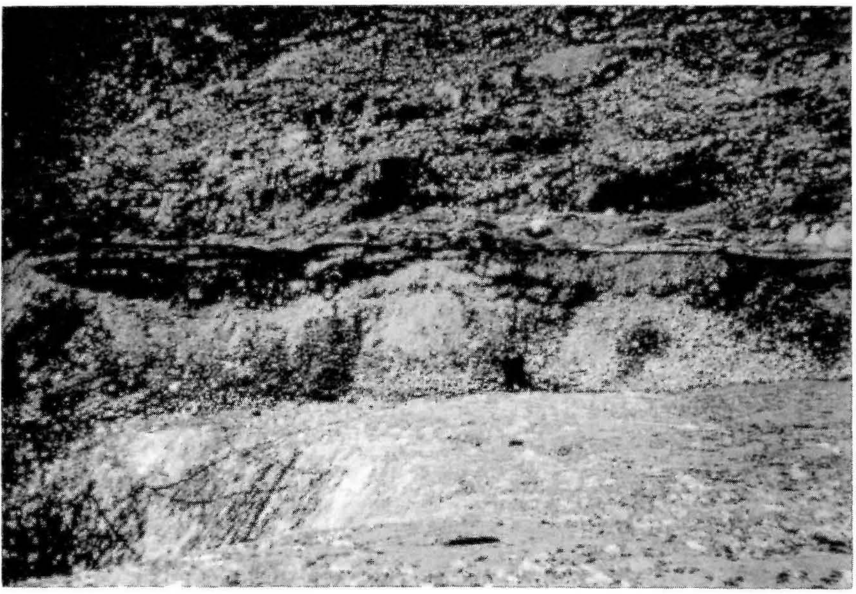


Figure 42. Photo of ice-contact face of kame terrace south of Nunatak A. Note pack in center and ice in foreground.

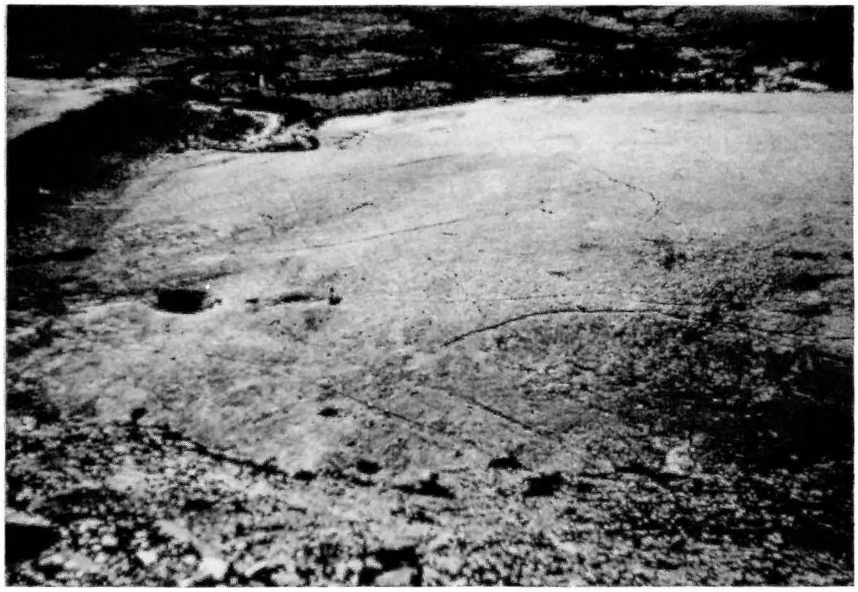


Figure 43. Photo of ice south of Nunatak A on July 14, 1969. Stream in upper left is about 3 m wide. Compare with Figure 44.

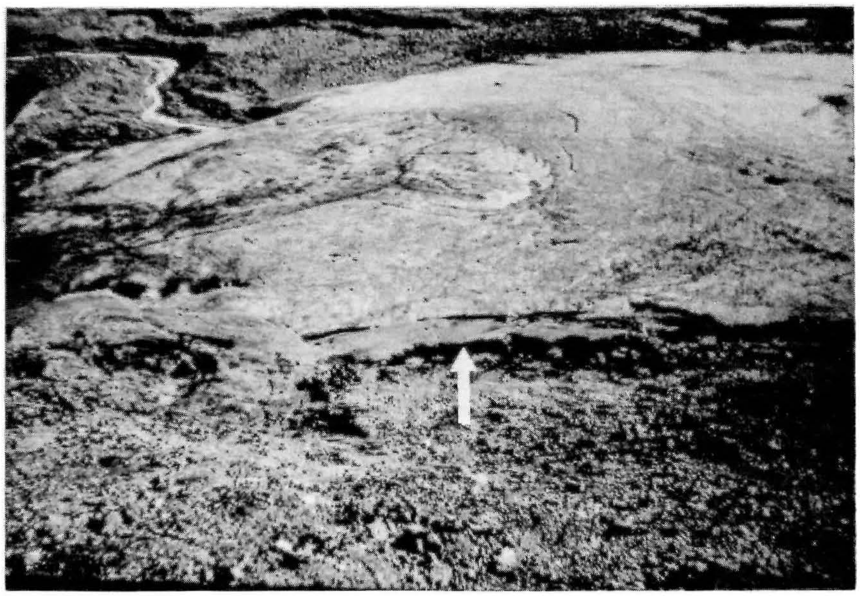


Figure 44. Photo of ice south of Nunatak A on August 12, 1970. Note kame terrace (arrow) which was exposed after Figure 43 was taken.

A similar but more extensive deposit occurs on Plateau Remnant, southeast of the Bruce Hills. Here sand and gravel, 1-3 m thick, was exposed between 1969 and 1970. The deposit has a nearly flat top and in places is capped by as much as 10 cm of silt. Some of the silt and fine cross-bedded sand in the deposit is deformed. Although part of the deposit is still within ice, the material seems to have been deposited from water flowing off the Bruce Hills and into the ice. The chute leading to the deposit has actively carried meltwater from the margin at least since 1960. Presently water is flowing at the base of the ice 10-15 m below the deposit.

A small ice-cored kame terrace formed in 1968 and early 1969 occurs on the southeast flank of Nunatak B (fig. 45 and 46). Here water flowing south along the nunatak near the edge of the ice deposited sand and gravel and finally silt on ice. This may have been submarginal in 1968, but by July 1969 the stream depositing the kame terrace flowed subaerially and ponding had occurred. By early August 1969, the stream had begun to down cut, exposing the ice core (fig. 45). By August 1970 the stream had cut a channel as much as 2 m deep in the ice and the kame terrace surface was collapsing (fig. 46).

That deposition has occurred on ice in these positions in the past can be seen by the sand and gravel blankets on much of the land surface on the lee sides of all of the nunataks and at the east end of the Bruce Hills. Figure 47 shows the hummocky sand and gravel blanket on the lee side of Nunatak E. Here a stream flowed across a col on the nunatak between 1961 (possibly before this but if so it was under ice) and 1965 or 1966 (See fig. 40 and 41 for 1963 and 1965 photos). Presumably the water flowed under the ice margin and out toward camp lake in its early stages and became submarginal only in its late stages. Only a small uncollapsed kame terrace segment low on the hillside remains.

Small uncollapsed kame terrace segments exist above the ice margin on the south flanks of Minnesota Ridge and the Bruce Hills. As discussed later, most of the drainage here is submarginal, subglacial or englacial and most terraces occur at the base of subaerial gullies which may have been submarginal chutes (fig. 48).

The slopes of three terraces less than five years old, ice margins adjacent to the terraces, and ice surfaces in the central part of the glacier were determined with a Wild A7 plotter (table 9). The slopes of the terraces decrease with increasing elevation as do the slopes of the adjacent ice margin and the ice surfaces in the central part of the glacier. Although the gradient of the lower terrace differs from the gradient of the adjacent ice margin, they are both steeper than the ice surface gradient. The slopes of the other two terraces agree fairly well. Thus the gradients of a number of individual terrace segments can be used in reconstructing the slope of the ice margin.

Figure 45. Photo of ice-cored kame terrace on August 11, 1969. Terrace front is about 1.5 m high.

Figure 46. Photo of kame terrace from same position as Figure 45 on August 12, 1970. Note collapse of terrace surface.

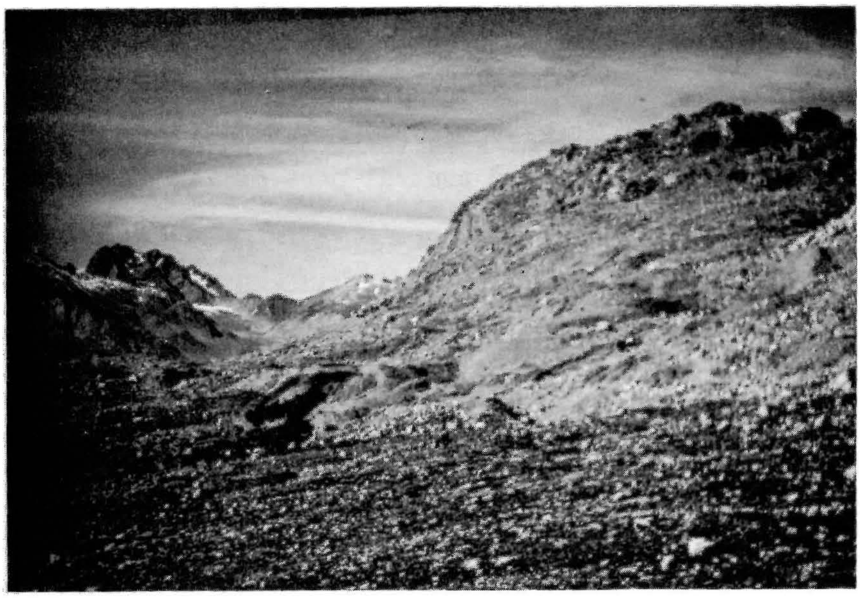


Figure 47. Photo of collapsed sand and gravel on lee side of Nunatak E. Largest boulders on ice in foreground are about 0.5 m in diameter.

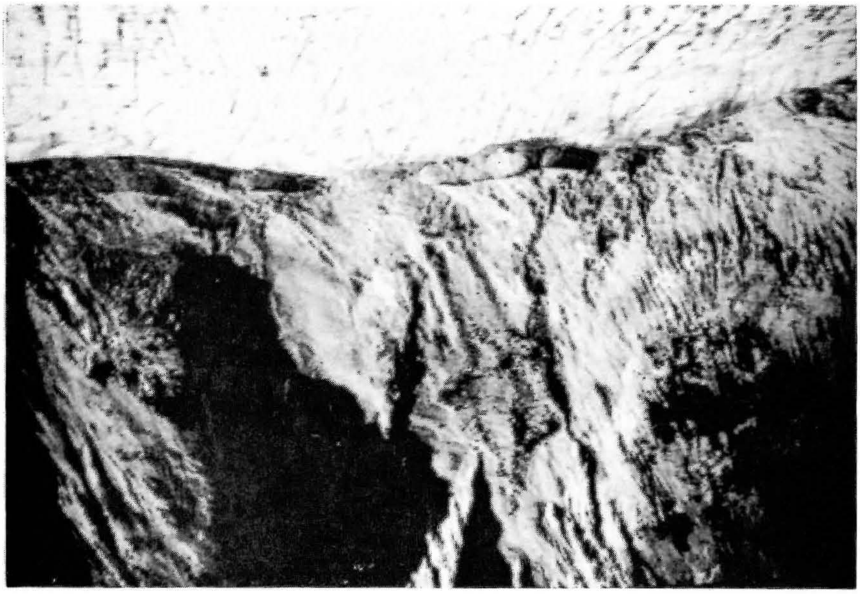


Figure 48. Photo of ice margin at base of Minnesota Ridge. Note small (about 20 m long) terrace segments along margin.

Table 9. Slopes of kame terraces, adjacent ice margin and central part of ice at base of Minnesota Ridge.

Range in Elevation of Kame Terrace (m)	Slope (%)		
	Kame Terrace	Ice Margin	Central Part of Ice
335-300	16.7	10.5	6.9
386-369	11.4	10.0	4.8
434-424	6.7	6.1	3.6

CHAPTER V

LAKE DEPOSITS AND OUTWASH

Ice-Dammed Lakes and Spillways

Ice-dammed lakes and spillways have been described around present glaciers (Thorarinsson, 1939; Lindsay, 1966) and their existence at the margins of Pleistocene ice sheets is well established. Fairchild (1899) described a number of channels related to lakes in the Finger Lakes region of New York. Shortly thereafter Kendall and Muff (1902), in a classic paper on marginal drainage in the Cheviot Hills of Scotland, attributed most channels to the draining of marginal lakes. Since that time the spillway origin of many channels has been advanced by Twidale (1956), Lougee (1940, 1953, 1956) and others. This origin for many channels has recently been questioned by Price (1960), Sissons (1958, 1960, 1961) and others.

That some marginal lakes did exist in many areas during the retreat of Pleistocene ice cannot be questioned; however, debate centers around the evidence necessary to prove the former existence of lakes and thus the origin of channels related to them. Sissons (1960) has cited the rarity of shorelines, rarity of deltas, lack of lake sediment, and the absence of channels in areas where they would be expected as evidence of marginal rather than overflow channels.

Although no large lakes exist around the terminus of the Burroughs Glacier, water has been and is presently ponded in several places along the ice margin. Several of these small lakes are described below with reference to the criteria listed above for the existence of a former lake and to the outlet channels which have formed.

Lakes Around Burroughs Glacier

One of the earliest lakes which developed along the retreating ice margin was one just south of station 7 (fig. 5). The lake came into existence in or just after 1930, was present at least until 1950, and had drained completely by 1956. Photos taken by W. O. Field in 1941 and 1950 are shown in Figures 49 and 50. Until sometime between 1941 (fig. 49) and 1946 the lake drained over the outlet at 167 m. A shoreline is visible at this level in the 1950 photograph (fig. 50) but by 1958 the shoreline position is indistinct on photographs because of tree cover. In 1970 the 1941-level shoreline on the south side of the lake was marked by sand about 10 cm thick and by a delta 8 to 10 m across at the base of the stream east of station 7. The delta surface is at approximately the same elevation as the outlet.

Figure 49. Photo of ice-dammed lake from station 7 in 1941. (Photo by W. O. Field)

Figure 50. Photo of ice-dammed lake from station 7 in 1950. Note 1941 shoreline about 20 m above lake. (Photo by W. O. Field)

Thin sand patches occur below the 1941 level throughout the basin. A distinct shoreline is present at the 1946-1948 lake level (150 m) and a second delta about 10 m across occurs below the delta described above. A small sand deposit, possibly built against ice is present at about this level just east of the present pond. By 1950 the lake level had dropped several meters and was evidently draining sub-marginally through a high channel leading to the Burroughs River (fig. 62). By 1956 the lake had drained completely.

Another small lake was present at least until 1950 on the north side of station 7. This also had drained by 1956, presumably englacially or subglacially to the Burroughs River. Two fairly small channels presently lead from the former lake position under the ice and it cannot be determined which channel carried water first or if they were both active at the same time. A blanket of sand, up to 1 m thick in places, covers the former lake bottom.

Muir Lake, which formed at the head of the drainage to Muir Inlet at the southeast end of Minnesota Ridge (fig. 5), evidently has a more detailed history. It evidently drained periodically through a subglacial chute (fig. 51). In late summer 1956 the ice margin was close to the col shown in Figure 52. By late August 1958 the margin had crossed the col and a small lakelet had formed (Field, 1959). The ponded water was present to the level of the col (154 m) in August 1959 but by 1960 the lake had drained, presumably subglacially. Ponded water was again present on August 16, 1961 to nearly the elevation of the divide. On August 24, 1963 the lake was present but seems to have been at a level several meters below the divide. By August 30, 1964 the lake was present but at a level about 6 m below that of 1963. Only oblique aerial photos are available for 1965, but the lake appears considerably smaller than in 1964. By 1969 the lake drained subglacially through a tunnel at the base of the ice.

The lake sediments occur at three indistinct levels in the basin. The highest, at about the level of the outlet, consists of poorly bedded sand deposited against gravel which may have been deposited earlier as a kame terrace. The surface of the sand has an estimated gradient of 1/100 toward the outlet. In the central part of the basin, about 0.5 m of silt (Appendix A, sample 133) grades upward to a silty sand near the top (Appendix A, sample 134) with a total thickness of about 8 m. A sediment level 7.5 m below the upper level (possibly related to the 1964 lake level) consists of poorly stratified sand and silt. It is present only in the part of the basin occupied by the 1964 lake. Cross bedding that was measured at this level indicated flow toward the basin center indicating drainage beneath the ice.

Another fairly distinct level occurs about 28 m below the high level and is sand and gravel blanketed in places by silt (Appendix A, sample 135) up to 0.5 m thick. This may represent the bottom deposit of the 1964 lake or may have been deposited later. In the central part

Figure 51. Photo of ice-cored lake deposits (foreground) and ice tunnel through which lake drained. Ice surface over tunnel entrance is about 10 m high.

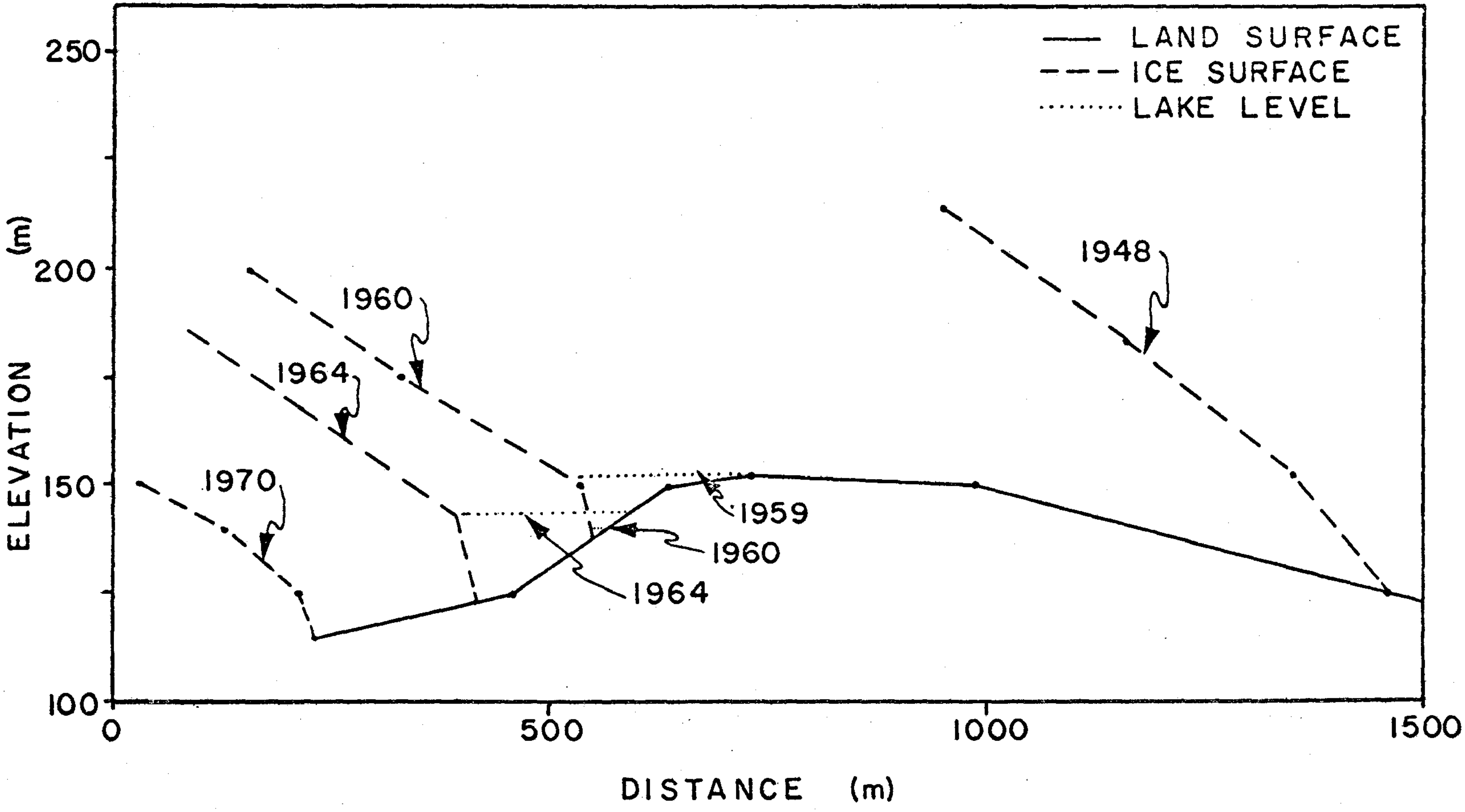


Figure 52. Longitudinal profile across Muir Lake area showing ice positions and lake levels.

of the basin a stream with an estimated discharge of about 6 cfs after a moderate rain flows toward the ice. A channel up to 4 m deep has been cut in the lake sediment.

A small lake located between Nunataks A and B came into existence when the ice margin ponded water against the col between the two hills in the melt seasons of 1961 or 1962. In 1963 the lake drained submarginally through the col channel shown in Figure 62. This channel is about 3 m deep in till and bedrock has been cut to a maximum depth of about 0.5 m. By late summer 1965 the margin had withdrawn from the overflow channel but the channel still carried water.

Shortly after this time water began draining around the east side of Nunatak A and the water level dropped to about 182 m. The outlet was along the ice margin and the channel has the same significance as a marginal channel. No lower outlets are present except beneath the ice margin where the lake drained in 1969 and 1970.

A small delta, possibly built against ice along part of its front, occurs on the west side of the lake with its surface at the same level as the post-1965 outlet channel (182 m). It was deposited when water from a small ice tongue between Nunataks B and C carried sand and gravel into the lake. This relationship indicates that ice was higher toward the northwest at the time the delta was deposited. With the exception of the delta, deposits in the lake above the 1970 level are thin (5-30 cm) and consist mainly of silt (Appendix A, sample 125). The deposits will be difficult to detect in several decades.

The last of the lakes to be described is Gull Lake, located about 200 m west of the boundary of Figure 5, southeast of Plateau Remnant. It seems to have been drained by marginal and submarginal outlets throughout its history and to have been less than 1 m deep since its inception after 1965. During 1969 and 1970 water level fluctuated daily up to 30 cm and there was always a current toward the outlet. Thus the deposit is midway between an ice-dammed lake deposit and an outwash deposit. Marginal or submarginal channels occur above the lake outlets (above A, fig. 53) and were formed as the ice margin retreated down the ridge between Gull Lake and Gull Stream. The highest shoreline on the lake is visible on aerial photographs and is marked by about 0.5 m of sand. Sediment thickness in the central part of the lake is variable but averages about 0.5 m of fine sand overlying 0.8 to 1.0 m of coarse sand. Much of the sand is evidently derived from the Burroughs Glacier through a subglacial tunnel in Plateau Remnant occupied by Gull Stream. This is discussed further under outwash.

Identification of Former Lakes after Deglaciation

Because of mixing due to tree throw, soil forming processes, burrowing animals and mass wasting, the presence of the small lakes around the Burroughs Glacier will probably be difficult to demonstrate

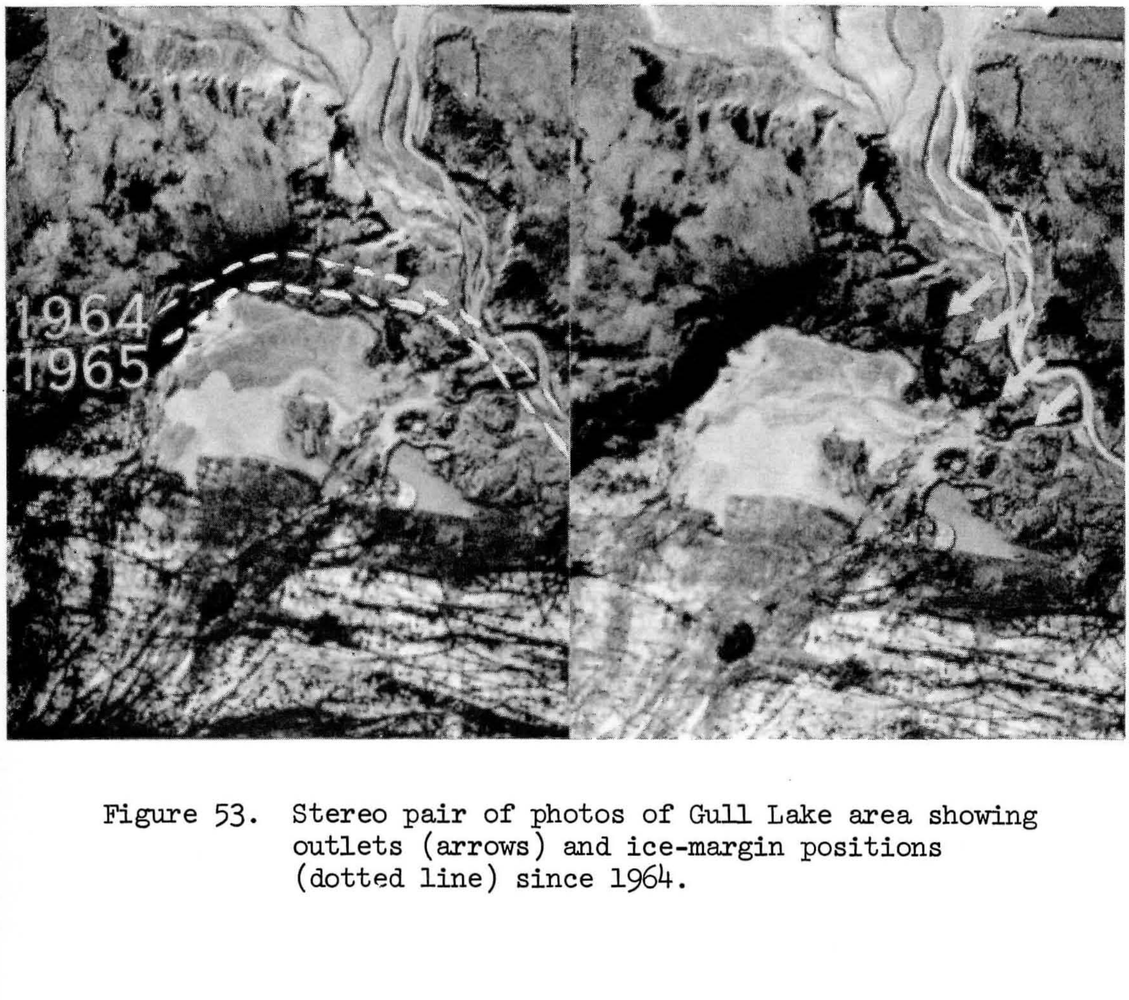


Figure 53. Stereo pair of photos of Gull Lake area showing outlets (arrows) and ice-margin positions (dotted line) since 1964.

on the basis of shorelines and lake sediment will be obscure in a century; only Muir Lake and Gull Lake have a significant thickness of lake sediment. Likewise the small deltas will be poorly preserved and will indicate little about the size of the lakes. The lack of these features is, therefore, probably not sound evidence against the presence of small short-lived ice marginal lakes. The absence of channels in certain positions might argue against the presence of a lake but it is difficult to prove that the lake did not drain supraglacially or englacially.

Another possible way of distinguishing lake outlets from marginal channels is the absence of the sand-silt apron found at the heads of nearly all marginal channels around the Burroughs Glacier (fig. 72). These deposits would be unlikely to form if a channel were used only as a lake outlet because lake spillways would be an area of increasing water velocity rather than decreasing water velocity as is the case in the late stage of marginal channel formation (See Chapter VI). The preservation of these features is probably as likely as the small deltas and thin deposits of a lake. Thus the presence of the thin apron and lack of lake sediment would argue strongly for the presence of a marginal channel rather than an outlet of an extensive lake.

Outwash

Outwash has been deposited as uncollapsed valley train, collapsed and pitted outwash, and outwash deltas during the wasting and retreat of the Burroughs Glacier. Subsequent down-cutting has occurred in nearly all of the valleys and on the deltas. The deltas have also been modified by wave action along the edge of Wachusett Inlet.

Continuous discharge measurements were made on three streams during the summer of 1970. Camp Creek, a stream which is not presently carrying glacial meltwater, had a maximum discharge of 36 cfs. This occurred during and just after a rainstorm and fluctuations throughout the summer were in phase with meteorological changes. In Bob Creek, which fluctuated daily, a gauge about 50 m downstream from the ice margin recorded discharge ranging from 2 to 43 cfs. A detailed discussion of stream discharge, its relation to ablation rates and to water tables in the ice is given by Larson (in preparation).

Valley Train Deposits

Outwash confined to channels cut in bedrock or drift occurs along the Burroughs River and its tributary Bob Creek, Gull Stream, and several short channels leading to Wachusett Inlet (fig. 5).

The Burroughs River, with an estimated discharge of 300-500 cfs, is fed by a subglacial stream which flows under the mass of ice south of Nunatak A, by John and Bob Creeks, and by other runoff within the basin. The river is presently confined to a single channel through most of its length and has built a delta at the edge of Wachusett Inlet. Since bedrock crops out along the channel in several places, it seems likely that the outwash is less than 3 or 4 m deep in most places except in the area of the delta.

Sand and gravel with boulders up to 0.5 m in diameter has been deposited in the valley. Composition of pebbles and rounding index at three locations are given in Appendix C. The lithologic counts suggest that the outwash is derived both from till and the Upper Van Horn gravels that are present beneath the till near the head of the stream. The rounding value is slightly higher for the sample collected farthest downstream (Appendix C, sample 157) suggesting that some rounding has taken place during transport.

Subaerial deposition began in the valley between 1941 and 1948. By 1948 an ice tongue had retreated about 700 m up this valley and a steep valley train was being deposited (fig. 54). By 1960 the ice tongue had withdrawn another 600 m and the stream flowed at a lower level than in 1948 except in the area of the fan where aggradation had taken place and the stream flowed at a slightly higher elevation. By 1970 the ice had withdrawn another 750 m and the river was flowing at a level nearly 12 m below the 1960 level. Remnants of intermediate levels are present locally along the valley although there are no well defined terraces (fig. 55). Outwash presumably deposited since 1964 surrounds eskers which occur south of Nunatak D and is derived both from the Burroughs River and from Bob Creek (fig. 55).

In Bob Creek outwash at least 2 m thick has been deposited between the stream gauge and the ice margin. Between mid-summer 1969 and mid-summer 1970 at least 0.25 m of degradation has taken place in the stream bottom 10-20 m upstream from the gauge (fig. 56). A small terrace remnant is present on the south side of the stream about 1.0 m above the present stream level. If it was deposited subaerially, it was deposited after 1964. Since that time degradation has taken place in the whole area above the bedrock threshold where the gauge is located. Outwash is presently being deposited on ice near the margin.

In Gull Creek debris is being carried through a bedrock channel up to 2 m deep. Its high discharge is estimated at about 200 cfs. Little outwash has been deposited except at the base of a steep bedrock-floored channel directly above Plateau Remnant and in Gull Lake. Sample 118 (Appendix C) has a rounding of 0.28 and was collected at the head of the stream. Another, collected about half way down the stream (Appendix C, sample 145), has a rounding of 0.30, and from the lower end of the stream at Plateau Remnant (Appendix C, sample 143) the sample has a rounding of 0.32. All samples have about the same plutonic rock/metasiltstone ratio but addition of debris through erosion of till and bedrock may account for the relatively small increase in rounding. Much of the outwash was deposited in or on ice and has subsequently collapsed.

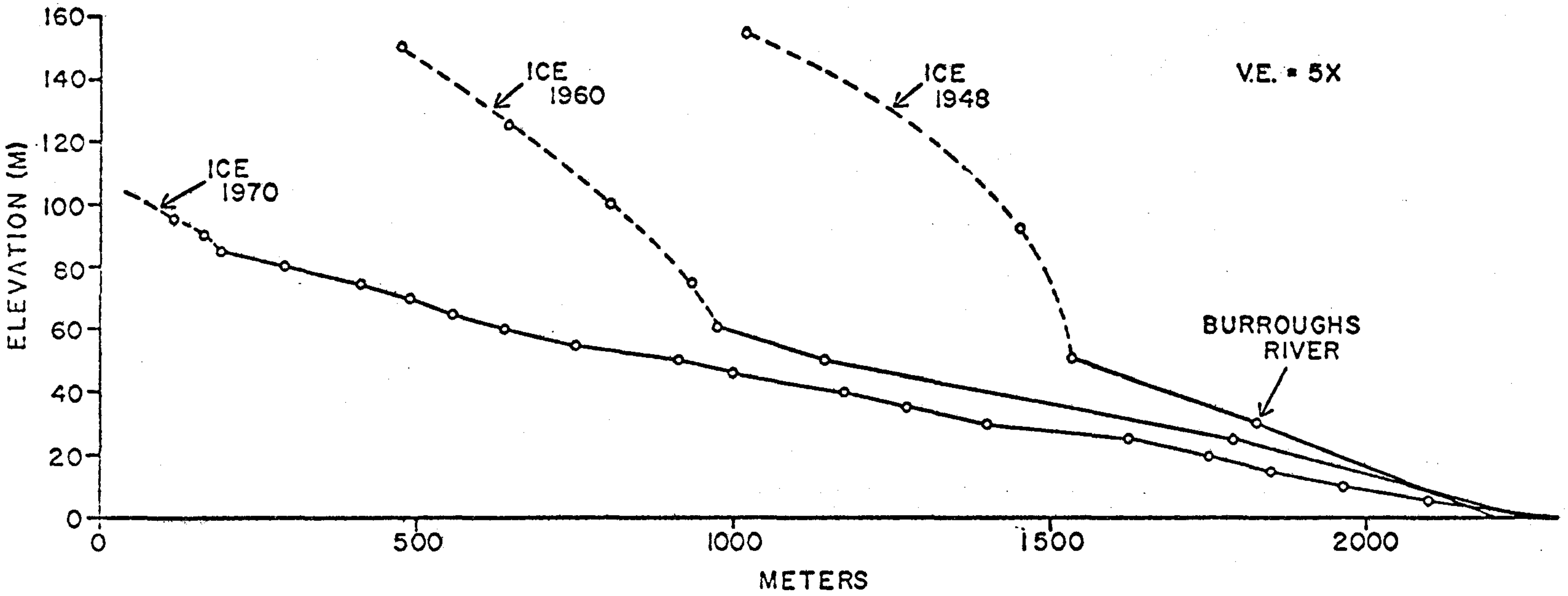


Figure 54. Longitudinal profiles of Burroughs River outwash.

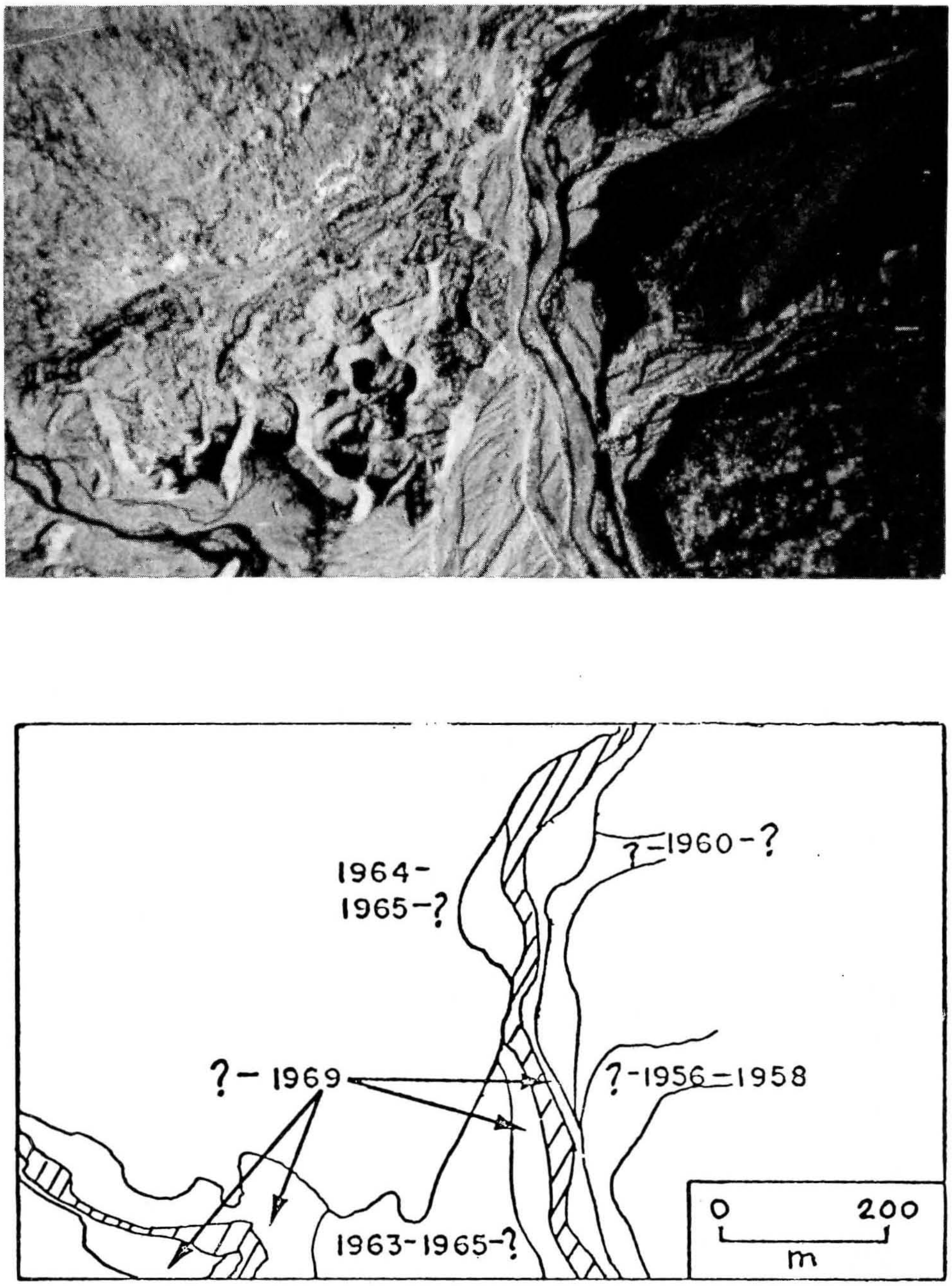


Figure 55. Photo of Burroughs River south of Nunatak D showing outwash levels active between the dates given.

Figure 56. Photos of outwash in Bob Creek in 1969 (above) and 1970 (below). Discharge was equal. Note cutting of about 20 cm. Stream is about 20 m wide.

Collapsed Outwash

The only extensive collapsed outwash in the area occurs around the margin and under Camp Lake. Between the mid-1950's and 1963, water flowing through a series of marginal and submarginal channels (see Chapter VI) carried sand and gravel onto a tongue of ice in the present position of Camp Lake. In 1959 the ice was buried in the eastern lake area (foreground of fig. 57, 58, 59). By 1962 the ice had withdrawn about 70 m and ponded water was present locally (fig. 58). Note that the stream was confined in both years to the southwest side of the valley. By 1969 a small terrace (fig. 59) was present at this location. Since the terrace is 3 m higher than the lake level and the lake is up to 6 m deep, a buried ice mass up to 9 m thick has melted out in a maximum of 8 years. This rate of melting is nearly 5 times that calculated by McKenzie (1969) in Adams Inlet. However, he did not include heat transfer by percolating surface water which is no doubt important in the Camp Lake area. After the ice surface melted down to the level of the lake outlet ponding would have occurred, increasing ablation and in late stages probably allowing the ice mass to float.

Outwash Deltas

Throughout much of Glacier Bay outwash deltas have been produced by meltwater streams flowing into inlets. The most extensive delta on the north shore of Wachusett Inlet developed at the mouth of Burroughs River (fig. 60). The delta is composed of sand and gravel with cobbles and boulders up to 25 cm in diameter. The stream cutting the delta remained within its channel during 1969 and 1970 and seems to have been in the same position since 1964. Aggradation of the delta above sea level seems to have halted.

Delta building began sometime between 1941 and 1948. No data on the total thickness of the delta are available but the fan seems to have increased in area by $2.7 \times 10^5 m^2$ between 1948 and 1960 and to have increased in area by $0.6 \times 10^5 m^2$ between 1960 and 1970. Profiles on the stream crossing the delta have poor control but they are shown in Figure 61. The delta seems to have prograded outward approximately from the level of high tide since 1948 and to have downcut above this. Since topographic control is poor relative to the slopes involved, calculations of value of sediment deposited would probably be unrealistic. Since the amount of material removed is greater than the amount deposited above mean sea level a considerable part of delta is probably being built below sea level at this time.

Finer-grained sediment is present on the surface of the deltas from a point just above low tide into deeper water. Clay minerals were analyzed from these deposits on the Gull Creek and Camp Creek deltas. The clay mineral assemblages, when compared to those in Gull Lake and Camp Lake (table 10), suggest that the brackish environment of the inlet has little effect on the clay-mineral assemblages. Similar results are

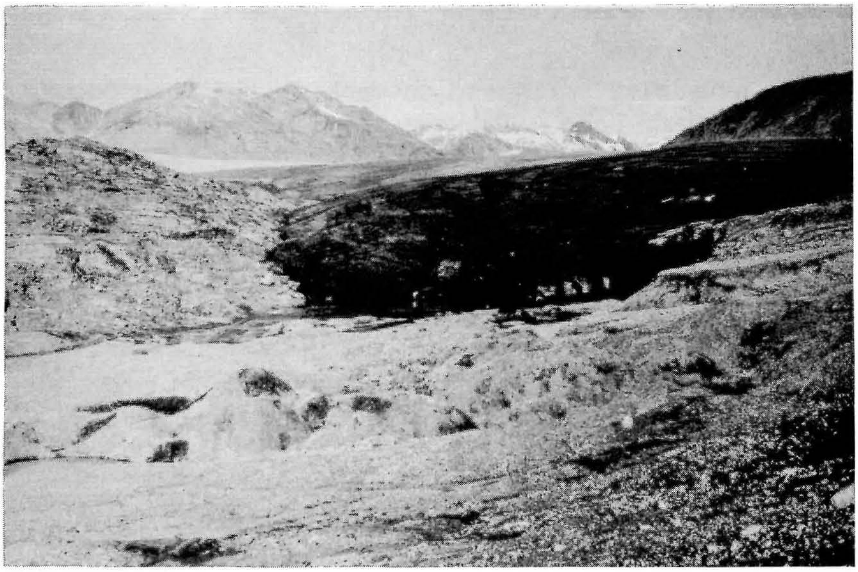


Figure 57. Photo of Camp Lake area taken in 1959. Note white boulder in left center (about 2 m in diameter). (Photo by L. Taylor)

Figure 58. Photo of Camp Lake area taken in 1961. Note white boulder in center. (Photo by R. Price)

Figure 59. Photo of Camp Lake area taken in 1969. Note collapse of outwash and formation of lake up to 6 m deep and position of white boulder.

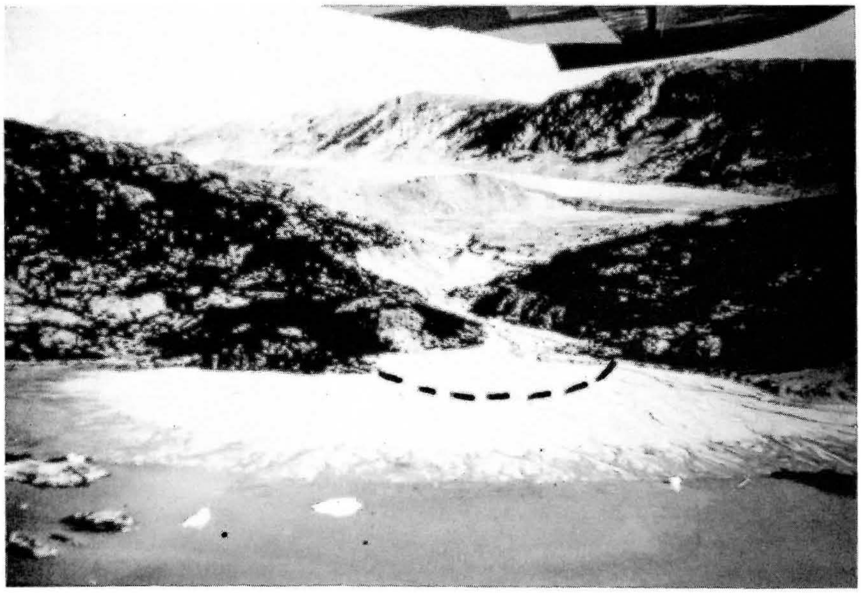


Figure 60. Aerial oblique photo of outwash delta at mouth of Burroughs River with approximate extent in 1948 and 1960 marked. Total width of delta is nearly 1 km.

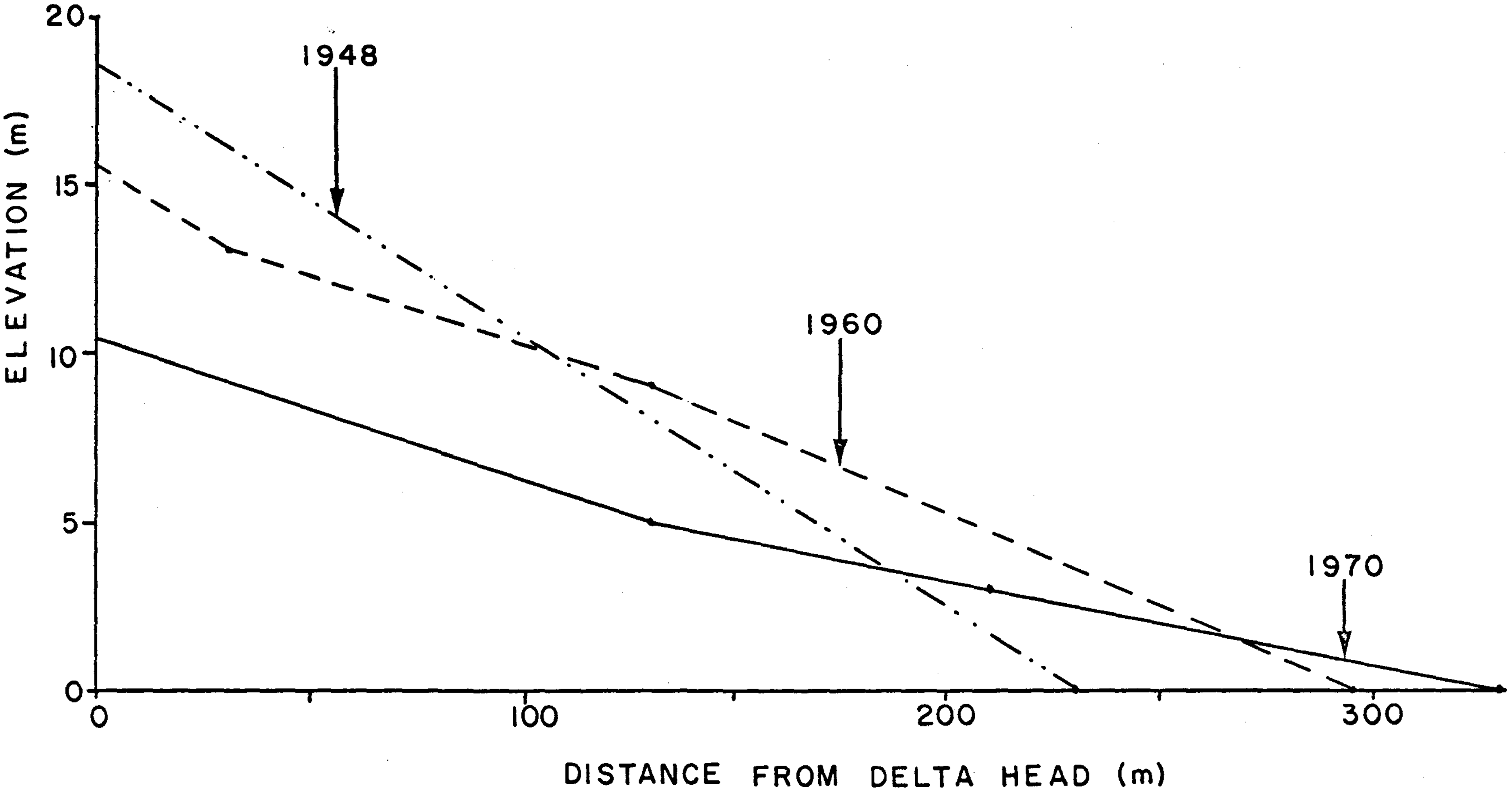


Figure 61. Longitudinal profiles of Burroughs River across outwash delta.

Table 10. Clay mineral percentages (rounded to nearest 5%) of < 2μ fraction at four locations.

Location	% Illite	% Vermiculite	% Quartz	% Chlorite	% Expandable
Gull Lake	5	T	15	65	10
Gull Creek Delta (Intertidal)	15	T	10	70	5
Camp Lake	15	10	30	45	T
Camp Creek Delta (Intertidal)	10	10	40	25	10

reported from Adams Inlet (McKenzie, 1968) and the central part of Muir Inlet and other inlets in Glacier Bay (O'Brien and Burrell, 1970).

CHAPTER VI

EROSIONAL FEATURES

Meltwater Channels

Since Clough (1895) suggested that dry channels in Great Britain might have been cut by glacial-meltwater streams, ice-related channels of Pleistocene age have been used extensively in hilly regions to determine the nature and positions of Pleistocene ice margins. Several types of meltwater channels have formed recently at the Burroughs Glacier and their characteristics, relation to the ice margin, and rate of formation are discussed below and their distribution is shown in Figure 62.

The classification used herein is a combination of those in the literature, especially those of Sissons (1960, 1961) and Derbyshire (1958). It is based primarily on the relationship of the channel to the ice margin at the time of meltwater channel formation. Direct channels, those that carry water directly away from the ice margin, are considered in the chapter on lake deposits and outwash. Glacial-lake spillways, because their formation is so closely tied to the history of the lake that they drain, are also considered elsewhere.

Marginal Channels

Marginal channels have traditionally figured in debates about the mode of late Wisconsin deglaciation, especially in New England (Goldthwait, 1938; Goldthwait, 1968; Lougee, 1940, 1956) and in other areas where downwasting and stagnation probably took place. Most arguments (Mannerfelt, 1949, 1960; Gjessing, 1960) assume that marginal channels reflect the gradient of the ice margin at the time the channels were cut. Thus if progressively lower marginal channels on valley walls change in gradient, then the slope of the margin also changed as the ice surface lowered. Likewise, if marginal channels in one area carried water in several different directions, then a stagnating ice mass with gradients in those directions can be postulated.

Marginal channels are considered to be channels that form along, or just beneath and roughly parallel to, an ice margin. Some channels may be marginal along part of their length but slightly submarginal in other parts. Distinguishing these channel segments as marginal or submarginal after deglaciation is difficult and probably not necessary if they are to be used to determine only the direction of ice margin slopes.

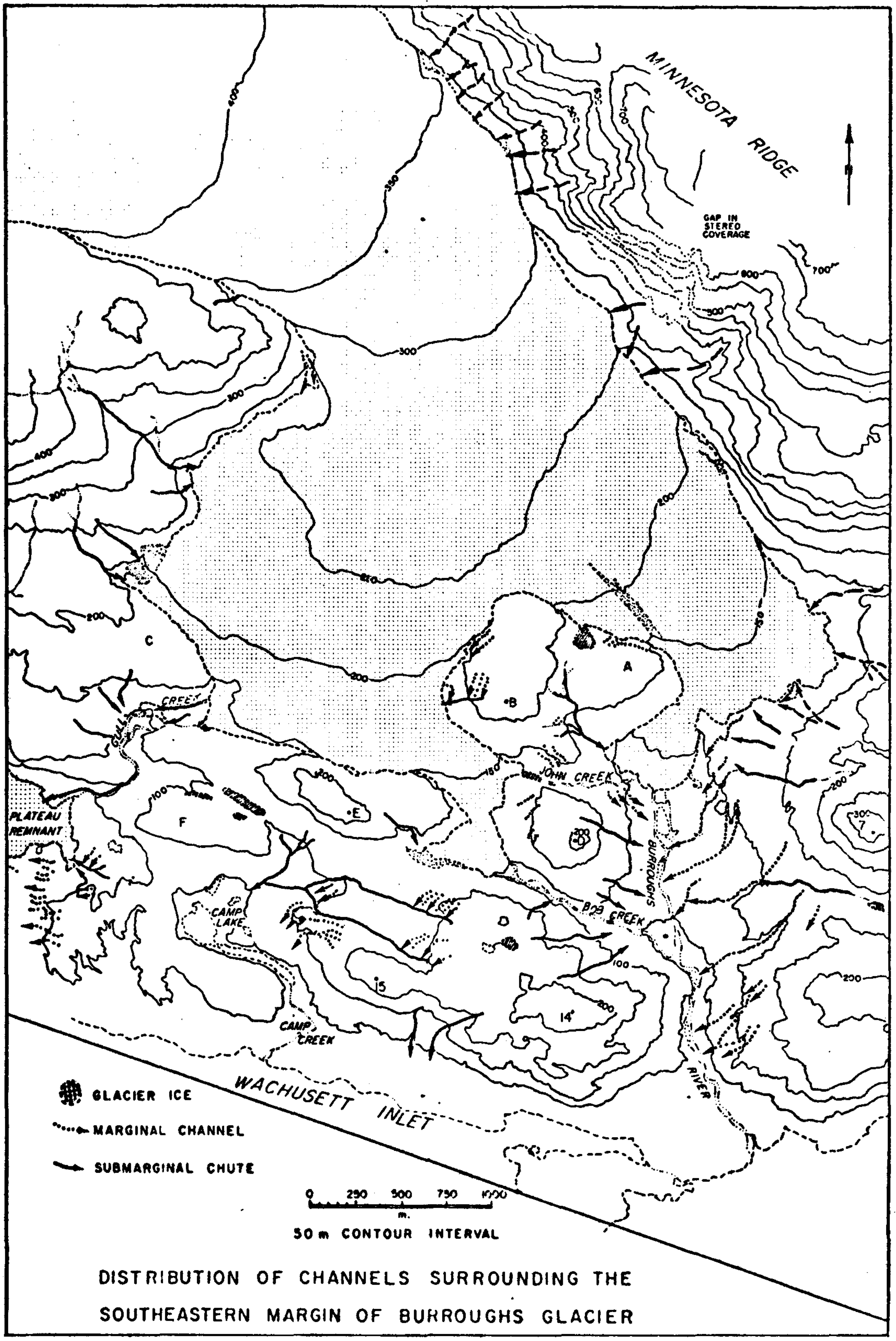


Figure 62. Distribution of ice-marginal channels and submarginal chutes near the southeast terminus of Burroughs Glacier.

Small Marginal Channels

Tarr (1908) described one-sided channels which were forming and had formed against the ice. After ice retreat, they appear as small cut or deposited terraces on the hillside (Mannerfelt, 1949). Since ice must have been present on their downslope sides, they have been considered marginal, although it is conceivable that they could also form submarginally. These channels commonly end by turning abruptly downhill into a submarginal chute. In Scandinavia they usually have an average gradient of 2-3 percent (Mannerfelt, 1960; Gjessing, 1960) and commonly occur in sets on hillsides.

One anastomosing set of these channels was observed forming during the summer of 1970 on the north side of Nunatak B. As ice retreated down a 10° slope at a rate of 0.8-1.0 m/week (slope distance) between June 15 and August 20, 2, 3, or 4 channels were cut to a depth of up to 2 m; some segments of these were left as one-sided, cut terraces (fig. 63). The stream, with a high discharge of 2 cfs occupied a channel only until the ice had retreated far enough to allow water to run along its margin. The channel segment that extended farthest under the ice (thus farthest down slope) had a smaller or non-existent downslope wall than the rest of the channel. As the ice withdrew from the channel, water cut through this wall (usually at high flow) and began to flow parallel to the ice margin until it again intersected a segment of the channel. Thus as the ice withdrew an anastomosing pattern of channels was formed.

Long profiles of the channel bottom and the ice margin along the channels were constructed for July 10 and August 17, 1970 (fig. 64). The gradient of the channel bottom reflects the gradient of the ice margin fairly closely in both cases except in the vicinity of the submarginal chute.

Three sets of marginal channels, which occur on the hillside above the previously described set, are shown in Figures 65 and 66. The channels vary from the one-sided cut terraces described above to distinct channels as much as 1 m deep. The gradients range from about 12 percent near the top to about 15 percent in the lower set, indicating a slight increase in gradient as the ice margin retreated.

The ice margins shown in Figure 66 are estimated from photographs taken in late summer of the years indicated. They indicate a more marked steepening of the margin between 1960 and 1963, but the highest channels were probably not formed until 1961 when some steepening of the margin had already occurred. Sets of small sub-parallel marginal channels are also present along the south side of Plateau Remnant and on the hillslope north of station 15. Both sets reflect the ice margin gradient, in general, but no photos are available to document their exact time of formation.

Figure 63. Photo of marginal channels on northwest side of Nunatak B. Channels are 1 to 2 m deep.

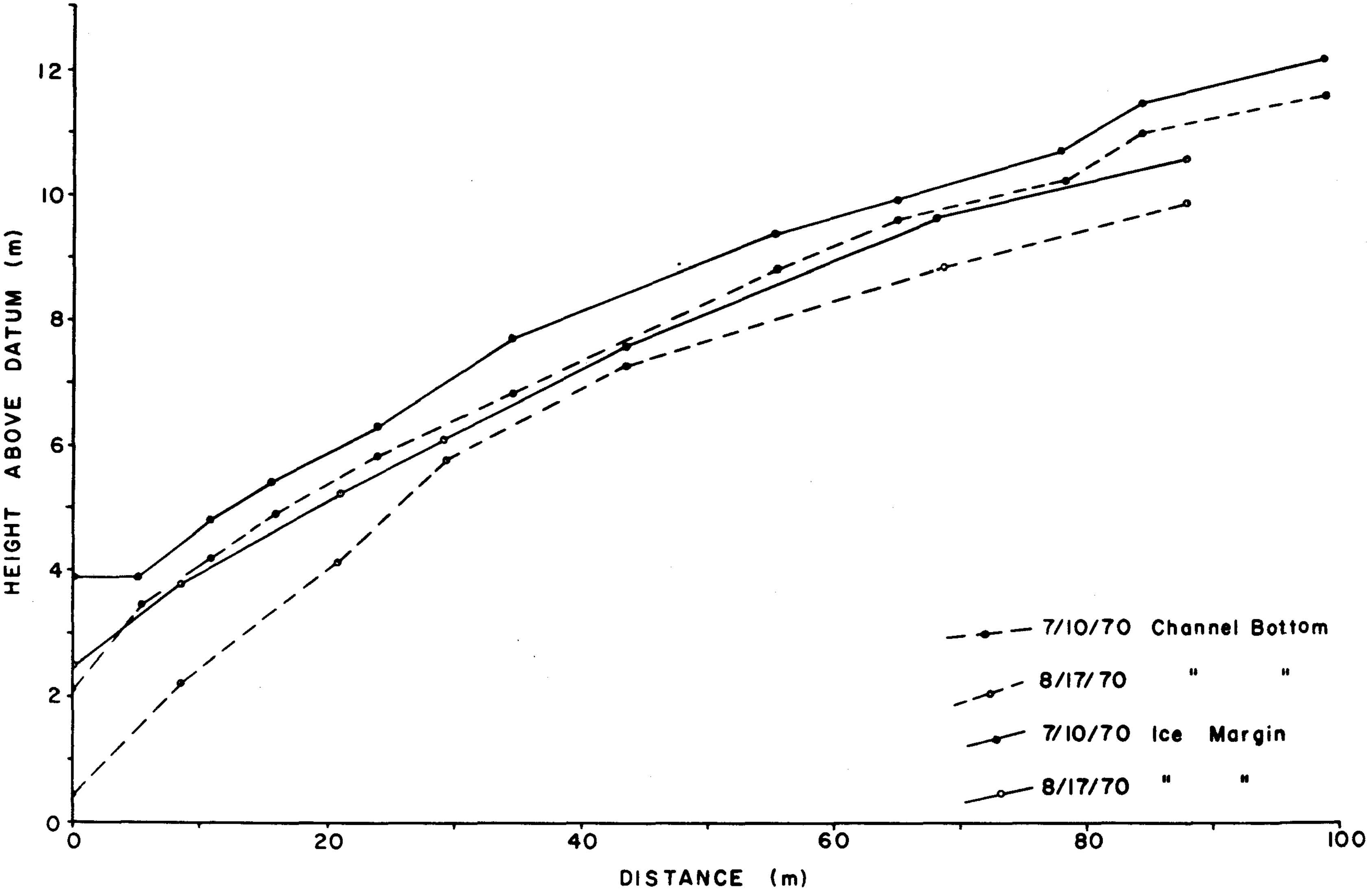


Figure 64. Relationship between ice margin and ice-marginal channel on two days in 1970.

Figure 65. Photo of sub-parallel marginal channels formed between 1960 and 1964 on Nunatak B.

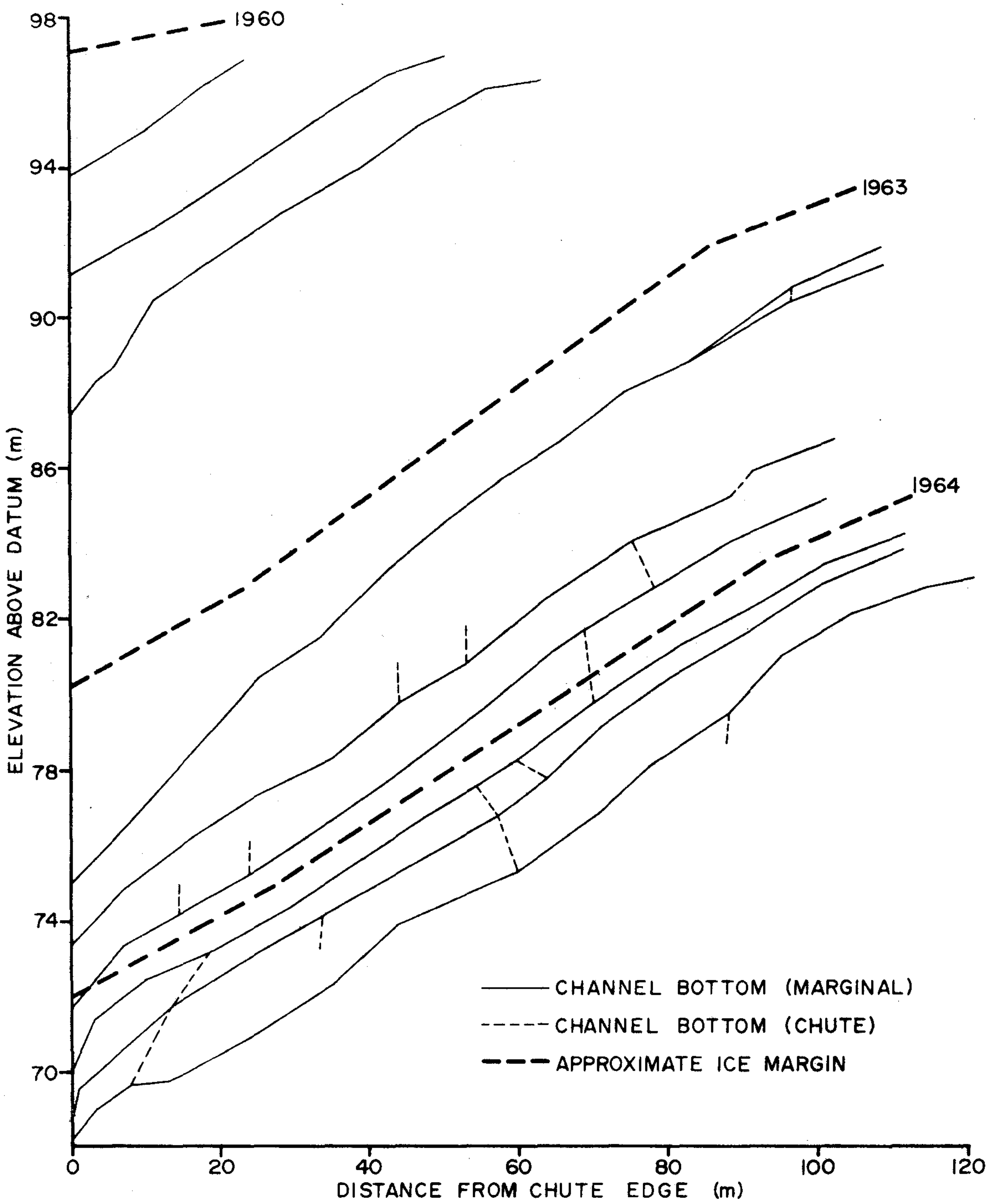


Figure 66. Longitudinal profiles of ice-marginal channels on Nunatak B. Ice margin positions estimated from aerial photos.

Large Marginal Channels

Deeper marginal or submarginal channels occur in the area between station 7 and Nunataks A, B and D (fig. 62). Although cutting of the channels was at first submarginal the channels are nearly parallel to former ice margins. The source of water was probably both meltwater from the margin and a subglacial or englacial tunnel carrying water from the direction of the crest of the Burroughs Glacier in a manner similar to the Burroughs River today. Discharge was probably on the order of 50 to 100 cfs. The ice margin gradient is best represented by the land slope along the edge of the channel rather than the gradient of the channel itself. Thus the position of the channel is controlled by the ice margin, but once cutting begins the gradient of the stream is independent of the gradient of the ice margin.

Another set of deep marginal channels (fig. 67) occurs north of Camp Lake. They seem to have begun cutting just under the ice margin as was the case of the channel cut in 1960. Again, the gradient and shape of the ice margin is best reflected by the gradient of the land surface along the edge of the channel rather than the gradient of the channel bottom (fig. 68, 69). Although times of occupation of each channel cannot be determined from photographs, all of the channels were probably cut between the mid-1950's and 1964. No marginal channels were formed by the superposition of supraglacial streams as described by Schytt (1956), Price (1960), Sissons (1958) and Clapperton (1968). This may be because there is little surface drainage on the ice at low elevations on the Burroughs Glacier due to the highly permeable surface ice. Some of the larger channels may have formed by superposition of an englacial stream, but this cannot be determined from photographs, and none were observed forming during the summers of 1969 and 1970.

In-and-Out Channels

Another type of marginal channel that can be used to obtain estimates of former ice-margin gradients is the in-and-out channel. First described by Smith (1932) in Great Britain, these crescent-shaped channels, incised on a hillside, hang at both ends. He suggested that the crescentic shape indicated a lobate margin with each channel being controlled by the shape of the ice margin. Common (1957), working in the Cheviot Hills of Scotland, suggested that this type of channel formed by the superposition of a meandering supraglacial stream on the underlying hillside. Although these hypotheses are not unreasonable, to the writer's knowledge neither of these mechanisms has previously been seen to produce an in-and-out channel.

A series of these channels was observed forming on the Burroughs Glacier during the summer of 1969 (fig. 70, 71). Three consecutively lower in-and-out channels in till were formed by one stream as the ice retreated down a 10° land slope at about 0.8-1.0 m/week (downslope

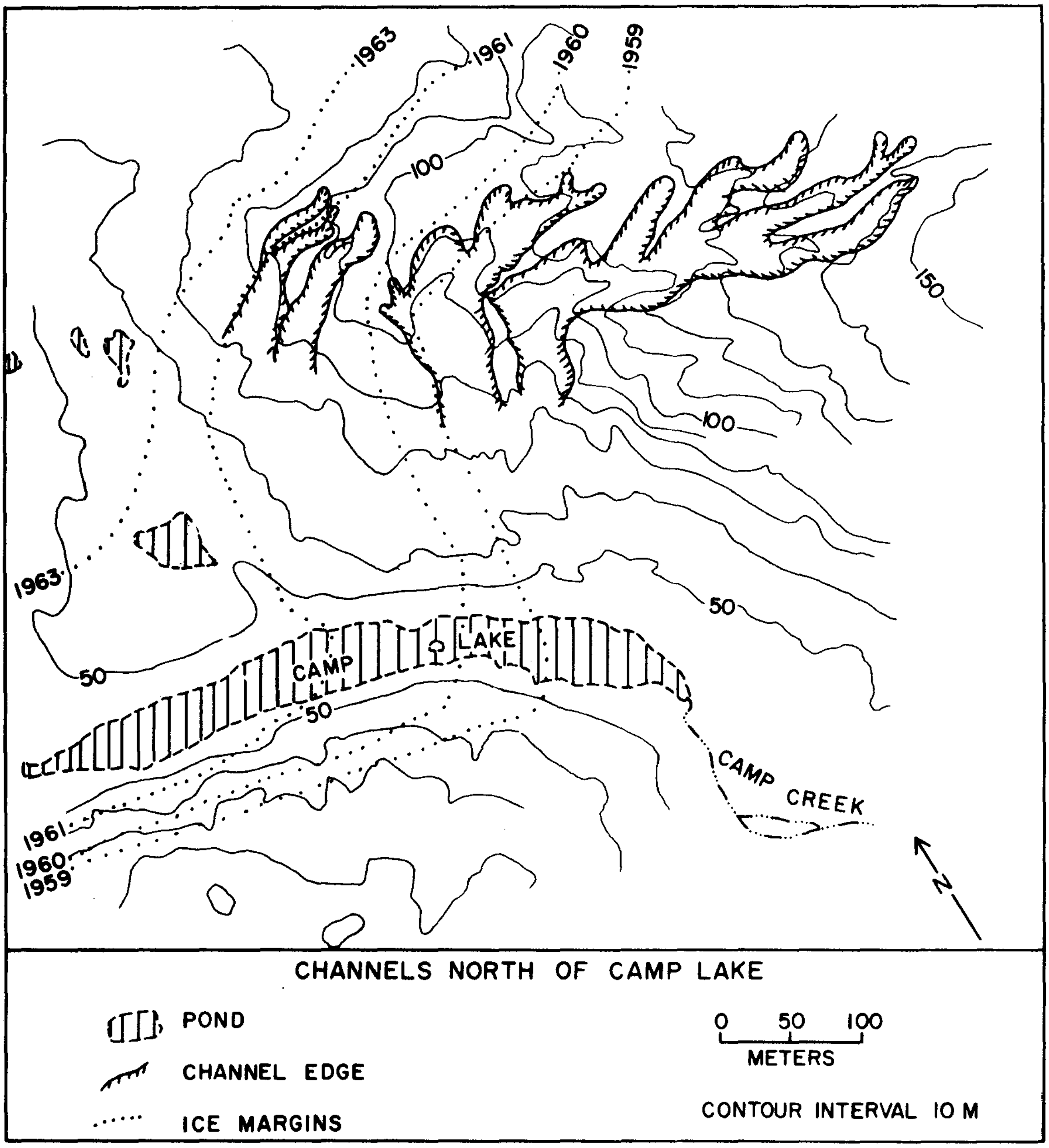


Figure 67. Channels north of Camp Lake and selected ice margin positions.

Figure 68. Photo of channels north of Camp Lake in 1959. Channels are up to 20 m deep at mouths. (Photo by L. Taylor)

Figure 69. Photo of channels north of Camp Lake in 1969. Note two lower channels which formed between 1959 and 1963 and the change in cross-sectional shape of the large channels.

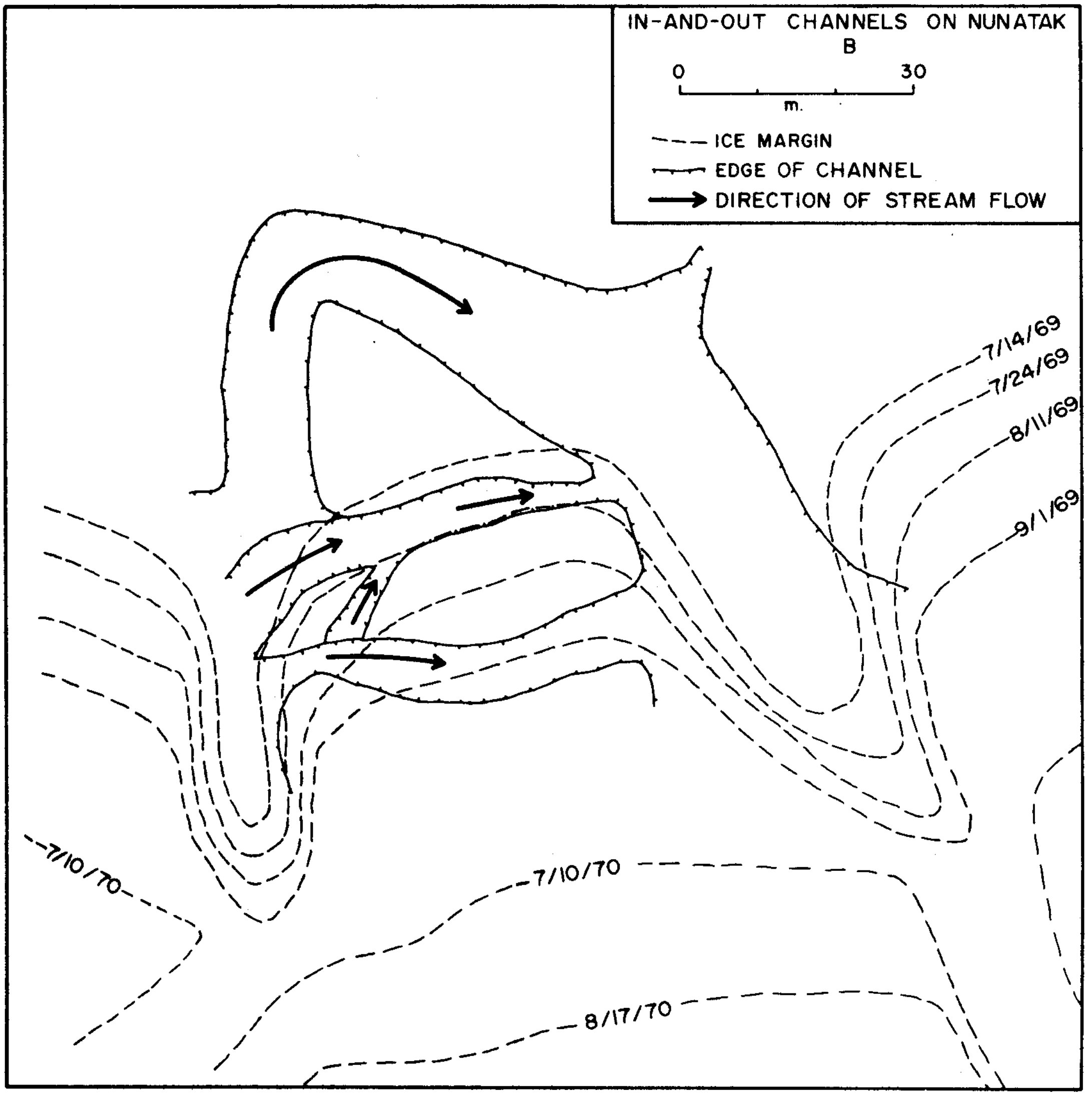


Figure 70. Map of in-and-out channels showing ice margin positions.

Figure 71. Photos of in-and-out channels formed between July 4, 1969 (above) and September 5, 1969 (below). Note figure in center of top photo.

distance). The ice-surface gradient perpendicular to the margin was approximately 20°. The upper channel was cut quickly in June to a depth of about 2.5 m until a lag concentrate of boulders and cobbles slowed further cutting. Ponding occurred at the head of the channel and a new channel formed along the ice margin about July 24. This sequence was repeated to form the lowest channel about September 1. The ponding of water that occurs in the late stages of channel development seems to be typical of most of the channels that head against or on ice. Figure 72 shows a sand and silt apron evidently deposited at a channel head during a late stage of channel development.

Marginal Channels and the Pattern of Ice Retreat

The overall distribution of channels reflects the nature of ice retreat as suggested by numerous studies of Pleistocene-age channels. For instance, Goldthwait (1968) found that in central New Hampshire, of 112 channels observed, 44 carried water toward directions between south and southeast and 23 between south and southwest, although a complete range of azimuths was found. He suggested that when the channels were formed, the ice had a gradient to the south but that stagnation and downwasting took place, exposing nunataks and producing a ragged ice mass. The number of marginal channels that carried water toward various octants in the area of Figure 62 are given in Table 11 with the percent of the ice margin with gradients in those directions on the 1960 topographic map (Taylor, 1962a, b).

Marginal Channels as Indicators of Rates of Retreat

Mannerfelt (1949) suggested that marginal channels could be considered annual and that a rate of ice retreat could be determined from their positions. He calculated that 3.5-5 m/yr of thinning had taken place in Scandinavia. This has been disputed by numerous writers and later discussions by Mannerfelt suggest that this interpretation may not be valid. Rate of formation of channels at the Burroughs Glacier varies from at least eight small, successively lower channels being formed in one year, to one large channel being occupied for several years. Channel spacing and depth are thus determined not only by the rate of downwasting but also by the slope, size of the stream cutting the channel, and the erodibility of the substrate.

Although the anastomosing channels on Nunatak B were only observed forming during one year, a possible way of determining annual sets of channels is suggested. Further observations on marginal channel gradients should be examined before this method is accepted. It can be seen (fig. 64) that the channel formed late in the 1970 season has a steeper gradient near the submarginal chute than that formed early in the season. One channel above this (fig. 66) also shows a steeper gradient over a longer distance than channels below it. This could have formed late in

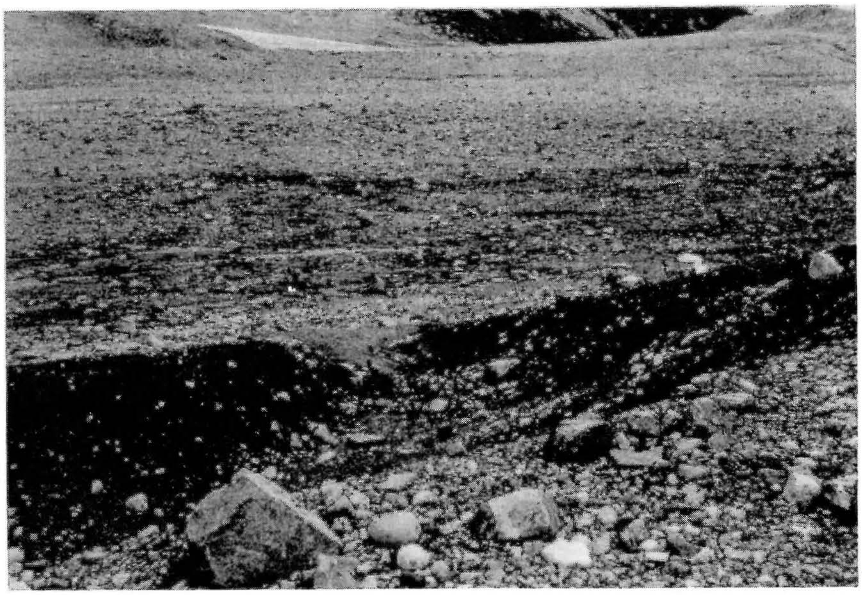


Figure 72. Photo of hanging channel head. Note sand apron at head of channel.

Table 11. Number of channels which carried water toward various octants and the percent of 1960 ice margin with gradient in those directions.

	Number of channels	% 1960 ice margin
N-NE	4	8
NE-E	5	1
E-SE	6	29
SE-S	12	10
S-SW	55	19
SW-W	59	22
W-NW	19	11
NW-N	0	0

1963 but this cannot be determined from photographs. Thus it seems possible that annual sets could be differentiated by this method if a complete set of channels is present.

Subglacial Channels

A number of subglacial channels are presently forming at Burroughs Glacier. Three types, submarginal channels, submarginal chutes and central channels are distinguished, although gradations exist between all three. Submarginal channels, those formed nearly parallel and close to the ice margin, were discussed with marginal channels as they can be used as indicators of ice margin gradients. Submarginal chutes (subglacial chutes of Mannerfelt, 1945) carry water from the ice margin directly under the ice. These are commonly associated with marginal channels (fig. 65). The third type, central subglacial channels, carry water under the main part of the ice, usually in the direction of the ice surface gradient.

Submarginal Chutes

Submarginal chutes occur at the end of marginal channels on the northwest side of Nunatak B (fig. 65) and several other localities, on the lee sides of nunataks and along the Bruce Hills and Minnesota Ridge (fig. 62). They indicate little about ice margin positions because parts of the channel seem to be occupied for a relatively long period of time during ice retreat. Those filled by eskers can yield information about the ice-surface gradient as described in Chapter IV.

On the Burroughs Glacier the chutes seem to be controlled by crevasses and fractures in the ice. This suggests that they form when water, flowing along the ice margin, follows a fracture or crevasse into the ice, and an englacial or subglacial tunnel forms by erosion of the channel floor (if subglacial) or by melting of ice along the tunnel walls as suggested by Mathews (1969), Stenborg (1969a, b), and Rothlisberger (1969). A description of water movement in englacial tubes on part of the Burroughs Glacier is given by Larson (in preparation).

On both sides of the high part of the Burroughs Glacier and along the south side of the Bruce Hills, water flows down chutes and beneath the ice margin (fig. 48). This water then evidently flows in englacial tubes to the lower part of the ice. As suggested by Stenborg (1969a, b) for Mikkaglaciären and Storglaciären in Sweden, tensile crevasses probably play an important role in determining the path of this water after it enters the ice. The absence of marginal channels, high on either side of the Bruce Hills or on the south side of Minnesota Ridge, is probably due to the presence of the crevasses. Here water, which would flow along the margin on the lower part of the glacier where most crevasses are closed, flows into crevasses without forming marginal channels.

Striations

Striations, grooves, and roches moutonnées are present on all nunatak tops within the study area. The orientation of striations is shown in Figure 19. In many cases crossing sets of striations are present, indicating the change in ice-flow direction that took place after 1907. Although a large number of grooves and roches moutonnées measurements were not made, most are parallel to the pre-1907 ice-flow direction. Thus striations formed during the change in ice-flow direction are oblique to the roches moutonnées on which they formed.

On the top of Nunatak A, striations have azimuths ranging from 300° to 330° (fig. 73), suggesting that the striations represent both the early and later ice-flow directions. In 1948 the ice flow across Nunatak A was from about 330° (plate V) and in 1920 from about 315°. By 1948 the ice was only 55 m thick and the basal shear stress is estimated at 0.56 bar (table 6).

On Nunataks B, C, E, and F, as well as other bedrock highs, striations show a similar or greater range of azimuths. On the island in Wachusett Inlet near the mouth of Burroughs River, only one set (275-285°) is present. This is consistent with ice flowing parallel to the inlet.

Figure 73. Photo of striations on top of Nunatak A. Azimuths range from 300° to 330°.

CHAPTER VII

GEOMORPHIC CHANGE IMMEDIATELY AFTER ICE RETREAT

After deglaciation a landscape is modified by runoff from heavy rain, snow melt, and slow dewatering of very wet till and outwash. This produces a more subdued microrelief and a surface with a high density of washed cobbles and boulders. An example of this can be seen in Figures 74 and 75. A crag and tail feature, photographed in 1959 just after it emerged from the ice, is fairly sharp-crested and surface cobbles are veneered with till. By 1970 the feature was more subdued and the surface cobbles had been washed. Similar surfaces deglaciated in 1969 had been washed by the beginning of the 1970 summer.

Slope Modification

Gully Erosion

Closely spaced subparallel gullies occur on steep till slopes (25-35°), and are well developed north of station 15 and on Nunatak D. The V-shaped gullies are as much as 1.5 m deep (fig. 76). Most are extensions of small gullies on the gentle slopes above them.

Although none were observed near the present ice margin, aerial photographs suggest that gullies form during or immediately after deglaciation. In 1964 gullies led obliquely to the ice margin and may have been submarginal chutes. Since neither the steep gullies nor the small gullies which lead to them are parallel to former ice margins, it is unlikely that they were cut by meltwater.

Since till at the ice margin is very wet, it seems likely that spring sapping may be in part responsible for the formation of the gullies. Those deglaciated for at least 5 years now carry water only during and shortly after rain, suggesting that surface runoff is responsible for their maintenance.

Ice-Marginal Mudflows

Flows of till both on and off the ice are common around the margin of the Burroughs Glacier. One type is of particular interest because the mud flows along the ice margin. After deglaciation a small terrace is left whose gradient reflects that of the ice margin. These features are thus useful as indicators of ice-margin gradients.

The terraces are best developed on the south side of the till build-up west of Nunatak D (fig. 77). Flat-bottomed mudflow scars

Figure 74. Photo of crag and tail feature (same as Figure 75) near ice margin in 1959.

Figure 75. Photo of crag and tail feature (same as Figure 74) in 1970.

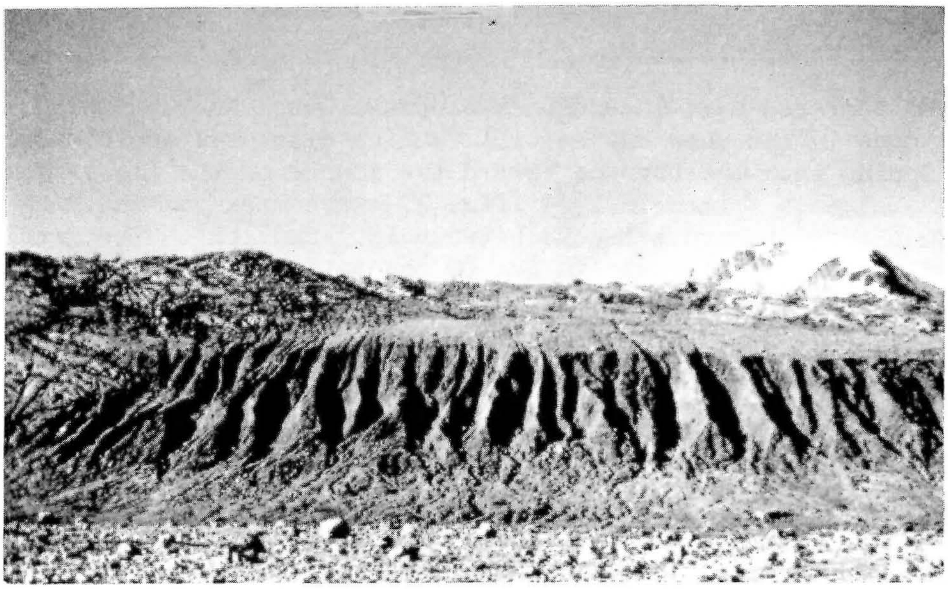


Figure 76. Photo of gullies north of station 15. Large boulder in left foreground is 1 m in diameter.

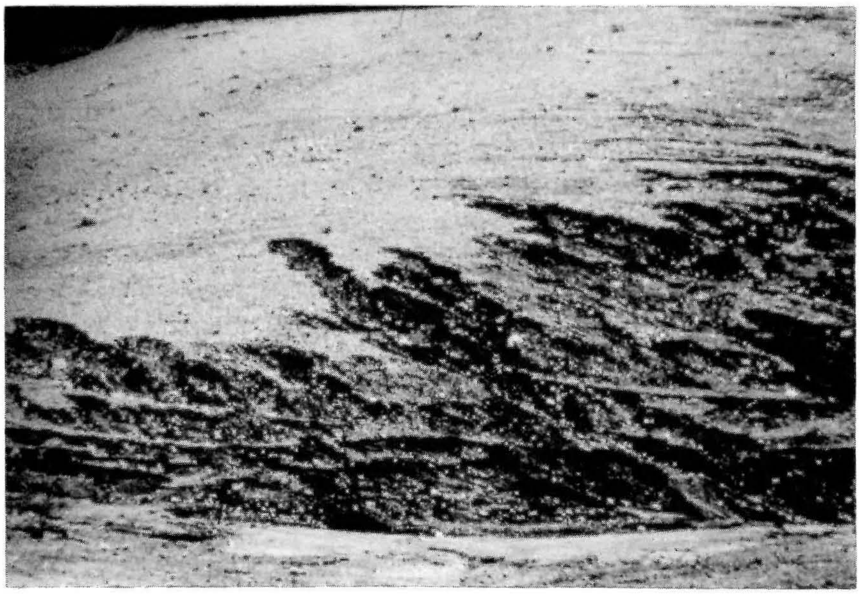


Figure 77. Photo of ice-marginal mudflows near Nunatak D. (Ice margin in foreground).

about 1.5 m deep lead to the terraces. The scars have a nearly vertical head and lack the cobble floor of otherwise similar open-ended marginal channels (e.g. fig. 71).

The terraces have ice-contact slopes as much as 0.5 m high. A fabric done in the same way as till fabrics discussed above showed a mode dipping into the terrace toward the source of the flow. The slopes of four terraces formed in 1969 (fig. 77) were measured and range from 3 to 16 percent toward azimuths between 150° and 182°. The gradient of the mid-summer 1970 ice margin was 60 percent toward an azimuth of 160°. A group of them, therefore, provide a fairly good estimate of the ice-margin gradient.

The flows seem to be triggered by rainstorms. Between the night of August 13 and the morning of August 15, 1969, 36 mm of rain fell. A large flow, then about 15 m from the ice margin (left center, fig. 77), occurred during that time.

Change in Channel Morphology

The cross-sectional shape of meltwater channels has been used as a criterion for their recognition by Derbyshire (1958) and others. Broad, flat floors and steep walls are cited as typical, although as noted by Derbyshire, they may be modified after deglaciation. The channels being cut in till near the ice margin of Burroughs Glacier have very steep walls and flat floors but the shapes change rapidly when the channels are no longer occupied. Figures 68 and 69 show the same channels in 1960, just after they were occupied, and in 1970. During the 10 years, the channels developed an asymmetrical V-shape because of mass movement down their walls. At the channel heads a marked change in the cross profile has taken place between 1958 and 1970.

In a shorter period of time similar changes have taken place in the in-and-out channels on the north flank of Nunatak B. A hand-leveled profile (fig. 78) shows the development of a more V-shaped cross section in one of the channels between July 1969 and August 1970. It is also significant that the long profile (fig. 78) has developed the up-and-down profile typical of subglacial channels. Thus measurement of the gradient of abandoned channels from topographic maps can lead to erroneous results. The original cobble stream bottom is preserved beneath the slumped till; field identification of this surface should be made before channel profiles are measured.

Compaction Mounds

Near the 1969 margin of Plateau Remnant a thin veneer (up to 0.25 m) of poorly stratified silty sand and fine gravel was evidently deposited in ponded water. In 1970, mounds as much as 0.5 m high and 1.5 m in diameter had formed (fig. 79) in an area 15 by 20 m. Each mound is over a boulder covered by both till and sand (fig. 80).

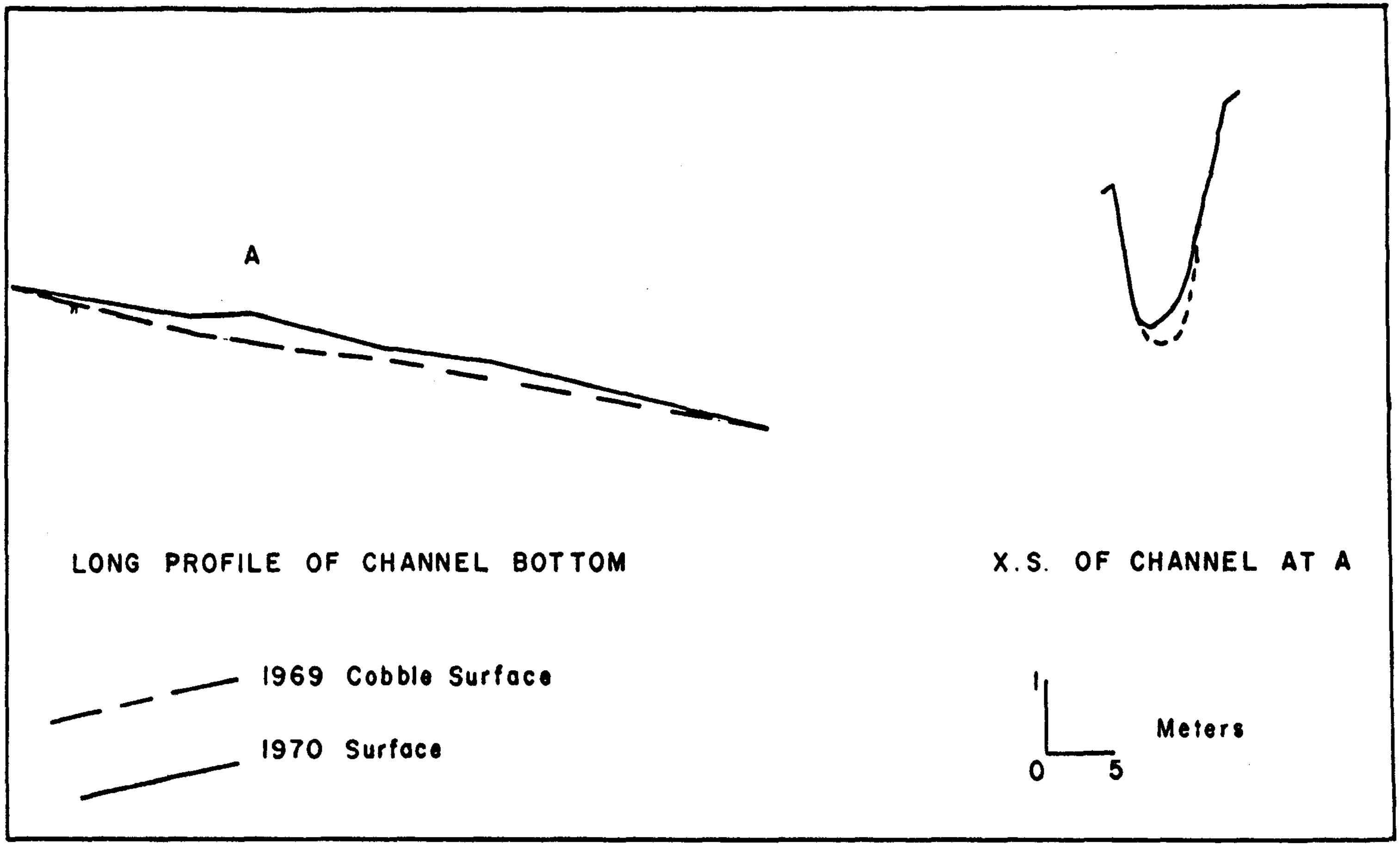


Figure 78. Longitudinal profile and cross section of channel in 1969 and 1970.

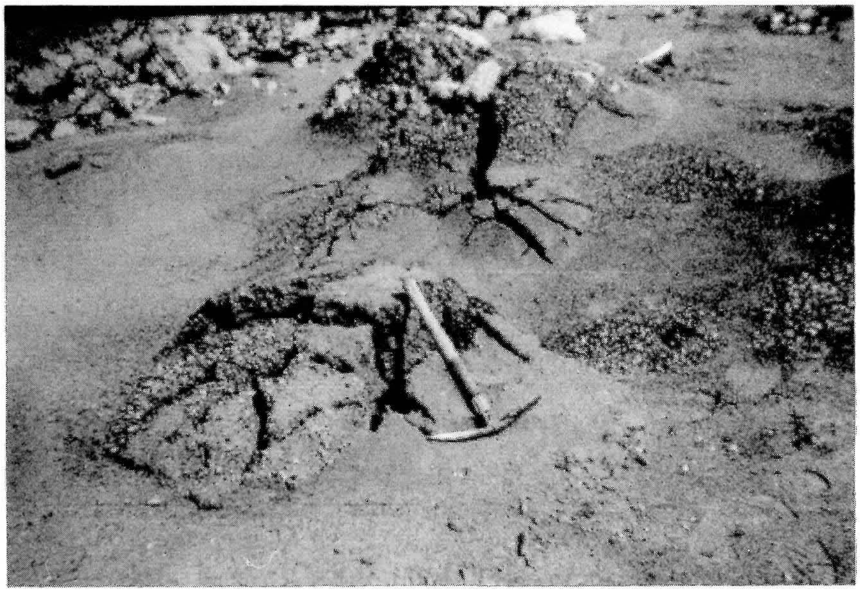


Figure 79. Photo of compaction mounds near Plateau Remnant.

Figure 80. Photo of internal section of compaction mound. Note boulder covered by till which is covered by sand and silt. Notebook is 12 by 20 cm.

Similar features, formed when sand was deposited on snow which subsequently melted, have been observed by R. Goldthwait (personal communication). However, since these boulders are actually in the till and the till surface rises over the boulders this mode of formation seems unlikely in this case. The ice in this area was free of ablation till during the summer of 1969 so it is unlikely that the till over the boulder is ablation till.

The mounds must indicate differential compaction of the sediment after surface water drained from the area in late 1969 or early 1970. Some compaction of the till may have occurred due to loading by snow or ice during the winter. Because till at the edge of the ice is typically saturated or supersaturated, compaction as water drained out of the silty sand and the till seems likely. The water-table surface is approximately 0.5 m below the top of the mound shown in Figure 80; thus, some dewatering has taken place.

SUMMARY

Chronological Summary of Events

1890-1910

In 1890 the tops of Idaho Ridge, Minnesota Ridge and peak 1960 of the Bruce Hills were ice-free. Patchy, hummocky sand and gravel were deposited high on Minnesota Ridge. The ice-flow direction at this time was about 300° in the southeastern part of Burroughs and Plateau Glaciers. Roches moutonnées, grooves, and striations with this orientation were being cut at this time. Till was being deposited in areas such as Adams Inlet to the southeast, and on the stoss sides of hills beneath the ice, such as Nunataks A, B and D.

Throughout this period, ice flowed through Glacier Pass from Muir Glacier to Burroughs Glacier. Striations, crag-and-tail features, and marginal channels indicate that ice continued to flow from Glacier Pass toward Burroughs Glacier until at least 1948.

1910-1930

The Bruce Hills emerged and became a controlling factor in determining ice-flow directions at the southeast end of Burroughs Glacier. Basal till was deposited at the rate of approximately 2 cm/yr in many areas although erosion still occurred on bedrock highs and on the lee side of Nunatak D. Marginal channels had begun to form near Glacier Pass by 1930 with their orientations demonstrating that at this time the ice-surface gradient was toward Cushing Glacier.

1930-1950

Soon after 1930 the col between peaks 1960 and 2221 of the Bruce Hills became ice-free. Water flowed across the col from Burroughs Glacier to Plateau Glacier and disappeared beneath the ice where small gravel terraces were formed. Plateau Glacier was lower than Burroughs Glacier partly because of higher ice loss due to calving at its margin. During this time the surface of Muir Glacier also became lower than that of the Burroughs.

Ice-flow direction continued to change in the southeastern part of Burroughs Glacier. Basal till was deposited in most areas except near the tops of bedrock highs. Till crag and tail features were formed near the ice margin both north and south of Wachusett Inlet. Throughout this period striations were cut on ice-covered nunatak tops as the ice thinned.

With the retreat of the Plateau Glacier margin, arc-shaped crevasses formed in the ice over Wachusett Inlet and its margins. Subsequently till was squeezed into some of these crevasses near the margin. This presumably has occurred progressively with margin retreat until the present time.

Marginal channels continued to be cut near Glacier Pass and indicate an ice margin that decreased in elevation from Glacier Pass toward the northwest. Crag and tail features were deposited here indicating ice flow also to the northwest.

As the southeast margin retreated, ice-marginal lakes formed north and south of station 7. Here as much as 2 m of sand was deposited in locations where streams entered the lakes. A now nearly indistinguishable shoreline formed at the highest level of the lake south of station 7. The lake level dropped when ice retreated from a lower outlet and the lake remained at this second level until about 1950 when it drained submarginally. Deposition of outwash in Burroughs River began shortly after 1941. A steep outwash delta was deposited at the edge of Wachusett Inlet.

Kame terrace segments were deposited along the ice margin against Minnesota Ridge. Gradients of the kame terraces are nearly parallel to that of the ice margin. A blanket of hummocky sand and gravel was deposited at the east end of the Bruce Hills.

1950-1970

In the past 20 years Burroughs Glacier became separated from Cushing and Plateau Glaciers. At its northwest terminus, outwash was, and is presently, deposited by streams flowing from the Burroughs Glacier. The Plateau Glacier has retreated rapidly and nunataks have emerged near the southeast terminus of Burroughs Glacier.

Because of rapid deglaciation of Wachusett Inlet, ice from the Burroughs Glacier flowed directly toward the inlet and perpendicular to the initial flow direction. The fabric of till that had been squeezed into crevasses was reoriented as crevasses closed and rotated while surface movement was on the order of 2 to 3 m/yr. Crag and tail features were evidently formed during this time, close to the margin, as they are oriented parallel to the last ice-movement direction. A number of features formed during shorter periods of time are discussed below.

1950-1960

With the emergence of Nunataks A, B, D, E, and F shortly after 1960, nearly stagnant masses of ice became stranded on their lee sides. Channels were cut by water flowing from the high ice on the stoss sides of the nunataks, across the nunataks and into the ice on the lee sides. These streams deposited what are now collapsed kame terraces that appear as blankets of sand. A system of small gravel eskers was deposited on the lee side of Nunatak D, presumably by the subglacial Burroughs River.

By 1956 the two ice-dammed lakes adjoining station 7 had drained submarginally. Muir Lake formed at approximately the same time when the ice margin retreated over a drainage divide. By 1960 the lake seems to have drained subglacially.

The first of the large marginal channels carrying water from near station 7 into the Burroughs River began as a submarginal channel about 1956. This subparallel set of channels remained active until 1964. The gradient of the ice margin adjacent to the channels during this time is recorded not in the gradient of the channel bottoms, but in the land surface gradient along the edges of the channels. Four large marginal channels were cut across the ridge north of Camp Lake beginning in about 1958. Each channel was occupied approximately 2 years and their gradients indicate a steep ice-surface gradient toward Wachusett Inlet. Outwash at the base of these channels was deposited on ice that subsequently melted.

By 1960 the Burroughs River had downcut its outwash 15 m from the 1964 level at the position of the 1948 ice margin. The delta had prograded about 50 m from the 1948 position.

1960-1970

By 1960 ice at the southeast end of Burroughs Glacier was confined to the valley bottoms. Eskers were formed under Plateau Remnant south of Nunatak F. One of these, composed primarily of silt at its downstream end, formed between 1964 and 1967. Its flow direction indicates that the ice surface gradient was toward Wachusett Inlet even after flow from the Burroughs Glacier had stopped and only a nearly stagnant ice mass remained.

Between 1962 and late 1964, small marginal channels were cut in till on the northwest side of Nunatak D indicating a steepening ice margin that is confirmed by the photographic record. A kame terrace formed in the lee of Nunatak D between 1960 and 1963 just beneath the ice margin. Another kame terrace was deposited in ice near the margin at the lee side of Nunatak A between 1967 and 1969 by a stream flowing from the high ice on the lee side of the nunatak. Much of this will probably collapse as buried ice continues to melt.

The last connection between Burroughs and Plateau Glaciers was in the valley of Gull Stream. Kame terraces along the valley walls have a gradient toward Wachusett Inlet. Several eskers were formed in the valley bottom and in places were not exposed until 1970. Since Gull

Stream flowed in and on Plateau Remnant as late as 1970, outwash there is ice-cored and will collapse in the future.

As the ice margin retreated Burroughs River downcut 8 m at the 1948 margin position and 12 m at the 1960 margin position between 1960 and 1970. The delta prograded about 35 m during that time.

Downwasting vs. Backwasting

As Johnson (1941) pointed out, margin retreat is a necessary result of downwasting. In the Burroughs Glacier area, where there is a decrease in the rate of ice-surface lowering with increasing elevation, the importance of backwasting is much greater than downwasting in the disappearance of a single ice mass. Three other factors, the presence of calving margins, the flow of ice from the northeastern terminus of the Burroughs Glacier toward Queen Inlet, and the emergence of hilltops have caused the separation of the Burroughs Glacier. Thus on a larger scale relative stagnation of separated ice masses has taken place.

Johnson (1941, p. 94) argued that "to recognize with certainty the loci of such stagnant masses... appears in the light of our present knowledge a well-nigh insoluble problem." The presence of marginal lakes, marginal channels, and other features at the southeast margin of Burroughs Glacier shows only the effects of local margin retreat. Marginal channels, crag and tail features, and till buildups on the stoss sides of hills and perhaps outwash at the northwest terminus, would argue for the isolation of the Burroughs Glacier. Thus these features will demonstrate the position of a nearly stagnant, separated ice mass even after complete deglaciation. Likewise, similar features in other areas should aid in the interpretation of the history of deglaciation.

APPENDIX A

MECHANICAL ANALYSES AND PERCENT METASEDIMENT FRAGMENTS IN HEAVY FRACTION OF SAND

	% of Total Sample	% of < 2μ Fraction (Silt and clay not distinguished in gravel samples)			
Sample Number	(< 2 mm)	Sand (2-0.0625 mm)	Silt (0.0625-0.002 mm)	Clay (< 0.002 mm)	% Metasediment Fragments
Granite Canyon Till					
121	72	49	40	11	26
126	81	48	42	10	36
Average	77	49	41	11	-
Glacier Bay Formation					
7	43	90	8	2	-
8	66	55	38	7	16
9	52	58	27	13	-
15	65	61	29	10	-
16	56	61	33	6	-
35	48	66	31	4	-
37	70	58	35	7	-
38	75	55	37	8	-
39	65	65	27	9	28
40	65	58	34	9	25
41	72	62	30	8	29
42	42	63	29	8	-
46	88	41	51	9	40
48	74	46	45	9	38
50	74	56	39	5	22
51	61	57	37	6	33
62	63	81	16	3	26
63	76	63	34	3	25
101	58	63	31	6	-
102	40	73	23	5	-
103	44	71	25	4	-
106	54	52	42	7	-
107	87	52	42	6	13

APPENDIX A (Continued)

	% of Total Sample	% of $< 2\mu$ Fraction (Silt and clay not distinguished in gravel samples)			
Sample Number	(< 2 mm)	Sand (2-0.0625 mm)	Silt (0.0625-0.002 mm)	Clay (< 0.002 mm)	% Metasediment Fragments
Glacier Bay Formation (continued)					
108	61	55	39	7	20
120	61	55	38	7	49
122	83	40	50	10	41
123	70	40	45	15	46
124	83	46	44	9	33
127	81	45	46	9	32
128	89	32	62	6	35
132	70	60	34	6	40
140	67	57	35	8	25
141	28	48	43	9	24
142	54	67	28	6	20
143	68	60	34	6	44
144	70	53	43	5	29
146	68	55	36	8	-
155	65	61	31	8	-
Average	68	57	36	7	-
Ablation Till					
105	51	66	30	4	-
145	54	69	28	4	-
147	58	44	46	10	-
148	71	83	13	3	-
149	72	61	37	2	-
150	76	67	28	5	-
154	61	73	25	3	-
Average	63	66	30	4	-

APPENDIX A (Continued)

	% of Total Sample	% of < 2μ Fraction (Silt and clay not distinguished in gravel samples)			
Sample Number	(< 2 mm)	Sand (2-0.0625 mm)	Silt (0.0625-0.002 mm)	Clay (< 0.002 mm)	% Metasediment Fragments
Lake Sediment					
125	100	0	95	5	-
133	100	2	92	6	-
134	100	30	68	3	47
135	100	1	93	6	-
151	100	2	88	10	-
152	100	2	92	6	-
Average	100	6	88	7	-
Eskers					
12	100	94		6	-
13	39	89		12	-
14	37	94		6	-
19	46	89		11	18
22	51	97		3	-
23	19	95		5	-
24	88	95		5	-
27	100	58		42	24
28	100	90		10	-
29	100	94		6	-
30*	100	5	48	2	-
31*	100	2	94	2	-
32*	100	51	31	1	-
33*	100	1	31	5	-
Average	68	90		11	-
* not considered in average					
Kame Terraces					
17	44	94		6	28
18	41	93		7	-

APPENDIX A (Continued)

	% of Total Sample	% of < 2μ Fraction (Silt and clay not distinguished in gravel samples)			
Sample Number	(< 2 mm)	Sand (2-0.0625 mm)	Silt (0.0625-0.002 mm)	Clay (< 0.002 mm)	% Metasediment Fragments
Kame Terraces (continued)					
113	44	93		7	8
114	89	98		2	12
117	33	81		19	24
118	63	94		6	30
129	38	86		14	46
130	52	96		4	52
137	95	65		35	25
Average	53	90		11	-
Outwash					
10	43	98		2	-
116	56	95		5	35
138	35	96		4	18
139	69	97		3	35
Average	51	97		4	-
Middle Van Horn Formation					
43	100	15	80	5	-
44	100	2	93	5	-
Average	100	9	87	5	-
Upper Van Horn Formation					
111	24	99		1	8
Other					
20	100	87		23	-
21	41	97		3	-
110	99	79		21	36
113	44	93		7	8

APPENDIX B

RELATIVE ABUNDANCE OF CLAY MINERALS AND QUARTZ IN < 2μ FRACTION (ROUNDED TO NEAREST 5%)

Sample Number	% Illite	% Vermiculite	% Quartz	% Chlorite	% Expandable
Granite Canyon Till					
121	T	20	45	35	T
126	15	20	15	40	5
Average	8	20	30	38	3
Glacier Bay Till					
39	20	15	10	50	T
40	25	T	10	55	5
41	15	10	15	55	5
42	25	5	5	60	5
46	25	10	20	35	10
48	T	T	40	60	T
50	15	15	30	35	5
51	10	T	5	75	10
62	T	20	45	35	T
63	T	5	50	40	5
122	10	35	15	35	T
123	T	T	10	65	20
124	T	40	25	35	T
127	5	T	10	80	T
128	10	T	10	60	20
145	15	20	30	35	T
146	T	T	20	70	5
Average	11	11	21	51	6
Other					
31	10	T	25	65	T
32	5	10	45	30	10
43	30	T	10	55	5
54	15	10	30	45	T
58	5	T	15	65	10
59	10	10	40	25	10
60	15	T	10	70	5

T = Trace (considered 1% when calculating averages)

APPENDIX C

PEBBLE COUNTS

Rounding Sample Number	Sediment Sample Number	% Granite	% Syenite	% Diorite	% Basalt	% Andesite-Dacite Porphyry	% Marble	% Metasiltstone	% Other	Plutonic/ Metasiltstone Ratio	Rounding (Krumbein Scale)	% Pebbles with Fresh Breaks	% Pebbles with Striations
Granite Canyon Till													
123	121	20	0	4	24	29	0	23	0	1.04	0.33	22	6
Glacier Bay Till													
105	146	13	1	12	26	22	0	23	3	1.13	0.42	32	18
110*	-	15	23	13	17	28	0	0	4	-	0.46	31	11
111*	-	14	34	16	14	24	0	2	6	27.0	0.42	37	17
114	-	7	0	23	30	10	0	25	5	1.20	0.26	13	4
115	110	5	0	19	14	18	0	25	21	1.04	0.32	16	8
121	-	12	1	1	31	10	0	45	0	1.04	0.31	74	5
122	120	8	0	4	26	10	0	51	1	0.24	0.24	14	9
124	122	9	1	6	22	14	0	42	4	0.38	0.31	15	11
125	123	8	0	11	21	18	0	42	0	0.45	0.29	14	8
126	124	12	3	4	18	26	0	35	2	0.54	0.32	19	7
130	132	5	1	8	19	16	2	48	1	0.29	0.27	24	2
132	142	16	2	8	24	29	0	19	0	1.37	0.37	12	10
138	-	8	0	6	23	26	0	34	3	0.41	0.28	8	7
141	143	12	0	10	18	10	1	49	0	0.45	0.31	7	10
142	8	21	9	25	9	21	1	10	4	5.50	0.31	10	3
144	62	16	1	13	28	10	0	31	1	0.97	0.32	14	12
161	155	10	0	9	23	8	0	46	4	0.41	0.30	15	11
Average		10.8	1.3	10.6	22.1	16.5	0.27	35	3.3	1.03	9.3;	15.1	8.3
Ablation Till													
103	150	18	1	25	19	17	1	16	3	2.75	0.29	23	4
104	145	34	3	21	18	13	1	7	3	8.29	0.34	32	7
106	154	32	4	22	13	15	5	4	4	14.50	0.36	23	1
107	-	20	2	13	28	20	2	11	4	3.18	0.32	21	2
108	-	30	3	14	24	19	0	7	3	6.71	0.33	26	1
112*	113	24	47	7	5	13	0	1	3	78.00	0.34	12	1
148	-	20	3	18	13	23	1	20	2	2.05	0.32	17	9
Average		25.7	2.7	18.8	19.2	17.8	1.7	10.8	3.2	6.25	0.35	23.7	4.0
147	-	2	0	1	14	4	6	59	13	0.05	0.14	0	0
160	-		0	1	7	2	71	18	1	0.06	0.19	7	0
Average							38.5	38.5	7.0	0.06	0.17	3.5	0

APPENDIX C (Continued)

Rounding Sample Number	Sediment Sample Number	% Granite	% Syenite	% Diorite	% Basalt	% Andesite-Dacite Porphyry	% Marble	% Metasiltstone	% Other	Plutonic/ Metasiltstone Ratio	Rounding (Krumbein Scale)	% Pebbles with Fresh Breaks	% Pebbles with Striations
Upper Member-Van Horn Formation													
109*	111	15	19	15	13	30	0	1	7	49.00	0.52	25	0
146	-	21	0	16	24	26	0	13	0	2.8	0.47	13	0
Eskers													
113*	-	12	25	12	22	24	0	2	3	24.50	0.46	35	3
135	19	19	2	8	22	20	4	22	2	1.62	0.41	12	0
136	-	23	0	15	12	24	0	23	3	1.65	0.42	12	1
137	27	24	1	13	19	25	0	17	1	2.24	0.44	11	0
140	-	18	1	15	15	29	0	21	1	1.62	0.39	11	1
Average		21.0	1.0	12.8	17.0	24.5	1.0	20.8	1.8	1.78	0.41	11.5	0.5
Kame Terraces													
116	114	15	1	23	27	13	1	17	3	2.17	0.29	16	2
117	-	6	0	5	18	9	0	56	6	0.20	0.24	10	1
119	117	17	0	11	17	21	0	34	0	0.82	0.34	10	0
120	118	17	0	5	24	11	0	42	1	0.51	0.30	8	0
127	129	9	1	1	16	18	1	52	2	0.21	0.24	15	2
128	130	9	1	3	22	13	0	51	1	0.25	0.25	12	0
129	131	7	1	6	28	11	0	41	5	0.34	0.30	14	1
131	134	21	1	6	17	21	0	34	0	0.82	0.31	10	2
133	137	28	2	5	15	15	1	24	10	1.40	0.41	11	4
139	17	4	0	3	23	17	5	45	3	0.14	0.25	10	2
Average		13.3	0.7	6.8	20.7	14.9	0.8	39.6	3.1	0.58	0.29	11.6	1.4
Outwash													
118	116	16	0	12	19	7	0	46	0	0.61	0.28	8	1
134	138	15	3	13	22	20	0	24	3	1.29	0.43	9	0
143	58	8	1	7	23	15	1	42	3	0.37	0.32	7	0
145	139	17	1	8	16	33	0	45	0	0.58	0.30	5	2
155	-	23	2	13	28	16	0	16	2	2.38	0.40	14	0
157	-	25	1	14	21	19	4	12	4	3.33	0.45	14	2
159	-	32	2	11	14	16	1	22	2	2.05	0.41	10	0
Average		19.4	1.43	11.1	20.4	18.0	0.9	29.6	2.0	1.52	0.37	9.57	0.7

* Across inlet-not counted in mean

APPENDIX D

CARBONATE ANALYSES OF THE -200 MESH FRACTION

Sample Number	Calcite (wt. %)	Dolomite (wt. %)	Total Carbonate
Granite Canyon Till			
121	0.7	0.6	1.3
Glacier Bay Till			
39	2.4	1.1	3.5
40	2.6	1.7	4.3
41	2.7	0.6	3.3
42	2.6	0.8	3.4
46	2.1	0.4	2.5
48	2.0	0.8	2.8
50	1.8	0.8	2.6
51	2.6	0.3	2.9
62	1.5	0.7	2.2
63	1.6	0.2	1.8
101	5.6	0.9	6.4
102	7.5	0.9	8.4
103	6.0	1.1	7.1
105	4.9	1.1	6.0
107	0.9	0.3	1.2
108	0.9	0.3	1.2
122	0.8	0.2	1.0
123	1.1	1.2	2.3
124	1.6	0.6	2.2
140	1.6	0.5	2.1
141	1.4	0.4	1.8
143	2.3	0.8	3.1
144	0.9	0.3	1.2
149	0.3	0.8	1.1

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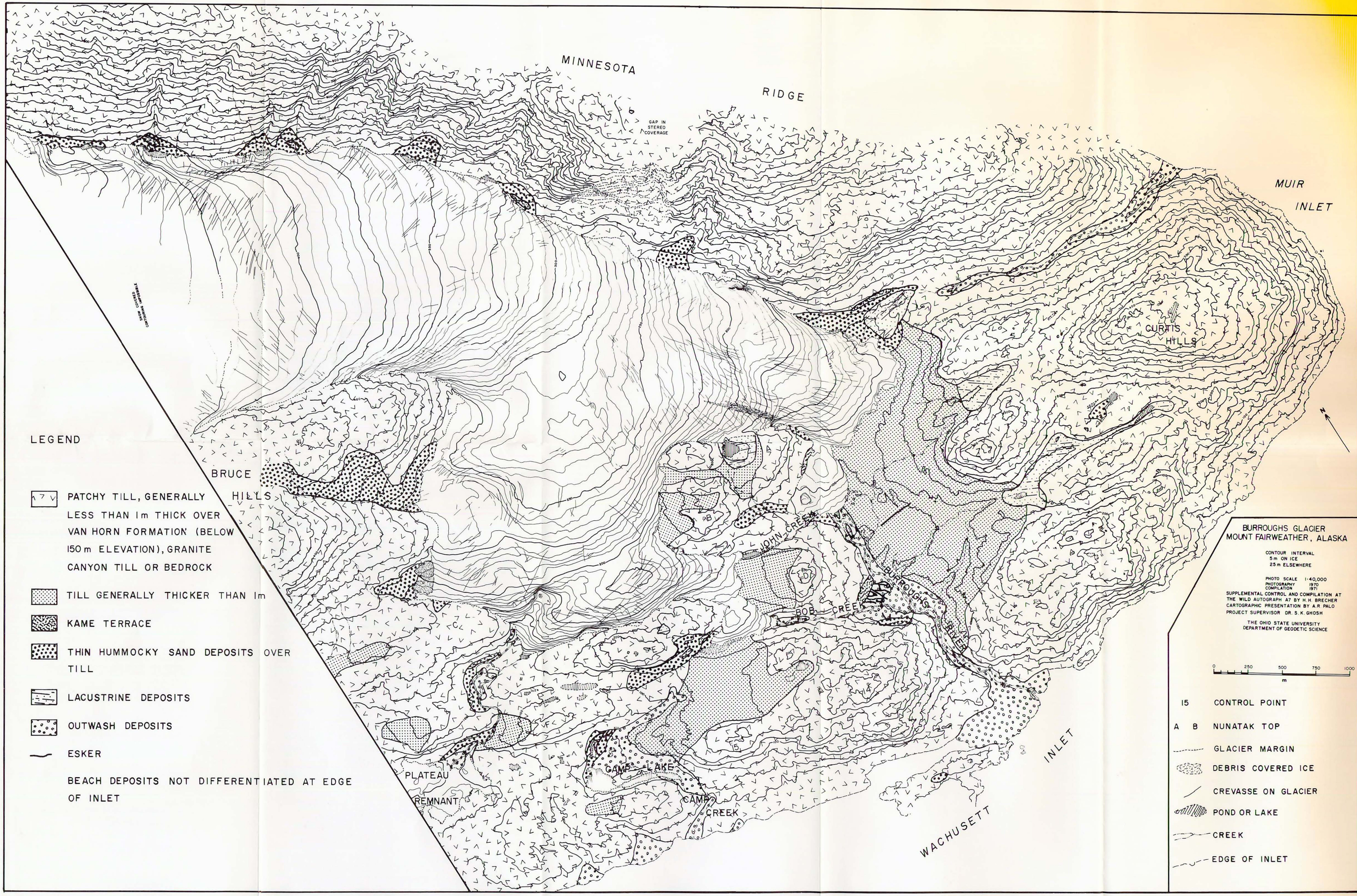

FIGURE 5. SURFICIAL GEOLOGY SURROUNDING THE SOUTHEAST TERMINUS OF BURROUGHS GLACIER.

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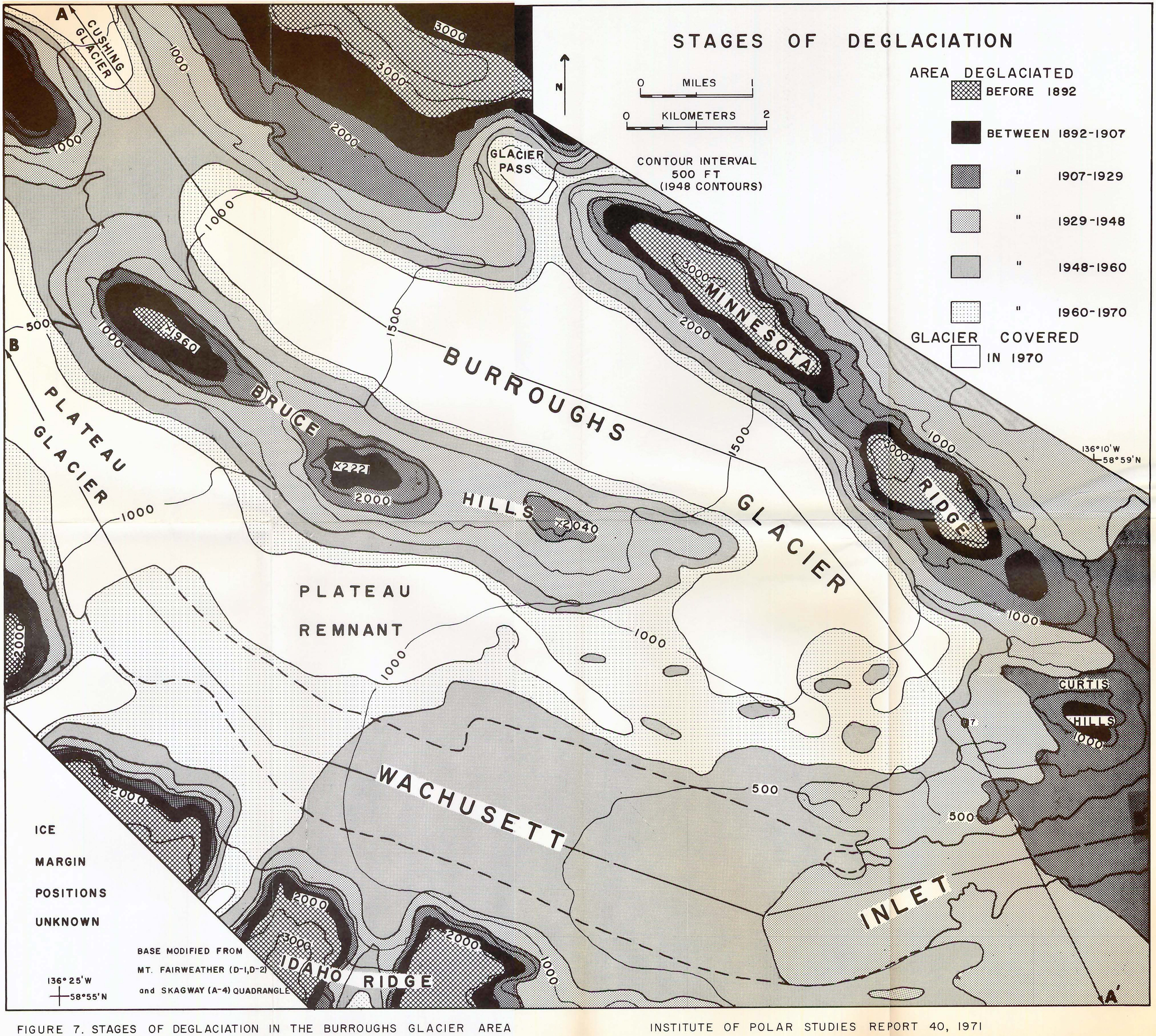

FIGURE 7. STAGES OF DEGLACIATION IN THE BURROUGHS GLACIER AREA

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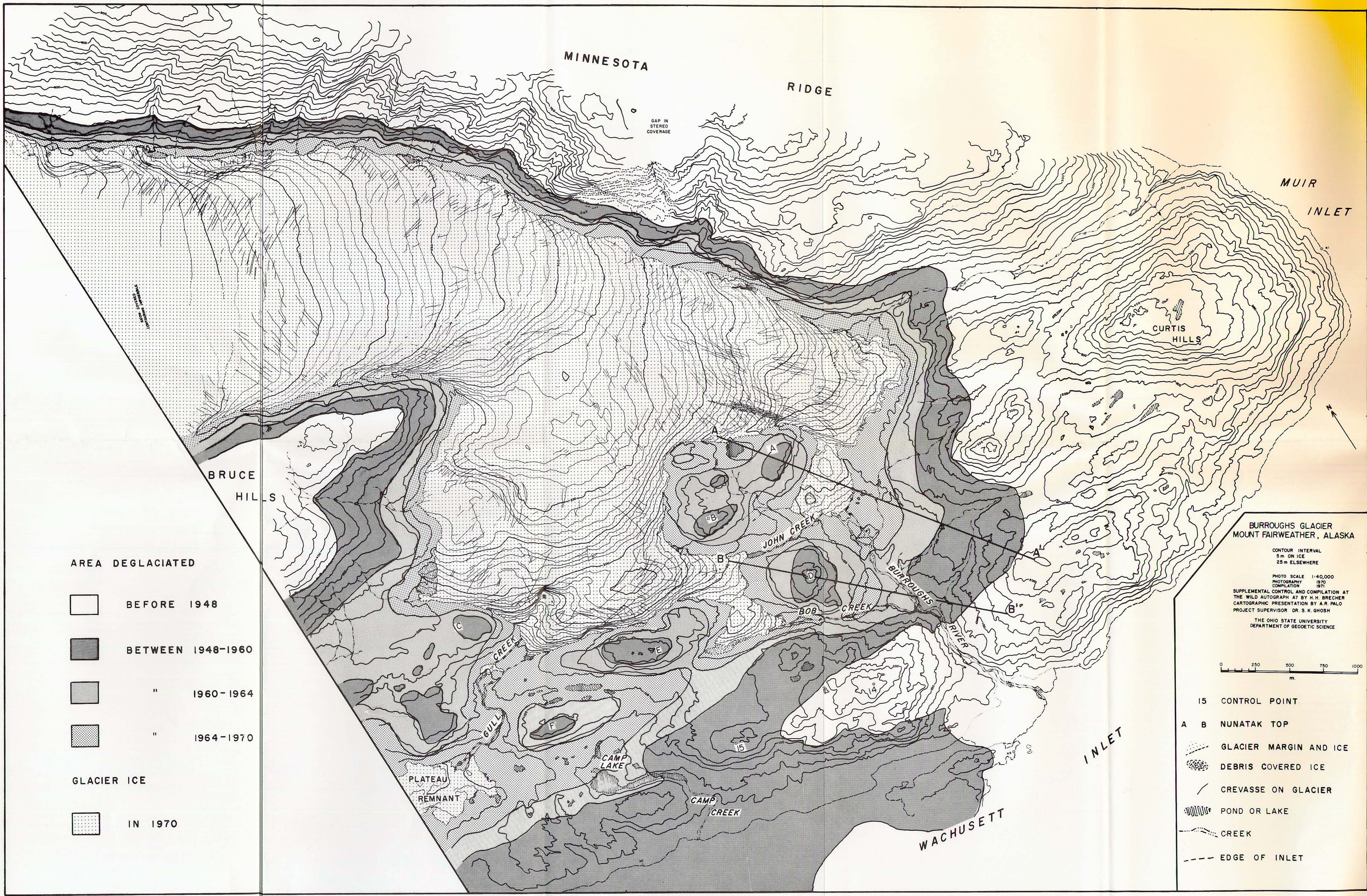

FIGURE 10. MARGIN POSITIONS OF THE SOUTHEAST TERMINUS OF BURROUGHS GLACIER

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