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GLACIAL GEOLOGY OF THE
BRADY GLACIER REGION, ALASKA

DISSERTATION
Presented in Partial Fulfillment of the Requirements for
the Degree Doctor of Philosophy in the Graduate
School of the Ohio State University

By
Stephen J. Derksen, B.S., M.S.

*****

The Ohio State University
1976

Reading Committee:
R. Goldthwait
G. McKenzie
K. Stanley

Approved By

Richard O. Goldthwait
Adviser
Department of Geology
and Mineralogy
ACKNOWLEDGMENTS

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VITA

October 4, 1946 ... Born - Cincinnati, Ohio

1968 ............. B.S., The University of Alabama, Tuscaloosa, Alabama

1972 - 1975 ...... Research Associate, Department of Geology and Mineralogy, The Ohio State University, Columbus, Ohio

1974 ............. M.Sc., The Ohio State University, Columbus, Ohio

1975 - 1976 ...... Research Associate, The Institute of Polar Studies, The Ohio State University, Columbus, Ohio

PUBLICATIONS


FIELDS OF STUDY

Major Field: Glacial Geology

Studies in Glacial Geology. Professor R.P. Goldthwait

Studies in Geomorphology. Professor S.E. White

Studies in Analysis of Soil Materials. Professor L. Wilding


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Plate 1. Surficial geology of the Brady Glacier region
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CHAPTER I
INTRODUCTION

Purpose of Investigation

The 520 km$^2$ Brady Glacier lies in the western part of Glacier Bay National Monument in southeastern Alaska. This large valley glacier has experienced major changes in its size and extent in the distant and recent past which have greatly affected the morphology and ecology of this region. It was the purpose of this investigation to: (1) unravel the complex glacial history of the Brady Glacier, and (2) to date, if possible, post-Wisconsin and older sea levels recorded by raised marine terraces and associated deposits along the coast from Icy Point to Dundas Bay. As the study progressed, it was augmented by: (1) a reconnaissance determination of the present Brady Glacier mass balance, (2) a description of the nature of modern and ancient glacial sediments found in the area, and (3) an examination of the orientation, relief, and vertical distribution of abandoned cirques formed in the mountainous terrain surrounding this glacier.
Previous Investigations

The earliest published references to the Brady Glacier and its environs are entirely descriptive. Capt. George Vancouver (1798) visited this area in 1794 and published a detailed description and map of the region. C.E.S. Wood (1882) briefly described the appearance of the Brady terminus in an account of a visit to the area in 1877. John Muir, accompanied by the Rev. S.H. Young, visited the glacier terminus in 1880 and published several descriptive accounts of his exploits in this area (Muir, 1897, 1909, 1915; Young, 1915). The Brady was named in 1893 by the U.S. Geological Survey for the Rev. John Green Brady, Alaskan missionary and later governor from 1897 - 1909. The glacier was first photographed and mapped by the Canadian Boundary Survey in 1894 (Alaskan Boundary Tribunal, 1904), and re-surveyed by the U.S. Boundary Survey in 1907 (Inter. Boundary Commis., 1952). The region was again mapped by the U.S. Geological Survey in 1961 on the basis of aerial photography taken in 1948.

Bedrock geology of various parts of this region has been mapped by Seitz (1959), Rossman (1963 a & b), Plafker (1967), and Mackevett et al. (1971). Rossman also mapped surficial deposits, but did not distinguish between stream gravel, glacial detritus, and talus. Glacial history of the Brady Glacier was mentioned briefly by Klotz (1899, 1907), and more recently by Streveiler and Paige (1971) and Streveiler and Worley (1974), but the most detailed study of recent glacial events was made by Bengtson (1962). Bengtson worked primarily on the northeastern side of the Brady, but did recognize the basic sequence of Holocene glacial
events which the present study has been able to amplify in considerable detail. Ancient marine tillites which crop out in this region have been described by D.J. Miller (1953, 1958), Rossman (1963b), Plafker (1967), and Plafker and Addicott (1976).

**Geographic Setting**

The 40 km-long Brady Glacier flows to the southeast down a long, fault-line valley just east of the Fairweather Range between 58° 20' and 58° 40' North latitude and 136° 35' and 137° 10' West longitude (map, Figure 1). The Brady is fed by a large icefield formed in the lee of the peaks of the Fairweathers which attain heights of up to 4700 m. Some of this ice flows northwest from a drainage divide to become the Reid and Lamplugh Glaciers, but more than 70% of it flows south-southeast as the Brady Glacier. The major terminus of the glacier rests above tidewater on its outwash plain built into Taylor Bay, but 9 ice lobes protrude some distance down smaller northeast–southwest trending valleys. The Brady is generally about 4.5 km wide and possibly as much as 800 m thick since the bottom of the glacier probably lies in a deep fiord excavated below sea level. Post and Mayo (1971) noted the presence of 19 ice-dammed lakes larger than 0.1 km² around the margins of the glacier. The ragged hills surrounding the lower part of the Brady are covered by a dense spruce-hemlock forest up to treeline at about 550 m elevation and provide a habitat for an abundance of wildlife. The glacier and its surrounding region as it appeared in 1973 is
Figure 1. Glacier Bay National Monument, Alaska, indicating the Brady Glacier region shown in detail by Plate 1.
shown by Figure 2. A small tributary glacier near the head of Palma Valley, here called the Palma Glacier, is now separated from Brady ice by an ice-dammed lake (Plate 1).

**Climate**

The cool, moist, maritime climate of the Brady Glacier region is typical of southeastern Alaska. Climate is dominated by the strong Aleutian low pressure system lying to the west which generates frequent storms which move in from the Gulf of Alaska bringing high amounts of precipitation to the mountainous coastal zone. The nearest weather station to the Brady Glacier is near sea level at Cape Spencer. Average monthly temperature varies here between 0.1°C and 11.5°C during the year, and annual precipitation is commonly greater than 280 cm; most of it falling during the Fall and Winter months. Only 10% of the days at Cape Spencer are cloud-free. Coastal fog is common between June and September. Precipitation in the form of snow is common from late October until early April at sea level, but the summer period becomes progressively shorter at higher elevations. Winds are generally from the southeast quadrant from the last of October to the middle of May and are often accompanied by rain or drizzle at sea level. Summer winds are weaker and more variable in direction.
Figure 2. The Brady Glacier region as it appeared in 1973. The highest peak visible is Mount Crillon, 3870 m; the prominent peak to the left of it is Mount LaPerouse, 3270 m. Icy Point, stretching into the Gulf of Alaska, is visible on the upper left. Most fieldwork was done in the rugged terrain on both sides of the lower part of the glacier. Note the extensive outwash plain deposited at the terminus which slopes down to tidewater in Taylor Bay (out of the photo on the bottom left), and the prominent abandoned cirque visible near the bottom right. (Photo by Austin Post)
Bedrock Geology

The high peaks of the Fairweather Range to the west of the Brady Glacier are in large part composed of a layered gabbroic intrusion of Cretaceous or early Tertiary age (map, Figure 3). Southwest of the glacier, the gabbro can be seen to intrude a large area of mostly Mesozoic (?) amphibole and biotite schist with a few discordant plutons of unfoliated granodiorite, adamellite, or granite.

Further west, across the Fairweather Fault near Icy Point occur sedimentary rocks of the Topsy and Yakataga Formations. The Topsy Formation in this area includes siltstones, sandstones, and conglomerates of early Oligocene to early Miocene age. The overlying Yakataga Formation contains siltstones, sandstones, conglomerates, and marine tillites and is of early Miocene to early Pleistocene in age. (Both formations have been dated on the basis of fossil mollusks).

Southeast of the glacier there is a region of strongly metamorphosed gneisses, mixed contact zones, and local occurrences of carbonate rocks. The gneisses are mainly tonalitic to dioritic, but some massive breccias also occur. Contact zones consist of large hornfels masses surrounded by rock of the adjacent intrusive body. North of the metamorphic zone bedrock is composed of partly discordant plutons of Mesozoic (?) age made up of foliated granitic rocks (mostly tonalite, diorite, and granodiorite).

The Glacier Bay region is dominated by a northwest-southeast trending fault system, of which the Fairweather Fault is typical. This strike-slip fault separates sedimentary rocks of Tertiary age from older
Figure 3. Map showing simplified bedrock geology in the Brady Glacier region. Sources: Seitz (1958), Roisman (1963 a & b), Plafker (1967), and Mackevett et al. (1971).
metamorphic rocks east of Icy Point, and has been active in historic and recent time (D.J. Miller, 1960; Tocher, 1960). A less-pronounced, complementary fault system running northeast-southwest is also present. Twenhofel and Sainsbury (1958) believed most of the major valleys in this region were fault controlled. The nearly vertical dip of rocks of the Yakataga Formation, as young as early Pleistocene, attests to the active tectonism of this region.
CHAPTER II
PLEISTOCENE GLACIAL HISTORY

Pre-Wisconsin Glaciations

The rugged terrain of this region sculptured into sharp peaks and ridges separated by deep, U-shaped valleys, the smoothed and worn lower mountains, the numerous fiords, and the virtual absence of more than a thin regolith over bedrock slopes everywhere attest to a long history of vigorous, repeated glaciation. The marine tillites of the Yakataga Formation exposed along the Gulf of Alaska coast near Icy Point (Figures 4 and 5) indicate that local mountain glaciers were large enough to calve into the sea by early Miocene (Plafker and Addicott, 1976). Near its base are horizons containing plankton indicative of glacial conditions (Brandy et al., 1969). Further north along the coast southeast of Lituya Bay occur subdued moraines of probable pre-Wisconsin age (Berksen, 1974, 1975). These moraines are the oldest preserved subaerial glacial deposits known in southeast Alaska. However, no pre-Wisconsin deposits are known from the Brady Glacier region east of Icy Point which suggests that the Wisconsin glaciation of this area was the most extensive of the Quaternary. This may reflect continuing uplift of the Brady's source area in the high peaks of the Fairweather Range.
Figure 4. Steeply-dipping marine tillites exposed along the Gulf of Alaska near Icy Point. Dropstone-rich layers are interstratified with layers containing few dropstones.

Figure 5. Closer view of marine tillites shown in Fig. 4. Note the angular pebbles and clasts of mixed lithology enclosed in a silty matrix.
Pre-Wisconsin Marine Platform Cutting

Northwest of the Brady Glacier in the area around Lituya Bay occur as many as five major raised marine terraces. The four highest terraces are believed to date from former interglacial periods while the lowest terrace has been cut and uplifted since Wisconsin glaciation (Derksen, 1974, 1975). Four of these major terraces extend as far south as Icy Point. Remnants of the lowest pre-Wisconsin terrace can be traced south across the Fairweather Fault at a similar upper elevation of about 30 m (Plate 1). This terrace is well preserved around the mouths of Dixon Harbor, Torch Bay, Graves Harbor, and at Cape Spencer (Figure 6).

Within Graves Harbor, at a protected area near sea level in Murphy's Cove, the 30-m terrace's surface retains glacial grooves and striations partially covered by till (Figure 7). Nearby, between Graves Harbor and Murk Bay, glacial grooves are still evident on an exposed bedrock point. Both of these localities are several kilometers beyond the outermost Neoglacial deposits, but indicate the presence of a major ice mass in Graves Harbor. The absolute age of the till resting on the terrace surface is unknown, but it lies in a geographic location that probably was eroded by massive Wisconsin ice sheets (Plate 1). Therefore, the till is probably as young as Late Wisconsin. Since all traces of any older marine terraces have been removed by repeated regional glaciation, the age of this 30-m terrace must be comparatively recent, but pre-Late Wisconsin. Therefore, it probably was cut during the long Sangamon Interglacial period of high sea level and then uplifted to its
Figure 6. 30-m, Raised marine terrace near Cape Spencer. Taylor Bay and the Brady Glacier lie in the distance on the right side of the photo.

Figure 7. Criss-crossing grooves and striations cut by Wisconsin ice near sea level in a protected part of Murphy's Cove. The notebook lies in a glacial groove running right to left; the striations were cut by later ice moving from the top to the bottom of the photo.
present elevation over the past 75,000 years. During a mid-Wisconsin warm period sea level rose eustatically (Shepard, 1963; Curray, 1965; Milliman and Emery, 1968; Guilcher, 1969; Hopkins, 1967, 1973), but the magnitude and duration of this rise in sea level was probably insufficient to cut this pronounced landscape feature.

Wisconsin Glaciation

Extensive glaciation of southeast Alaska occurred as recently as Wisconsin time when a Cordilleran ice sheet thousands of meters thick buried most of the coastal ranges and locally formed a floating ice shelf which calved into the eustatically lowered Gulf of Alaska (Coulter et al., 1962; Flint, 1971). In the Alaska-Canada Boundary Range, M.M. Miller (1964a, 1975) recognized both an Early and Late Wisconsin stage separated by a cool "interglacial". A peat layer buried by Late Wisconsin outwash near Juneau was dated at greater than 39,000 B.P. (W-2721; R.D. Miller, 1973b). The Late Wisconsin stage lasted from perhaps 40,000 to 14,000 B.P., and was followed by three substages recognized on the basis of weathering differences between till units and changing flow directions of progressively smaller mountain ice sheets. The recession of Late Wisconsin ice to at least its present limits occurred by 12,000 B.P. on the eastern flanks of the St. Elias Range to the north of Glacier Bay (Denton and Stuiver, 1967), and by 12,000 to 9,000 B.P. in the Juneau Icefield and Glacier Bay regions (Goldthwait et al., 1963; Goldthwait, 1966; McKenzie and Goldthwait,..
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There are no absolute age determinations for Wisconsin glacial events recognized in the vicinity of the Brady Glacier. However, the general chronology of Wisconsin glaciation recognized elsewhere in the same region can probably be safely applied to the Brady.

The paucity of glacial deposits of known Wisconsin age in the vicinity of the Brady Glacier, and elsewhere in southeastern Alaska, seems remarkable. This is probably due to: (1) nearly complete burial and erosion of terrain at lower elevations by vast mountain icesheets, (2) earlier glaciations removing easily-eroded, weathered bedrock from hillsides, (3) thick icesheets depositing most debris at their periphery where it was drowned by post-glacial rise in sea level, (4) steep slopes and abundant rainfall aiding post-glacial mass wasting and erosion of mountainside glacial deposits, (5) extensive Neoglacial ice advances eroding or burying older glacial deposits, and perhaps (6) relatively rapid deglaciation at the end of the Wisconsin (Broecker and Van Donk, 1970) which might preclude the formation of significant recessional moraines and other glacial deposits at the margins of the receding ice.

The orientation of glacial grooves and striations was measured on high ridge crests overridden by Wisconsin ice from Delangle Mountain to the ridges around western Dundas Bay. These markings show that ice came from the east-northeast at Delangle Mountain, but flowed more from the northeast at Graves Harbor and Dundas Bay (Plate 1). This pattern indicates that ice from the greatly enlarged Brady icefield merged with ice from Glacier Bay to form a vast icesheet which flowed seaward during the
height of Late Wisconsin glaciation. Glacial markings left by later, waning stages of glaciation at lower elevations are today mostly hidden by a thick cover of vegetation. The greatest Quaternary glaciation of the Brady Glacier region probably occurred in Late Wisconsin time since there is no discernable change in bedrock weathering, lichen cover, or presence of intermediate moraines or other glacial deposits between the highest glacial limit recognized on the sides of former nunataks and the highest lateral moraines of known Neoglacial age at lower elevations.

Retention of small-scale glacial markings caused by ice abrasion is a function of bedrock type and orientation to the moving ice. The hard, granitic rocks to the east of the Brady Glacier best retain ice scour features such as grooves, striations, crescentic gouges, chattermarks, and some glacial polish. In contrast, finer glacial markings have been mostly lost from the less-resistant schist to the southwest of the glacier. In places, the schist has weathered and eroded as much as 15 cm from its post-glacial surface which can be recognized by the flattened, striated surfaces of projecting quartz inclusions within the schist. Nevertheless, even the schist retains larger features of glacial abrasion on projecting rock spurs and on low saddles between mountains where ice motion was concentrated. Especially well developed roche moutonées and grooves 30 cm wide, 10 cm deep, and several meters long were cut into the saddles between Torch Bay and South Deception Lake, on the ridge northwest of Graves Harbor, and on the gneissic terrain between southwest Dundas Bay, Taylor Bay, and Cross Sound (Plate 1).
Late Wisconsin Glacial Limits and Ice-free Areas

The glacial limits of Late Wisconsin ice in the Brady Glacier region can be recognized by the contrast between smoothed, glacially-abraded topography formed below the ice margins (Figure 8) littered by occasional erratic rock fragments, and frost-shattered "tank and tor" topography above (Figure 9). Figure 10 illustrates the change in slope topography from rounded, below the limit, to jagged bedrock outcrops above on the northeastern flank of DeLangle Mountain at an elevation of about 700 m. The glacial limit on the southern end of this mountain is also marked by the rare presence of a small moraine containing heavily lichen-covered erratic boulders at an elevation of 625 m (Figure 11). Similar criteria were used to define ice limits in the nearby Juneau Icefield region (M.M. Miller, 1975), Glacier Bay (McKenzie and Goldthwait, 1971), and elsewhere (Flint and Fidalgo, 1964).

Contouring of a bathymetric navigational chart (NOAA, 1972) of the offshore area from Icy Point to Cape Spencer reveals two interesting features of the sea floor (Figure 12). First, a series of linear, closed basins extends southeast from Icy Point suggesting the continuation of the great Fairweather Fault in this direction. Secondly, the character of bottom topography changes from smooth and undulating at depths below 55 ± 5 fathoms (~100 m ± 10 m) to more uneven and contorted above. A large, irregular shallow area lies at this depth off the mouth of Graves Harbor suggestive of a moraine deposited at the end of a deeply scoured fiord. A similar situation occurs at Torch Bay and seaward of the Fairweather Glacier further north as well. If these shallow areas
Figure 8. Glacially-smoothed, rounded topography characteristic of the Brady Glacier area below the glacial limit.

Figure 9. Frost-shattered tors above the glacial limit on the Astrolabe Peninsula. Note the figures for scale.
Figure 10. The northeastern flank of DeLangle Mountain. Note the change in slope morphology from smooth and rounded to jagged at the glacial limit (ca. 700 m elevation).

Figure 11. Small Wisconsin-age lateral moraine deposited on the edge of a cliff at an elevation of 625 m on the southeastern flank of DeLangle Mountain. View is to the southwest, toward the Gulf of Alaska.
Figure 12. Bathymetry of the sea floor from Icy Point to Graves Harbor. Note the continuation of the Fairweather Fault southeast from Icy Point and the change in bottom morphology at a depth of about 55 fathoms.
were merely bedrock "lips" formed at the end of fiords, they should lie further inland where confining valley walls suddenly diverge.

A plausible explanation for these phenomena is simply that this depth of about 55 fathoms reflects sea level during former glacial periods. Mountain ice sheets probably terminated at or near the coastline without the formation of a major floating ice shelf, much as the nearby LaPerouse Glacier and exposed portions of the Antarctic or Greenland coast do today. Walcott (1972) developed a crude model of global isostatic deformation which predicts that lowest Wisconsin-age sea level deposits in southeastern Alaska should lie at approximately 70% of their former depth due to post-glacial isostatic adjustments of both continents and ocean basins. Using Walcott's model, Chappel (1974) estimated the maximum eustatic lowering of sea level during the Wisconsin to have been 130 to 135 m. Thus, lowest Wisconsin sea level in southeastern Alaska should be observed at depths of -90 to -95 m according to this model. This predicted former shoreline agrees well with the 55 \pm 5 fathom line (-100 \pm 10 m) on this chart.

Even during the climax of Wisconsin glaciation certain parts of this region remained ice-free (Figure 13). A few peaks above about 850 m on the east side of the glacier, and DeLangle Mountain, the Astrolabe Peninsula, and Horn Mountain at an elevation above about 600 m protruded through Wisconsin ice as nunataks. This indicates a flowline ice surface gradient toward the coast of about 1.4%, fairly typical of the lower portions of large, modern Alaskan coastal glaciers such as the Malaspina.
At lower elevations, along the coast north of Icy Point the bulk of prominent moraines built around Lituya Bay, Crillon Lake, and by the Dagelet, LaPerouse, and Finger Glaciers was probably deposited by Late Wisconsin ice, leaving large coastal areas ice-free (Heusser, 1960: D.J. Miller, 1961: Goldthwait et al., 1963: Derksen, 1974). Ice from the seaward side of the Fairweather Range was channelized along the Fairweather Fault and only found easy access to the sea by a few exits through the coastal (1000 m) mountains (Figures 1 and 3). Late Wisconsin sea level may have been eustatically lowered by as much as 135 m (Chappel, 1974), but parts of southeastern Alaska may have been depressed isostatically by as much as 213 m based on the present elevation of post-glacial marine deposits in the Juneau area (Buddington, 1927: Twenhofel, 1952: R.D. Miller, 1973a). This suggests that the coastal area near Icy Point may have been partially submerged during the Wisconsin. However, post-glacial marine deposits in the Glacier Bay area have only been found up to a present elevation of 60 m (McKenzie and Goldthwait, 1971). Furthermore, Derksen (1974) cited evidence that the marine terraces above 30 m elevation southeast of Lituya Bay were of considerably greater age than the Wisconsin. Therefore, most of the area from Icy Point north to the LaPerouse Glacier probably remained ice-free and above sea level (Figure 1 and 13). This area might have served as a biologic refugium for those organisms able to withstand the harsh climatic conditions of Wisconsin time.
Figure 13. Late Wisconsin maximum of the Brady Glacier, ca. 15,000 - 20,000 B.P.

Figure 14. Post-glacial marine transgression, ca. 9,000 - 13,000 B.P.
Relict Periglacial Features

No relict periglacial features have been found on the ice-free areas at lower elevations, but this terrain is mostly covered by dense vegetation. However, at elevations above treeline relict periglacial features are common, especially on mountain summits that stood above Wisconsin ice limits. The most common of these features are frost-shattered rock tors up to 20 m in height (Figure 9). Many of these are still actively forming on the highest summits, but relict forms can be found below modern treeline. Also common below the glacial limit are rock "tanks", described by M.M. Miller (1975) and explained as the result of differential plucking by moving ice and subsequent modification by nivation.

Above the glacial limits on Delangle Mountain occur a rock glacier, nivation hollows, small cirque moraines, a blockfield, stone stripes, and frost-shattered bedrock (Figures 15 and 16). These lichen-covered, partially vegetation-covered features appear inactive under present climatic conditions. The lobate rock glacier shown in Figure 16 lies 300 m below presently active rock glaciers in the Alaskan coastal ranges (M.M. Miller, 1975). A slight difference in lichen cover between the uppermost and lower lobes of this rock glacier suggests that it has been recently reactivated, probably during Neoglacial periods of colder climate (Heusser, 1960). Two small cirque glaciers and periglacial features such as nivation hollows and stone stripes may have been active in Neoglacial time as well. However, the contrast in development of periglacial features above and below the Wisconsin glacial
Figure 15. Frost-shattered, gabbroic bedrock above the glacial limit on the southern end of DeLangle Mountain at an elevation of 650 m.

Figure 16. Multi-lobate rock glacier above the glacial limit on the western side of DeLangle Mountain. Palma Valley lies below on the right, the Gulf of Alaska to the left.
limits indicates that most of these features are relict from much harsher Wisconsin climatic conditions.
As the glaciers waned at the close of the Wisconsin glacial, marine waters transgressed over the still isostatically depressed landscape (Figure 14). A small raised marine terrace and sea cliff up to 5-11 m above mean high tide found throughout the Brady Glacier region was probably cut at this time. This cliff is especially well-developed on the southeastern side of the Boussole Valley and in parts of Thistle Cove, Dixon Harbor, Torch Bay, Graves Harbor, and both Taylor and Dundas Bays (Plate 1). Relief on this sharp cliff is often up to 25 m, and neither the cliff or terrace bear any trace of glaciation. No till is present on the terrace surface.

In the southwestern arm of Dundas Bay this prominent sea cliff was cut in what is a very sheltered area today. The presence of this cliff in this quiet part of the bay suggests open water between Dundas and Taylor Bay during its cutting, allowing entry of high winds and Pacific swells from Cross Sound (Plate 1). Since this opening was blocked by the deposition of Neoglacial sediments as long ago as 2870 ± 80 years B.P. (DIC-460; see Appendix A), the cutting of this cliff and terrace
must have been more than 2900 years ago.

Withdrawal of ice from the Katalla Valley 400 km to the northwest was followed by a marine invasion prior to 14,000 B.P. (Sirkkin and Tut- hill, 1969). The post-glacial marine Forest Creek Formation in upper Glacier Bay has been radiocarbon dated by five dates between 10,000 ± 220 B.P. and 13,960 ± 360 B.P. (McKenzie and Goldthwait, 1971). A prominent terrace and sea cliff at an elevation of 6-15 m above modern beach heads in Icy Strait, 40 km southeast of Glacier Bay, has been dated between 13,350 ± 100 B.P. and 9,130 ± 130 B.P. (T.D. Hamilton, manuscript in preparation, 1976). Hamilton suggests this terrace formed during a slowdown of isostatic recovery caused by renewed glacial advances 11,500 to 11,000 years B.P. However, stumps exposed near sea level in Lituya Bay are as old as 9150 ± 275 B.P. (Goldthwait et al., 1965) which suggests that the isostatic rebound of the region had about reached its present state by this date. Therefore, the cutting of the widespread 5-11 m terrace and cliff in the Brady Glacier region must have occurred more than 2900 years ago and probably was completed prior to about 9000 B.P., when isostatic uplift raised the land above the eustatically rising sea.

**Hypsithermal Ice Recession**

The Hypsithermal was a long period of warmer (by ca. 1°C), dryer climate from about 7000 to 4100 years B.P. This interval was characterized by smaller glaciers than at present in this region except in the
northwest part of Glacier Bay (Heusser, 1960; Goldthwait, 1963, 1966). It was a time of valley filling by great thicknesses of outwash derived from the shrunken glaciers in the upper parts of Glacier Bay (Goldthwait, 1963, McKenzie and Goldthwait, 1971). No till or outwash deposits in the vicinity of the Brady Glacier have been found to date from this interval which suggests that this glacier also shrank to less than its present size. Figure 17 illustrates the Brady as it might have appeared at the beginning of this warm period. However, since the Brady Glacier fills what would probably be a fiord if the ice were to recede, and since fiord glaciers are subject to catastrophic recession (Mercer, 1961b; Post, 1975), the entire glacier may have retreated during the warmest part of the Hypsithermal until only small remnants were left streaming down the eastern flanks of the Fairweather Range. Taylor Bay might have become a long fiord extending 70 km from Cross Sound to upper Glacier Bay. In a similar manner, ice in nearby Glacier Bay has rapidly receded over the past 225 years to open up a fiord system over 100 km in length (Cooper, 1923, 1937; Field, 1947; Lawrence, 1958; Goldthwait, 1963).
Figure 17. The Hypsithermal Interval, ca. 4,100 - 7,000 B.P.

Figure 18. The Early Neoglacial maximum advance of the Brady Glacier, ca. 1,230 - 1,960 B.P.
Early Neoglacialization

Evidence of Multiple Ice Advances

A geographic and stratigraphic description of 15 radiocarbon dates from the Brady Glacier region is given in Appendix A. Of these dates, 8 relate directly to Early Neoglacial deposits left by the Brady Glacier (Y-9; DIC-284, 285, 458, 459, 460, 462, 555). These dates range in age from 4680 ± 160 B.P. (Y-9) to 1230 ± 60 B.P. (DIC-459). However, 6 of these 8 dates (DIC-284, 285, 458, 459, 462, 555) seem to cluster in the range from 1960 to 1230 radiocarbon years B.P. There is evidence that these younger dates come from a more recent, maximum phase of Early Neoglacialization, and that the remaining two dates, Y-9 and DIC-460, relate to older, probably separate advances of the Early Neoglacial period.

The oldest date, 4680 ± 160 B.P. (Y-9) was collected by K.B. Bengtson in 1950 from very compact till on the east side of the terminus of the Reid Glacier, the northeastern distributary of the Brady Icefield (Preston, 1955). Thus, the Brady Icefield must have existed in at least its present state by that date.

It is open to speculation whether or not this advance about 4700 B.P. of the Brady Glacier was followed by recession, but by 2870 ± 80 B.P. (DIC-460) the terminus must have advanced far enough down Taylor Bay to deposit intercalated outwash and intertidal sediments near southwestern Dundas Bay (Plate 1). The glacier must have been considerably larger than at present at about this time.
Following this 2900 B.P. advance, the Brady must have receded to at least its present margins. The next date in chronological order, 1960 ± 90 B.P. (DIC-458), comes from a forest layer on outwash buried by sediments from the rising waters of a glacially-dammed lake which was eventually overridden by advancing ice. This site, in the upper Boussole Valley (Plate 1), is remarkable because of the nature of the lake's confining boundaries. Advancing ice could easily have dammed the lake on the north, a bedrock ridge lies to the southeast, but to the southwest lies only a thick section of till and outwash filling Boussole Valley and exposed along the Dixon River. Thus, the third boundary of this glacial lake must have been pre-existing glacial sediments, probably deposited at the time of the 2900 B.P. advance. Therefore, (1) ice must have advanced to some extended position in the Boussole Valley depositing moraines and outwash about 2900 B.P., (2) the ice receded leaving these glacial deposits as higher ground, (3) a forest grew "inside" the older moraines, and (4) the forest was killed by an ice-dammed lake as the Brady advanced again some 1960 radiocarbon years B.P.

The climax of the Early Neoglacial is well dated by the remaining six radiocarbon dates from about 1960 to 1230 radiocarbon years B.P. This advance deposited the outermost moraines and outwash in the Boussole Valley, the Dixon River Valley, and southwest Dundas Bay, and probably built the outermost moraines found in the Palma Valley, Graves Harbor, Taylor Bay, and in two valleys near the northwest arm of Dundas Bay as well (Plate 1). A reconstruction of the glacier's margins at this time is shown by Figure 18.
Age and Correlation of Lateral and End Moraines
in Taylor Valley

Within Taylor Valley, high on the walls on each side, are perched lateral moraines that appear to have been deposited by Early Neoglacial advances of the Brady Glacier. Elevations of the highest moraines were measured by altimeter on both sides of the valley. These moraines, end moraines in Taylor Bay, and the centerline profile and margins of the modern Brady Glacier are all projected onto the centerline of Taylor Valley in Figure 19. Distance is measured from the 1948 "modern" terminal position of the glacier. A curve was "best-fit" by eye to the modern Brady profile and reproduced through the altimeter stations indicated. It can be seen that the 1948 glacier margins do tend to parallel the glacier centerline profile, but are disturbed by ice escaping down branch valleys. Thus, the projection of the "best-fit" 1948 profile through the altimeter stations onto the positions of end moraines in Taylor Bay seems a reasonable, if only approximate, correlation between lateral and end moraines. Taylor Valley is generally uniform over much of this distance.

The high lateral moraines on the eastern side of the valley seem to fit well with the outermost, best-developed terminal moraine "1" on the eastern side of the Bay (Figure 19; Plate 1). Moraine "1" appears to be an extension of the radiocarbon-dated Early Neoglacial deposits (DIC-460, 555) found a few kilometers northeast. Thus, the highest lateral moraines on the eastern side of the valley are probably of Early Neoglacial age.
DISTANCE MEASURED FROM THE MODERN (1948) BRADY TERMINUS (Km)

ELEVATION ($\times 10^2$ M)

Figure 19. Correlation of Early Neoglacial lateral and end moraines of the Brady Glacier in Taylor Valley. Perched lateral moraines, end moraines, and the 1948 centerline profile and margins of the Brady Glacier are shown projected onto the centerline of Taylor Valley. A best-fit curve to the 1948 glacier profile is reproduced passing through the altimeter stations on each side of the valley.
However, the highest (and also best-developed) lateral moraines on the western side of the valley are somewhat lower (as is the case with modern ice margins), and seem to fit best with small end moraine "III". Although this correlation is highly uncertain, it does suggest that moraines "I" and "II" further down the Bay might date from some older advance than Early Neoglacial. Furthermore, a few large, cavernously-weathered, granodiorite boulders occur on the surface of moraines "I" and "II", and these appear to have weathered in place (Figures 20, 21). Weathering and erosion of granitic rocks is believed to require considerable time (Calkin and Cailleux, 1962), although rates of 2-3 cm/1000 years have been observed in the humid climate of Germany (Winkler, 1965). A few rounded granodiorite boulders were observed on moraines "I", "IV", and "V", but the degree of weathering was markedly less. Two of the boulders on moraine "IV" were striated and, on one of these, the abrasion could be seen to post-date the rounding.

Thus, two possibilities for the ages of these moraines exist. First, that all the outermost moraines are Early Neoglacial. The few cavernously-weathered boulders could have been transported a short distance from much older deposits, or might even have weathered in place over the past 1200 - 2900 years. Second, there is the possibility that moraines "I" and "II" date from some much earlier period of ice advance, such as a waning stage of Late Wisconsin. However, the evidence is too meager to lend much strength to this second hypothesis. The first possibility seems the best explanation for the available data.
Figure 20. Moraine "II" extending into southwestern Taylor Bay at low tide. Moraine "I" lies on the lower right. The view is to the north, up Taylor Bay toward the present Brady Glacier terminus.

Figure 21. Cavernously weathered granodiorite boulders similar to those found on the surface of moraines "I" and "II" in southwestern Taylor Bay. These particular boulders, however, lie on the coast southeast of Lituya Bay, and were probably derived from an Early Wisconsin moraine (Derksen, 1974).
Lowering of Firm Line During the Maximum of Early Neoglacialion

Several methods for estimating the height of a former firm line are summarized by Flint (1971) and Andrews (1975). They include: (1) lowest elevation of the floor of north-facing cirques formed at a particular time, (2) an accumulation area ratio (AAR) technique applied to a reconstructed glacier, and (3) the maximum elevation of lateral moraines formed by a given glacial advance.

First, in the Brady Glacier region there are only six abandoned cirques out of 154 examined in air photos that bear deposits of possible Neoglaclial age. The mean elevation of these cirque floors is 427 m, but the standard deviation is 152 m. The lowest of these oriented in a northerly direction is at 336 m elevation. Perhaps more significant, out of a population of 65 abandoned cirques oriented between 300° and 60° north, most have relief of over 100 m and probably date from Wisconsin time. The 13 cirques with relief less than 100 m tend to occur at higher elevations with a peak between 503 ± 46 m elevation (Figure 22). This elevation might reflect an average position of Early Neoglaclial firm line.

Second, the assumptions and sources of error inherent in an accumulation area ratio technique applied to a reconstructed glacier are fully discussed by Porter (1970, 1975). Basically, it assumes: (1) that the glacier's margins and topography are known in detail for a particular time, (2) that the mass balance of the glacier was in equilibrium at some former time was like that of modern glaciers in equilibrium within
Figure 22. The vertical distribution of cirque floor elevations of 65 cirques oriented between 300° and 60° North in the Brady Glacier region is shown by the upper histogram. The distribution of the 13 of these which have less than 100 m of relief and probably formed since Wisconsin Glaciation is shown by the lower, shaded histogram.
the same region. (See also the discussion of this technique in the section "Reconnaissance Glaciology" of this report.) Reconstructing the topography of the Brady at its Neoglacial maximum position (Figure 23) suggests that the former firm line was 572 ± 38 m elevation.

Finally, Early Neoglacial firm line must have been at a somewhat higher elevation than the highest lateral moraines deposited at this time. The highest lateral moraines observed are at an elevation of about 430 m on the eastern side of Taylor Valley and at about 520 m along the western margin.

In conclusion, three lines of evidence indicate that the Early Neoglacial climax of the Brady Glacier came into equilibrium with a firm line at about 534 ± 77 m elevation. Surprisingly, this indicates a maximum lowering of firm line of less than 123 m from its present height near 580 m (Bengtson, 1962).

Late Neoglaciación

Post-Early Neoglacial Ice Recession and Marine Transgression

Following the Early Neoglacial advances of the Brady, marked recession occurred as the glacier shrank to less than its present volume prior to 685 B.P. (Bengtson, 1962). During the interval from about 1200 to 700 B.P., Taylor Bay must have lengthened considerably as Brady ice receded. Bengtson (1962) discovered stumps and logs of lowland
Figure 23. Reconstructed margins and topography of the Brady Glacier during the maximum phase of Early Neoglacialiation. Topography is shown from sea level to 1000 m elevation which is more than sufficient for determining the equilibrium line altitude from accumulation area ratios.
RECONSTRUCTED
EARLY NEOGlacIAL
MAXIMUM OF THE BRADY GLACIER
(Topography Shown 0-1000 m)

Modern
Ice Limits

Estimated
ELA Zone

Drainage
Divide

CONTOUR INTERVAL 100 m
coniferous trees above present treeline incorporated in Late Neoglacial till at a point 24 km north of the present terminus. This led Bengtson to conclude that the Brady must have shrunk dramatically, allowing a lowland forest to grow far up the shores of a lengthened Taylor Bay. Readvancing Late Neoglacial ice destroyed the forest about 685 ± 40 B.P. \((UW-14)\) and carried its remnants to their present site downglacier from their place of growth.

Marine waters also spread far up the Dixon River Valley during this period of ice recession. Exposures of shell-bearing, compact, gray silty clay occur sporadically up the Dixon River Valley a few meters above river level from just north of the Early Neoglacial terminal moraines near Dixon Harbor to a point several kilometers north between the Dixon and Deception Lobes of the Brady Glacier (Plate 1). Shells of the shallow marine mollusc *Macoma balthica* are found in these deposits as well as in modern silty, inter- to subtidal environments throughout southeastern Alaska (identification by Dr. W. Zinsmeister, Institute of Polar Studies, and Mr. David Duggins of the University of Washington). In the Dixon River Valley, these raised marine deposits are undeformed and are conformably overlain by a meter or so of brown alluvial sand. Near the Dixon Lobe Bengtson (1962) discovered a stump layer in the overlying alluvium dated at 433 ± 80 B.P. \((UW-21)\). Further downriver, worm-drilled marine driftwood at the clay-sand interface gave an age of 340 ± 100 B.P. \((DIC-287)\). Thus, the Dixon River Valley marine invasion occurred following recession of Early Neoglacial ice from terminal moraines at the lower end of the valley, and then gradually regressed due to encroachment of proglacial outwash deposited as Late Neoglacial Brady ice readvanced
into the valley.

Climax of the Late Neoglacial "Little Ice Age" Advance

The Late Neoglacial period of glacial advance has been termed the "Little Ice Age" by other workers in this region (Bengtson, 1962; Goldthwait, 1963) after the informal usage by Matthes for the most recent episode of minor glacier activity in the North American Cordillera. This glacial advance built many small moraines (Figures 24, 26) and out still visible trimlines (Figure 25) in many places along the glacier's margins. Bengtson (1962) dated the climax of this advance at 1876 A.D. on the basis of an ice-damaged hemlock at a point 11 km north of the present terminus on the eastern margin of the glacier (Figures 26, 27). This tree was re-examined in 1975 and Bengtson's date confirmed by tree-ring count. Ice-damaged trees that are still alive after the passage of a century are quite rare. However, a spruce found along the Little Ice Age trimline of the Deception Lobe (Figure 25) was severely bent and damaged in the year 1888 A.D. Thus, after beginning to advance in the 13th Century, the Brady reached its maximum during the last quarter of the 19th.

The Canadian Boundary Survey mapped the Glacier Bay region in the year 1894. Their Sheet No. 15 (Alaskan Boundary Tribunal, 1904) shows the Brady Glacier terminus in nearly its present position. O.J. Klotz, of the Canadian Survey, compared this map with a much older map (Figure 28) made by Capt. George Vancouver during his visit in July, 1794, and
Figure 24. The arrow indicates a small "Little Ice Age" moraine left by the Boussole Lobe of the Brady Glacier in the upper Dixon River Valley. View is to the southwest.

Figure 25. Trimlines cut by the "Little Ice Age" advance of the Deception Lobe of the Brady Glacier shown by arrows. The Dixon River lies on the right. The middle and left arrows indicate the trimline cut by ice-dammed North Deception Lake, formed as the Deception Lobe advanced to the bedrock spur located at the middle arrow. View to the east.
Figure 26. A prominent "Little Ice Age" moraine that damaged a still-living hemlock at a point 11 km north of the Brady Terminus on the eastern side of the glacier. Note the ice-thinning that has taken place since the deposition of this moraine.

Figure 27. A close-up of the ice-damaged hemlock shown in Fig. 26. Coring of the boulder-scarred area visible in this photo indicated the moraine was being deposited about 1876 A.D.
concluded that the glacier had advanced 5 miles (8 km) in the intervening century (Klotz, 1899, 1907). Since ice in nearby Glacier Bay was undergoing marked recession from its Little Ice Age maximum during the same interval (Lawrence, 1958; Goldthwait, 1963), the persistent advance by the Brady Glacier deduced by Klotz received some attention in scientific literature (Bengtson, 1962; M.H. Miller, 1964b; Amer. Geogr. Soc., 1966) and popular accounts (Bohn, 1967). However, there are three lines of evidence that suggest that this remarkable advance did not take place; that in fact, the Brady terminus in 1794 was already in about its present position.

First, Vancouver's chart does roughly reproduce the regional coastline but is greatly distorted in detail. His ships only spent a few days in the area and Vancouver himself did not actually visit Taylor Bay; the mapping was done by his lieutenant, Whidbey, on a stormy day and the final chart later compiled from his observations. Following is an excerpt from Vancouver's (1798) narrative of Whidbey's explorations near the Brady terminus:

Having at length effected this object (Cape Spencer), the continental shore from the cape above-mentioned was found to take a nearly north direction for about three leagues to a low pebbly point; N.N.W. from which, five miles further, a small brook flowed into the sound... To reach this station, the party had advanced up an arm (Taylor Bay) about six miles wide at its entrance, but which had decreased to about half that width, and their further progress was now stopped by an immense body of compact perpendicular ice (the Brady Glacier), extending from shore to shore, and connected with a range of lofty mountains that formed the head of the arm, and... gave support to this body of ice on each side. Their course was now directed across the arm, and on its eastern side... (the) shores are composed of a border of low land, which on high tide is overflowed, and becomes broken into islands... (This low land was) a few miles below the icy barrier...
The "low pebbly point" mentioned in the narrative is almost certainly terminal moraine "II" which extends quite some distance into southwestern Taylor Bay at low tide (Figure 20, Plate 1). The narrative indicates that the 1794 terminus position was five miles (8 km) north-northwest of this point. This is exactly the distance between the modern terminus and terminal moraine "II". Furthermore, the low land "a few miles below the icy barrier" can only be the low, marshy ground between Taylor and southwestern Dundas Bays (Plate 1). Raised marine deposits of recent age on this low ground suggest that it may well have been partially intertidal at the time of Vancouver's visit. Thus, Vancouver's descriptive account of this region in 1794 suggests that the Brady Glacier had already advanced to about its present position.

Secondly, John Muir (1915) in an account of his visit to the Brady terminus in 1880 describes a small outwash plain in front of the glacier (not mentioned in Vancouver's narrative) and the presence of a few stumps well out on the outwash "showing that its present bare, raw condition was not the condition of 50 or a hundred years ago". He continues, "In front of this (eastern) part of the glacier (terminus) there is a small moraine lake about half a mile in length, around the margin of which are a considerable number of trees standing knee-deep, and of course dead". Since the Brady in 1880 was near its Late Neoglacial maximum, if Klotz was correct the ice must have: (1) advanced about 8 km into Taylor Bay, (2) built up a stable outwash plain, and (3) trees became established on the outwash surface, all in just 82 years. This time period seems much too short for this magnitude of change, even making allowance for surge behavior on the part of the Brady.
Finally, the Late Neoglacial advance of the Brady greatly enlarged ice-dammed North Deception Lake (Figure 25, Plate 1). Rising lake waters, blocked by the advancing Deception Lobe, killed trees about 5 m below the highest level reached by the lake 300 ± 105 radiocarbon years B.P. (DIC-556). Thus, the minimum date for the death of these trees is 1755 A.D. This implies that the Deception Lobe (several kilometers below Klotz's deduced 1794 position of the main Brady terminus) must have been even more extended than at present at least 39 years before Vancouver's explorations.

In conclusion, the Late Neoglacial growth of the Brady Glacier, which began in the 13th Century, reached its present proportions by about 1650 ± 105 A.D., then swelled to a maximum about a century ago. Klotz's deductions from Vancouver's chart of rapid advance in the 19th Century are probably in error.

In the upper Palma Valley, Brady ice merged with ice from the tributary Palma Glacier to deposit a Little Ice Age moraine prior to the late 19th Century climax reached by the main body of the Brady Glacier. The age and composition of the vegetation on this moraine suggest it dates from the mid-1700's (Streveler and Worley, pers. comm., 1974).
Recent Events

Ice Recession and Readvance

Since the Late Neoglacial climax of the Brady nearly a century ago, its behavior has been characterized by slow, halting recession and thinning, documented by maps made in 1894, 1907, and 1948, and photographs taken in 1894, 1929, 1948, 1958, 1961, 1963, 1964, 1966, 1973, 1974, and 1975. During Muir's visit in 1880 the terminus had been advancing, showing up a push-moraine, but this advance halted prior to 1886 (Young, 1915). On the basis of the 1:160,000 scale map made in 1894 (Alaska Boundary Tribunal, 1904), the 1:250,000 scale map made in 1907 (Inter. Boundary Comm., 1952), and the aerial photographs taken by the U.S. Navy and U.S. Geological Survey in 1929, a minor, 0.6 km readvance of the terminus took place during the interval 1894 – 1929, which must have climaxed shortly before 1929. Rapid recession occurred between 1929 and 1948, but from 1948 to 1958 the terminus was essentially stagnant and only thinning of the snout took place (M.M. Miller, 1964b). The terminus began to advance again in 1961 (Bengtson, 1962), and has continued an advance of about 0.3 km up to the present time.

Figure 26 shows a recession curve derived from the Deception Lobe of the Brady Glacier on the basis of tree-ring dated minor moraines and vertical aerial photographs taken in 1929 and 1948. Recession here also appears to have been most rapid during the interval 1928 – 1948. The Deception Lobe also began to readvance within the past few years, and
Figure 28. Recession of the Deception Lobe of the Brady Glacier from 1888 to 1975 based on dendrochronologic and air photo data. This lobe has recently begun to advance and in 1975 was crushing trees that began to grow in 1967.
in 1975 was destroying alders that began to grow in 1967. The general character of the curve shown in Figure 28 is similar to other recession curves drawn for glaciers in this region (Heusser and Marcus, 1964).

In the Palma Valley, a steep moraine was forming between 1889 and 1914 when large spruce trees along its margin were bent and damaged. Another spruce was damaged 51 years ago in 1925, but here also recession must have started around this time since the oldest spruce found growing on the crest of the moraine is 48 years old. Another damaged spruce suddenly showed a rapid increase in growth 44 years ago in 1932 suggesting that the glacier had retreated from the immediate area. Thus, recent history of the Palma Glacier, which is separated from the Brady by an ice-dammed lake, seems generally in phase with recent behavior of the larger ice body.

Growth of Outwash Into Taylor Bay

The areal growth of supratidal outwash into Taylor Bay over the period 1894 - 1973 is shown by Figure 29. These data are derived from maps made in 1894, 1907, and 1948, and aerial photographs taken in 1929, 1948, 1963, and 1973. The approximate variation in area due to a mean tidal change of 3.2 m on a shoreline slope of 1.2% is also indicated (data from navigation charts; NOAA, 1974). A linear regression of the measured area increase since 1894 is plotted as a "best-fit" line. This line's extension to the x-axis suggests that, if the rate of outwash growth has been constant, the outwash plain must have made its first
Figure 29. Areal growth of outwash into Taylor Bay, 1894 - 1973. The straight, heavy line is best-fit to the outwash areas measured from maps and air photos with no tidal correction applied. The thin, vertical bar indicates the probable variation in area at a given time due to tidal fluctuation. Thus, the broken, thin line indicates the probable variations in growth rate during this interval relative to mean high tide.
visible appearance in Taylor Bay during the 1860's. This agrees with observations of the terminus made by Vancouver in 1794 ("compact perpendicular ice" at the head of the Bay; presumably an ice cliff calving into seawater). By the time of John Muir's visit in 1880, the outwash area should have been more than 4 km$^2$. This also seems to agree with Muir and Young's written descriptions of the size of the outwash plain (Muir, 1909; Young, 1915). If this growth rate of about 0.34 km$^2$/yr continues, all Taylor Bay will be dry land by the early part of the next century. However, these data do suggest some variation in growth rate; the rate appears to have increased somewhat during the interval 1929 - 1948. This corresponds to a period of generally warmer temperatures in this region (Marcus, 1964; M.M. Miller, 1975), and to the period of greatest recession of the main Brady terminus, the Deception Lobe, and the Palma Glacier.

The volume of outwash present in Taylor Bay and its rate of growth is more uncertain, but making reasonable estimates as to the slope of the outwash plain both above and below sea level (0.4° and 0.7°), and to the former average depth of Taylor "Fiord" (about 60 m; all data from navigation chart, NOAA, 1974), the total volume of outwash present in 1973 was about 1.1 km$^3$. The rate of volume growth was calculated at about 6.0 x 10$^6$ m$^3$/yr in the most uniform part of Taylor Valley between 1894 and 1929. A similar rate of 5.4 x 10$^6$ m$^3$/yr was determined for an outwash delta being deposited in front of the North Grillon Glacier (Goldthwait et al., 1963). If this rate of volume growth has remained fairly constant, the entire volume of outwash present could have just been deposited since the time of Vancouver's visit in 1794. This also
seems in agreement with the Brady terminus advancing to near its present position by this date (cf. page 46).

**Dumping Lakes**

Post and Mayo (1971) noted the presence of 19 ice-dammed lakes larger than 0.1 km$^2$ around the margins of the Brady Glacier. Draining of glacier-dammed lakes can occur suddenly in the form of an often catastrophic outburst flood; thus the term "dumping lakes". All of the lakes in the mapped area of Plate 1 discharge into known or inferred glacier outburst flood courses according to Post and Mayo (1971). Using Post's terminology (pers. comm., 1975), a lake between the Brady and the Palma Glacier ("Bearhole Lake"), a lake dammed just west of the Dixon Lobe ("Dixon Lake"), and the two Trick Lakes are the largest lakes with a known history of abrupt drainage (Plate 1). The Bearhole Lake probably drains subglacially under the Palma Glacier and down the Palma River. The Dixon Lake drains subglacially under the Brady and escapes down the Dixon Lobe to empty into the Dixon River (Mitchell, 1960). North Trick Lake drains subglacially into South Trick Lake which drains underneath the Brady into Taylor Bay, or, if the lake level rises close to 90 m, can overflow down Annoksek Creek into Murk Bay.

The Bearhole Lake (Figure 30) must have flooded the Palma Valley some years before 1948, since aerial photographs taken at that time show that the valley had recently been stripped of much of its vegetation.
Figure 30. The Bearhole Lake trapped between the Brady and Palma Glaciers as it appeared in August, 1974.

Figure 31. Trimlines of the 1918 flood of the Bearhole Lake down the Palma Valley. The trees that stood above the flood are mostly spruce; the area denuded by the flood is today covered by dense alder and young spruce thickets.
These flood trimlines were still visible in 1974 (Figure 31). Streveler (pers. comm., 1974) tree-ring dated the vegetation that grew up on this denuded surface and estimated that this catastrophic flood occurred about 1918. This corresponds to the period of moraine formation in the upper part of Palma Valley (ca. 1889 – 1925) suggesting that the advancing ice had substantially increased the size of the lake. Patches of vegetation that escaped this flood are still of relatively youthful age indicating that outburst floods from this lake are a fairly common occurrence.

The Dixon Lake must have flooded the Dixon River Valley shortly before 1894 since photographs taken in that year show the valley floor stripped clean of vegetation and a sharp trimline cut into the forest on the valley sides. However, the oldest trees found growing below the trimline in the formerly denuded area near the Early Neoglacial moraines at the end of the valley were 55 year-old alders, suggesting the passage of yet another flood about 1920. Mitchell (1960) observed the drainage of the Dixon Lake between Sept. 7th and 9th, 1960. The lake had been about 0.4 km\(^2\) in size and extended quite some distance under the glacier's surface along its margins before abruptly emptying almost completely in a two-day period.

The Trick Lakes were both higher some time before 1948; numerous strandlines are evident above lake level in photos taken that year. North Trick Lake appears to have been quite high in a 1929 photo. These lakes continued to lower until about 1964, but by 1973 they had risen to their highest level since the 1920's. South Trick Lake began overflowing down Annoksek Creek in 1974. Periods of high level of these
Recovery From Neoglacial Isostatic Depression

Whidbey, in Vancouver (1798) reports that:

(Between Cape Spencer and Frederick Sound were seen) several places where the ocean was evidently encroaching very rapidly on the land, and that the low borders extending from the base of the mountains to the seaside had, at no very remote period of time, produced tall and stately timber, as many of their dead trunks were found standing erect... the shorter stumps in some instances at low water mark were even with or below the surface of the sea.

Was Whidbey observing the effects of recent faulting, or could the land in this region have sunk due to its increased burden of Neoglacial ice? There is considerable evidence that the depression Whidbey observed was due to the latter, since isostatic recovery is taking place in the Brady Glacier region today.

First, Pierce (1961) and Hicks and Shofnos (1965) studied tidal gauge records for this area of southeastern Alaska stretching back as far as 1687 and concluded that: (1) the land in this region was steadily rising, and (2) the rates of uplift formed a concentric pattern with the highest uplift centered on Glacier Bay (Figure 32). Hicks and Shofnos' data indicate that the Brady Glacier area is rising at rates between 1.9 and 2.3 cm/yr. Goldthwait, from a study of young raised marine beaches in Glacier Bay from Bartlett Cove to Muir Glacier, deduced a similar pattern of uplift, but with the zone of greatest uplift occurring further up the Bay (pers. comm., 1976). This concentric pattern of regional uplift is consistent with the hypothesis of Neoglacial isostatic depression followed by post-Neoglacial isostatic recovery.
Figure 32. Figure 1 from Hicks and Shofnos' (1965) paper showing concentric rates of uplift centered on Glacier Bay.

Figure 33. An ancient marine sea cliff near Icy Point at an elevation of 40 m at the base, recently re-eroded by wave action due to Neoglacial isostatic depression of the region. Note the youth of the forest at its base.
Second, numerous raised marine and shoreline deposits including beach sediments, sand dunes, and recent erosion of coastal portions of ancient sea cliffs (Figure 33) is widespread in the Brady Glacier region. These features occur on both sides of the Fairweather Fault from Icy Point to Dundas Bay at elevations up to about 14 m above mean sea level (Plate 1). In eastern Dundas Bay, a forest growing near the base of an ancient sea cliff was submerged and buried by marine sediments about 1960 ± 65 radiocarbon years B.P. (DIC-461). This corresponds to the beginning of the maximum Early Neoglacial advance of the Brady Glacier, dated at 1960 ± 90 B.P. (DIC-458), and ice advance in nearby Glacier Bay (Bengtson, 1962; Goldthwait, 1963, 1966; McKenzie, 1970; McKenzie and Goldthwait, 1971). Today, the remnants of this forest still lie in the intertidal zone. A horizontal log in a raised beach deposit in western Taylor Bay was dated at 1030 ± B.P. (DIC-554), indicating that the land was still being depressed at this time, probably by the growing mass of ice in the Glacier Bay region. A small marine terrace, 4-5 m above mean high tide, also formed in the Icy Strait area due to isostatic depression of the Glacier Bay region from 3,185 ± 100 B.P. to 850 ± 50 B.P. Renewed depression occurred during the Late Neoglacial, forming a second terrace 0-1 m below the first (T.D. Hamilton, manuscript in preparation, 1976).

Uplift of all these Neoglacial marine deposits must have begun within the past one and one-half centuries, since the forest growing on their surface is quite young. Very few dead trees litter the ground, forest duff is only a few centimeters thick over little-weathered sand, and most trees are no older than about 80 years (although one 150 year-old spruce
was found growing on the lee/side of forested sand dunes in Astrolabe Bay. Marine clay in the lower Dixon River Valley at an elevation of 3.6 m above mean sea level contains driftwood in its upper part dated at 340 ± 100 B.P. (DIC-287). The high amount of clay in this deposit (66%) and negligible sand suggest it was deposited in quiet water below the higher energy intertidal zone. Thus, uplift of this deposit must have occurred at a minimum rate between 0.8 and 1.4 cm/yr. These values seem comparable to Hicks and Shofnos' rates of 1.9 to 2.3 cm/yr derived from tidal records.

In conclusion, data from historical observations, tidal records, raised marine deposits, and dated glacial history indicate that this region was isostatically depressed by a growing load of Neoglacial ice. It has only begun to rise from the sea since the close of the Little Ice Age and the onset of over 100 km of ice recession in Glacier Bay.

Summary of Holocene Glacial History and Comparison With Other Areas

A simplified interpretation of the history of the Brady Glacier from about 12,000 B.P. to the present is illustrated by Figure 34. The 12 radiocarbon dates which define the chronology of this history are indicated by appropriate symbols. Although events in the Brady Glacier area are only dated with any certainty back to about 4700 B.P., the older part of this curve can be reconstructed fairly reasonably on the basis of the generally similar, radiocarbon-dated glacial history derived from numerous other sites in southeastern Alaska and British
Figure 34. A simplified interpretation of the history of the Brady Glacier from about 12,000 B.P. to the present. The dashed line is based on glacial events dated in Glacier Bay and elsewhere in southeast Alaska. The radiocarbon dates indicated refer specifically to fluctuations of Brady ice except for sample Y-9 which was deposited by ice draining north from the Brady Icefield via the Reid Glacier.

Sources: Bengtson, 1962; Derksen, 1975, this report.

The first, independently-dated Early Neoglacial advance of the Brady around 4700 B.P. correlates with a period of glacial advance noted in many alpine areas of the world between about 5800 and 4900 B.P. (Denton and Karlen, 1973; Porter, 1974), although ice did not begin advancing into the upper part of Glacier Bay until about 4150 B.P. (Goldthwait, 1966). The extended position of the Brady Glacier indicated by the ca. 2900 B.P. date from glaciomarine sediments near Dundas Bay also correlates with a well-established period of glacial advance in many alpine areas between about 3500 and 2400 B.P. (Denton and Karlen, 1973; Porter, 1974), and in Glacier Bay between 4150 and 2200 B.P. (Goldthwait, 1966, 1974).

The best-dated part of the curve, the climax of Early Neoglacialiation of the Brady between about 1960 B.P. and 1230 B.P. correlates well with an advance of the nearby Geike Glacier dated at 1520 ± 140 B.P. and 1540 ± 130 B.P. (Bengtson, 1962), and with five dates between 1770 ± 100 B.P. and 1535 ± 100 B.P. from the Neoglacial Adams Formation in Glacier Bay (McKenzie, 1970; McKenzie and Goldthwait, 1971). However, this is not a well-known period of glacial advance from other parts of the world, in fact, the period 2050 to 1270 B.P. is associated with glacier recession and a rise in treeline on the lee side of the St. Elias Range to the north (Denton and Karlen, 1973).
Evidence of ice recession between about 1100 to 700 B.P. in this region comes from Adams Inlet in Glacier Bay (McKenzie, 1970; McKenzie and Goldthwait, 1971), the Juneau Icefield (M.M. Miller, 1975), and the northern St. Elias Range (Denton and Karlen, 1973). Thus, the deduced, dramatic recession of the Brady during this interval also fits into a regional context.

The most recent major advance of Brady ice, from about 700 B.P. to the latter part of the 19th Century, corresponds to very well known glacier advances in Glacier Bay and many other alpine areas around the world that occurred during the global cooling of the "Little Ice Age" (Denton and Karlen, 1973; Flint, 1971; Goldthwait, 1963, 1966, 1974; McKenzie and Goldthwait, 1971; M.M. Miller, 1975; Porter, 1974). This Little Ice Age was followed by ice recession to about the present.

Thus, the independently-dated history of fluctuations of the Brady Glacier seems to fit into the general pattern of glacial events in this region and other alpine areas. The present readvance of the Brady and several other glaciers fed by the high peaks of the Fairweather Range (Johns Hopkins, Lamplugh, Margerie, and Grand Pacific Glaciers; G. Haselton, pers. comm., 1976) indicates that climatic conditions have recently become more favorable for glacier growth in this region. This may correspond to a general decline in global and regional temperatures since about 1945 (L.J. Mitchell, 1961; Marcus, 1964; M.M. Miller, 1975).
Aside from meteorological observations at a high camp on the glacier during one summer (Bengtson, 1962), little glaciological data exists for the Brady Glacier. A series of aerial photos have been taken in various years since 1894, and meteorological records from Cape Spencer exist for the period 1939 - 1974, when the station was abandoned. However, it is possible to gain at least some idea as to the present mass balance of the glacier, its sensitivity to climatic change, and to speculate on its future behavior.

**Determination of the Modern Equilibrium Line Altitude**

The firm line on the Brady Glacier in 1950, 1960, and 1961 was about 580 m (Bengtson, 1962), although aerial photographs taken late in the ablation seasons of 1948, 1963, 1964, 1969, 1973 and 1974 indicate it has fluctuated through a range of 90 m since 1948 (between 520 and 610 m elevation). For a temperate glacier in equilibrium with its mass balance, the firm line and equilibrium line will nearly coincide (Paterson, 1969). Since the Brady has changed little over the past century, Bengtson concluded that the equilibrium line altitude
(ELA) of the Brady must also be around 580 m. This conclusion can be checked by two different techniques if the vertical distribution of glacier surface area is known.

Figure 35 shows 90% of the surface area distribution of the Brady Glacier. The remaining 10% is spread out in small increments above this to 3565 m elevation. Comparison with a map of the Brady (Figure 1) indicates that about 70% of the glacier's accumulation area (from firm line up to 1065 m elevation) is made up by the gently sloping surface of the Brady Icefield itself. Furthermore, nearly 16% of the total glacier area lies within ± 90 m of the assumed ELA at 580 m. This diagram shows that any persistent change in firm line elevation has potentially great effect on the mass balance of this glacier.

A crude technique for estimating the equilibrium line altitude from known distribution of glacier surface area was outlined by Mercer (1961a). This method assumes that rates of accumulation and ablation are a linear function of elevation above or below the equilibrium line. Ablation is considered to increase in the ratio 1 : 3 : 5 : 7 and so on with 500 ft (152 m) increments of decreasing elevation from the ELA; accumulation is considered to increase in the ratio 2 : 5 : 7 : 9 and so on with 500 ft (152 m) increments of increasing elevation from the ELA. Within the first increment of the accumulation area, 152 m above the equilibrium line, there is a change from a simple linear relation due to differences in the surface albedo of this zone. By multiplication of glacier areas contained in vertical increments of 500 ft (152 m) times the corresponding ratios relative to some assumed ELA, a "Mass Balance Index" can be calculated by the following equation:
Figure 35. A histogram of 90% of the surface area of the Brady Glacier from near sea level to 1372 meters elevation. The remaining 10% of the glacier's surface is spread out between 1372 and 3565 meters elevation.
\[ M.B.I. = \frac{\sum A}{B + \sum A} \]

where \( A \) equals the values obtained for the area of the accumulation zone, and \( B \) equals the values obtained for the area of the ablation zone. A Mass Balance Index (M.B.I.) of 0.50 should correspond to a glacier in equilibrium. If the glacier is known to be in equilibrium and a M.B.I. value of greater than 0.50 is obtained, then the assumed elevation of the equilibrium line must be too low and the data must be recalculated with a slightly higher ELA. By trial and error, the approximate elevation of the glacier's equilibrium line may be found within ± 250 ft (± 76 m).

Mercer found that this technique worked reasonably well with glaciers in the Gulf of Alaska region. However, in the case of the Brady, fed in part by snows from the high peaks of the Fairweather Range, the technique is complicated by a decrease in snowfall above 2,000 m elevation (Marcus and Ragle, 1970). However, this effects only 8.5% of the accumulation area, and effects the calculated result by only an additional ± 5%. Applying this method to the Brady Glacier, an assumed ELA of 457 m gave a M.B.I. of 0.70 ± 0.01, and an assumed ELA of 610 m gave a M.B.I. value of 0.44 ± 0.02. Therefore, the elevation of the equilibrium line on the Brady Glacier in 1948 (when the photography for the topographic map was made) was about 533 ± 76 m. This value agrees favorably with Bengtsson's estimate of around 580 m from more recent firm line observations.

Another reconnaissance technique for estimating the equilibrium line altitude from known distribution of glacier surface area with
respect to elevation is by calculation of an accumulation area ratio (AAR). Meier (1962) defined accumulation area ratio by the following relation:

\[
AAR = \frac{\text{Glacier Accumulation Area}}{\text{Total Glacier Surface Area}}
\]

These areas are easily measured by planimeter from an accurate topographic map of a glacier. Figure 36 illustrates accumulation area ratios calculated for the Brady Glacier given various hypothetical elevations of firm line (which delimits the accumulation area) from sea level to 3,500 m. Meier and Post (1962) calculated the AAR's for glaciers in the northwestern part of North America and values ranged from 0.5 to 0.8, but found that values greater than 0.6 indicated a "healthy" mass balance for glaciers in the Gulf of Alaska region. Glen (1963) suggested an AAR of 0.7 for a glacier in steady-state equilibrium was a good "rule of thumb". Porter (1963) found equilibrium values of 0.68 for two Pleistocene glaciers in Washington, and Andrews (1975) suggests that equilibrium for most glaciers occurs at AAR values in the range 0.6 - 0.7. Therefore, using an AAR of 0.65 ± 0.5 to define equilibrium mass balance on the Brady Glacier, the curve in Figure 36 may be used to indicate the elevation of the glacier's equilibrium line. This curve suggests the 1948 ELA of the Brady Glacier was about 595 ± 75 m which further lends support to Bengtson's more recent estimate of 580 m.

The effects of a change in firm line from the vicinity of the ELA can also be seen in Figure 36. Relative to a firm line at 580 m, a
Figure 36. Accumulation area ratios based on the 1948 surface area distribution of the Brady Glacier. Since the glacier has been in near equilibrium for almost a century, AAR values between 0.6 and 0.7 suggest that the equilibrium line altitude of the glacier lies between 520 and 670 meters.
change of ±150 m could cause the AAR to vary between 0.54 and 0.79, resulting in either a markedly negative or positive glacier mass balance. Using a July environmental lapse rate of 0.53°C/100 m at Juneau Airport as representative (Marcus, 1964), this change in mass balance could be accomplished by a change in average ablation season temperature of only 0.8°C. For comparison, "Little Ice Age" climate was estimated to have been 2°C cooler than at present in this region (Goldthwait, 1966).

In conclusion, the glacier's present equilibrium line altitude lies between 457 and 670 m elevation and is probably close to about 580 m. The Brady is quite sensitive to changes in firm line because so much of its area lies near this critical elevation. The maximum Early Neoglacial advance of the Brady came into equilibrium with an average firm line elevation of less than 123 m lower than that of the present (Chapter III).

Glacial Stability

Fiord glaciers are characteristically unstable because of: (1) the effect of variations in fiord width on glacier terminus positions, and (2) the effect of deep water at the glacier terminus (Mercer, 1961b; Post, 1975). Fiord glaciers terminate where the rate of ice discharge can keep pace with the high rate of ice loss due to calving into the fiord. Thus, fiord glaciers are subject to major adjustments in size due to relatively minor climatic fluctuations. However, the presence
of any sort of supportive shoal at the snout can stabilize the glacier by greatly reducing ice wastage due to ice calving into deep water (Goldthwait et al., 1963; Post, 1975). Figure 37 illustrates the contrast between the Brady Glacier profile during the height of the "Little Ice Age" advance in the late 1800's and its modern profile within the ablation zone. In the upper parts of the glacier, as much as 56 m of thinning along the glacier's margins has occurred, but in 1975 the Brady terminus had locally exceeded the Late Neoglacial maximum position.

Bengtson (1962) postulated that the growth of the outwash plain into Taylor Bay (Chapter III) stabilized the glacier, allowing it to remain in near-equilibrium since the Late Neoglacial advance. The growth of the outwash allowed the glacier to achieve a lower gradient profile. Thus, the terminus may now advance further without being affected by even the highest storm tides.

The stabilizing effect of the growth of the Brady outwash plain on the glacier's behavior may also explain the contrast between recent history of the Brady Glacier and events in nearby Glacier Bay. The Brady had advanced to about its present limits by 1650 ± 105 A.D., but reached its maximum in the late 1800's and changed little since. In contrast, ice in Glacier Bay reached its maximum around 1750 A.D., but lacking a major supportive shoal at the terminus, it probably became unstable and has receded over 100 km up the Bay in the past two centuries.

At present, the Brady Glacier is not subject to calving at the terminus and can maintain a very large accumulation area because of the size of the glacier itself (Figure 35). If, for some reason the Brady were to recede from the outwash at its snout into deep water again, catastrophic
Figure 37. Trimline elevations along the western flank of the Brady Glacier. Data are derived mostly from aerial photography, therefore trimline elevations are only approximate.
recession of the ice front might occur here as well. The glacier would recede until the accumulation area could keep pace with the increased rate of ice loss at the terminus, even if climatic conditions remained constant.

**Future Glacier Behavior**

Climatic conditions in southeastern Alaska were generally unfavorable for widespread snow accumulation during the period 1925-1945, but have become increasingly more favorable since that time (Mercer, 1961b; Marcus, 1964; M.M. Miller, 1975). Ten-year running means of average summer temperature (June-Sept.) and winter precipitation (Nov.-April) for the five coastal weather stations nearest the Brady Glacier are shown in Figures 38 and 39. Yakutat, Cape Spencer, and Sitka are all located along the coast of the Gulf of Alaska within 160 km of the Brady, and to have similar climatic trends. Records from Juneau, located 125 km from the coast of the Gulf, are more variable and probably reflect the greater continentality of its climate. These data are only a crude indication of mass balance trends of the Brady Glacier, but Marcus (1964) showed that correlation of records from several weather stations in the Juneau region with isolated camps on the Juneau Icefield was possible. All the stations nearest the Brady experienced a downward trend in ablation season temperature from about 1940 to the present, yet winter precipitation remained fairly high even during "dryer" periods. Thus, the present Brady advance, beginning in 1961, is not surprising.
Figure 38. Ten-year running means of average summer temperature (June - Sept.) for five coastal stations nearest the Brady Glacier over periods for which published data is available.
Figure 39. Ten-year running means of winter precipitation (Nov. - April) for five coastal stations nearest the Brady Glacier over periods for which published data is available.
A rate of surface flow of 0.18 km/yr was determined for the central part of the Brady's ablation zone by the progress downglacier of conspicuous dirt bands in air photos over an 11-year interval. Since variations in mass balance of the glacier are propagated down the glacier from the equilibrium line as kinematic waves of increased or decreased flow which move at approximately four times ice velocity (Paterson, 1969), the main response of the Brady terminus to a sudden change in mass balance should lag behind the event by a minimum of 30 years. This approximation seems reasonable given the generally uniform shape of the Brady Glacier from the equilibrium line 21 km south to the terminus. However, diffusion of kinematic waves may greatly prolong the time required for the glacier to come into equilibrium with the new mass balance. Thus, for harmonic changes in mass balance with periods of less than about 600 years, the maxima and minima of terminal position lag behind maxima and minima of mass balance by roughly one-quarter of the period (Paterson, 1969). M.M. Miller (1975) and Willett (1976) cite evidence based on sunspot cycles that the recent 30-year global cooling trend will continue to the end of this century. Therefore, it is reasonable to speculate that the current slow advance of the Brady may continue or even accelerate over the next few decades.
CHAPTER V
NATURE OF TILL DEPOSITS

Analytical Procedures

Eighteen till samples ranging from modern to Wisconsin in age were analyzed for grain-size distribution, quartz grain roundness of the -1 to -2 $\phi$ fraction, Trask coefficients of dispersal and skewness, percentage of heavy minerals, and clay mineralogy in order to characterize tills of different types and ages in the Brady Glacier region (Appendices B & C). Assuming that these samples were derived from a normally-distributed population, the statistical significance of data variations was calculated using the standard "Student's t-test" and significance tables for a two-tailed test (Fisher and Yates, 1953).

The grain-size composition was determined using sieve and hydrometer techniques outlined by the A.S.T.M. (1964). The fraction greater than -1 $\phi$ in diameter was calculated separately as a percentage of the total sample while the remaining sand, silt, and clay fractions are reported as totalling 100%. From cumulative percentage curves of grain-size data, coefficients of dispersal (sorting) and skewness were calculated by the equations derived by Trask (1932). Skewness is reported as the $\log_{10}$ of the Trask coefficient to indicate a coarse
grained bias (positive value) or fine-grained bias (negative value) (Pettijohn, 1957). Heavy minerals were separated from the 2 to 4 φ fractions using tetrabromoethane, specific gravity 2.96, and their percentages calculated on the basis of the total sand, silt, and clay fraction. Roundness of 20 quartz grains in each sample from the -1 to -2 φ fraction was measured under a binocular microscope using the Krumbein (1941) scale. Quartz grains were used since quartz is: (1) easily identified, (2) monomineralic, therefore of uniform hardness, and (3) common to all samples in sufficient amounts to produce statistically significant data. Similar results were obtained using an ubiquitous grain-type composed of gray, micaceous schist, but these data are considered less reliable since the mineralogy and hardness of the schist grains are variable.

Identification and semiquantitative determination of abundances of the clay minerals in the till from X-ray diffraction data were carried out using the method of Johns et al. (1954), modified by Wilding and Drees (1966, unpublished report). The same technique was used in the analysis of glacial sediments from nearby Glacier Bay by other investigators (Haselton, 1966; McKenzie, 1970; Mickelson, 1971). These data are reported as percentage of the total clay fraction and are considered accurate to about ±5%. In this calculation, the minerals illite, vermiculite, chlorite, quartz, and expandable clays (mostly montmorillonite ?) were considered to comprise the entire clay fraction even though the diffraction data indicates minor amounts of feldspar and amphibole are also present. No dolomite or calcite was identified in any of the samples, and is known to occur only very locally in the
Significance of Data

Basal and Ablation Till

Two types of till were distinguished in the field. Dense, compact, "hard" till deposits were considered basal till, while less-consolidated, more gravelly, "soft" till units (including unconsolidated, very recent till from ice-cored moraines) were classed as ablation till. In at least one locality in the Dixon River Valley, Early Neoglacial "ablation" till overlies a compact "basal" unit of probably about the same age. Fabric strength which may (Drake, 1971) or may not (Mickelson, 1971) be diagnostic of basal versus ablation till, was not measured in the field.

Tables 1 and 2 summarize the laboratory data for hard and soft tills in the Brady Glacier region. The analytic techniques used do not distinguish between these two types of till with any high degree of statistical significance. The range in these data (indicated by the relatively large standard deviations from the means) is too great and the number of sample localities too small. However, the trend of means calculated from these data is suggestive.

Ablation till is carried within or on the surface of a glacier until deposition by melting of the supporting ice near the glacier's
Table 1. Comparative analyses of basal and ablation tills from the Brady Glacier region, Alaska.

<table>
<thead>
<tr>
<th>Type</th>
<th>n</th>
<th>% &gt;-1®</th>
<th>-1 to -2®</th>
<th>% Sand</th>
<th>% Silt</th>
<th>% Clay</th>
<th>Sorting</th>
<th>Skewness</th>
<th>% Heavy Minerals</th>
</tr>
</thead>
<tbody>
<tr>
<td>&quot;Basal&quot;</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>mean</td>
<td>6</td>
<td>26.2</td>
<td>0.32</td>
<td>22.0</td>
<td>60.4</td>
<td>7.6</td>
<td>2.69</td>
<td>0.02</td>
<td>25.1</td>
</tr>
<tr>
<td>std. dev.</td>
<td></td>
<td>21.6</td>
<td>0.07</td>
<td>24.6</td>
<td>21.0</td>
<td>4.3</td>
<td>0.78</td>
<td>0.03</td>
<td>21.0</td>
</tr>
<tr>
<td>&quot;Ablation&quot;</td>
<td></td>
<td>43.8</td>
<td>0.24</td>
<td>57.8</td>
<td>37.6</td>
<td>4.6</td>
<td>3.75</td>
<td>-0.05</td>
<td>36.5</td>
</tr>
<tr>
<td>mean</td>
<td>12</td>
<td>17.4</td>
<td>0.07</td>
<td>14.4</td>
<td>11.3</td>
<td>3.4</td>
<td>1.78</td>
<td>0.27</td>
<td>14.7</td>
</tr>
<tr>
<td>std. dev.</td>
<td></td>
<td>17.4</td>
<td>0.07</td>
<td>14.4</td>
<td>11.3</td>
<td>3.4</td>
<td>1.78</td>
<td>0.27</td>
<td>14.7</td>
</tr>
</tbody>
</table>

*aPercent of total sample<br>bPercent of sand/silt/clay fraction

Table 2. Comparative clay mineralogy data for basal and ablation tills in the Brady Glacier region, Alaska.

<table>
<thead>
<tr>
<th>Type</th>
<th>n</th>
<th>% Illite</th>
<th>% Vermiculite</th>
<th>% Chlorite</th>
<th>% Quartz</th>
<th>% Expandable</th>
</tr>
</thead>
<tbody>
<tr>
<td>&quot;Basal&quot;</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>mean</td>
<td>6</td>
<td>22</td>
<td>25</td>
<td>27</td>
<td>6</td>
<td>10</td>
</tr>
<tr>
<td>std. dev.</td>
<td></td>
<td>21</td>
<td>15</td>
<td>17</td>
<td>4</td>
<td>9</td>
</tr>
<tr>
<td>&quot;Ablation&quot;</td>
<td></td>
<td>25</td>
<td>25</td>
<td>27</td>
<td>2</td>
<td>14</td>
</tr>
<tr>
<td>mean</td>
<td>9</td>
<td>25</td>
<td>25</td>
<td>27</td>
<td>2</td>
<td>14</td>
</tr>
<tr>
<td>std. dev.</td>
<td></td>
<td>21</td>
<td>14</td>
<td>14</td>
<td>7</td>
<td>16</td>
</tr>
</tbody>
</table>

*aPercent of clay fraction ± 5%
terminus. Thus, this variety of till is subject to less grinding and crushing in transport, but the supraglacial load may be washed and transported some distance by rain and meltwater. Therefore, ablation till might be expected to be: (1) coarser, (2) more angular, (3) positively skewed, and (4) better sorted than basal till.

Brady Glacier soft (ablation ?) till is generally coarser than the hard till, averaging 43.8% of the total sample greater than -1\% versus 36.2% for the hard till. Of the sand/silt/clay fraction, soft tills average 57.8% sand while the hard tills average only 32.0%. Figure 40 illustrates the mechanical composition of hard and soft tills plotted on a triangular coordinate diagram along with other fine-grained glacial sediments from the Brady Glacier region. The range of values of all the till samples is quite similar to the range of basal tills found in the upper parts of Glacier Bay (McKenzie, 1970, Figure 8; Nickelson, 1971, Figure 16). Therefore, coarseness of the soft Brady Glacier tills alone is not necessarily an indication of deposition by an ablation process. However, the occurrence of both coarse- and fine-grained ancient tills in the same local area is suggestive of different modes of deposition, particularly in the one locality where the soft will could be seen to overlie the more compact till.

Somewhat greater average roundness of quartz grains suggests that the soft tills were subjected to the action of running water, although these tills also seem more poorly sorted and more negatively skewed which tends to contradict the ablation hypothesis. Figure 41 is a plot of log\(_{10}\) skewness versus sorting for hard and soft tills and modern outwash and lake sediments in the Brady Glacier region. The soft tills are
Figure 40. Mechanical composition of tills and other glacial sediments in the Brady Glacier region.
Figure 41. Trask sorting versus log_{10} Trask skewness for hard and soft tills and modern outwash in the Brady Glacier region, Alaska.
subject to a greater range of both skewness and sorting than the hard tills. This wide range of values might reflect differences between subaerically-exposed supraglacial debris and englacial debris, either or both of which can be deposited as ablation till. More uniform values of sorting and skewness for hard tills suggests similar conditions of deposition by basal ice. Thus, the range of values for these parameters may be more significant than the calculated means of a few samples.

If all till samples are considered together, they tend to be symmetrically distributed about zero skewness and more poorly sorted than outwash or lake sediments. This supports the conclusions of Landim and Frakes (1968) that ancient tillite deposits can be distinguished from alluvial fan, outwash, and mudflow deposits partially on this basis.

In conclusion, these laboratory data do not statistically justify the field distinction between "basal" and "ablation" tills in the Brady Glacier region on the basis of compaction. However, trends in the data are suggestive of different modes of till deposition. Tills in the Brady Glacier region can be distinguished from modern fine-grained glaciofluvial deposits on the basis of lab data with considerable success.

Characteristics of Tills of Different Ages

Tables 3 and 4 indicate the results of analyses of tills of different ages in the Brady Glacier region. Only one unweathered till
Table 3. Comparative analyses of tills of different ages in the Brady Glacier region, Alaska.

<table>
<thead>
<tr>
<th>Age</th>
<th>n</th>
<th>%1-1/4</th>
<th>% Sand</th>
<th>% Silt</th>
<th>% Clay</th>
<th>Sorting</th>
<th>Skewness</th>
<th>% Heavy Minerals (2-64)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wisconsin</td>
<td>1</td>
<td>46.8</td>
<td>0.34</td>
<td>16.3</td>
<td>76.1</td>
<td>2.29</td>
<td>-0.03</td>
<td>11.1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0</td>
<td>0.06</td>
<td>13.0</td>
<td>10.9</td>
<td>2.5</td>
<td>1.24</td>
<td>0.17</td>
</tr>
<tr>
<td>Modern to Recent</td>
<td>5</td>
<td>45.3</td>
<td>0.35</td>
<td>46.7</td>
<td>46.8</td>
<td>6.2</td>
<td>4.25</td>
<td>0.10</td>
</tr>
<tr>
<td></td>
<td></td>
<td>24.2</td>
<td>0.08</td>
<td>17.0</td>
<td>13.0</td>
<td>4.5</td>
<td>2.32</td>
<td>0.30</td>
</tr>
</tbody>
</table>

*Percent of total sample

*Percent of sand/silt/clay fraction

Table 4. Comparative clay mineralogy data for till of different ages in the Brady Glacier region, Alaska.

<table>
<thead>
<tr>
<th>Age</th>
<th>n</th>
<th>% Illite</th>
<th>% Vermiculite</th>
<th>% Chlorite</th>
<th>% Quartz</th>
<th>% Expandables</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wisconsin</td>
<td>1</td>
<td>13</td>
<td>36</td>
<td>38</td>
<td>13</td>
<td>trace</td>
</tr>
<tr>
<td></td>
<td></td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>Early Neoglacial</td>
<td>10</td>
<td>16</td>
<td>28</td>
<td>30</td>
<td>2</td>
<td>12</td>
</tr>
<tr>
<td></td>
<td></td>
<td>15</td>
<td>16</td>
<td>16</td>
<td>6</td>
<td>15</td>
</tr>
<tr>
<td>Modern to Recent</td>
<td>4</td>
<td>56</td>
<td>16</td>
<td>18</td>
<td>2</td>
<td>14</td>
</tr>
<tr>
<td></td>
<td></td>
<td>9</td>
<td>4</td>
<td>3</td>
<td>5</td>
<td>9</td>
</tr>
</tbody>
</table>

*Percent of clay fraction ± 5%
sample of probably Wisconsin age was found in the lee of a roche mouton in the saddle of the ridge between South Deception Lake and the northeastern arm of Torch Bay (Plate 1). The data from this one sample are fairly similar to the range of values of Early Neoglacial and modern till. It contains less illite, heavy minerals, and sand, but more silt than younger tills, but these differences are small and probably not representative of Wisconsin tills as a whole in the region. With considerably more samples to work with, Mickelson (1971) found that he could not distinguish between Wisconsin and Neoglacial tills on the basis of mechanical composition or clay mineralogy in the Burroughs Glacier area of upper Glacier Bay. However, Haselton (1966) and McKenzie (1970), working in other parts of the same region found that they could. Differences in these parameters might exist if large Wisconsin-age glaciers transported debris to a locality from a greatly different bedrock source area than that of later ice advances. In the Brady Glacier region, however, ice flow patterns were probably not greatly different from that of the present, even during the Wisconsin.

There are no significant differences between Early Neoglacial and Modern and Recent tills except in the percentage of illite in the very young samples. The four Modern samples analyzed for clay mineralogy were collected from near the snouts of the Boussole Lobe, Dixon Lobe, Deception Lobe, and the central part of the Brady terminus in Taylor Valley, all areas underlain by micaceous schist. Thus, the higher percentages of illite in these samples probably reflects the composition of the parent rock.
In conclusion, there appear to be few significant differences between tills of different ages in the Brady Glacier region. Minor variations in till composition are probably related to bedrock sources.

Characteristics of Tills From Different Bedrock Sources

The lower part of the Brady Glacier lies between mostly granitic rocks to the east and metamorphic rocks to the west (Figure 3). Results of analyses of five till samples collected in the granitic area and eleven till samples collected in the metamorphic bedrock area are shown in Tables 5 and 6. The only statistically significant difference between tills collected in (and at least partially derived from) these two different bedrock areas is the percentage of heavy minerals in the till. Tills from the granitic area contain an average of 51\% heavy minerals in the 2 to 4\(\phi\) fraction while tills from the metamorphic zone average only about 29\%. If these two means represent normally-distributed populations, then the populations are significantly different at a probability level of 90\%. However, the composition of the suite of heavy minerals is generally similar. Garnet, biotite, iron oxides, and amphibole grains are common with apatite, pyroxene, epidote, rutile, sphene, and zircon present in minor amounts in almost all samples. There is a tendency for the metamorphic tills to contain more garnet, amphiboles, and biotite, and less iron oxide than the granitic tills, but detailed grain-counts were not made on all samples.
Table 5. Comparative analyses of tills derived from granitic and metamorphic bedrock sources in the Brady Glacier region, Alaska.

<table>
<thead>
<tr>
<th>Type</th>
<th>n</th>
<th>% -1Ø</th>
<th>Quartz Round, -1 to -2Ø</th>
<th>% Sand</th>
<th>% Silt</th>
<th>% Clay</th>
<th>Sorting</th>
<th>Trask</th>
<th>Skewness</th>
<th>% Heavy Minerals (2 - 4Ø)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Granitic</td>
<td>5</td>
<td>36.2</td>
<td>42.4</td>
<td>50.9</td>
<td>6.7</td>
<td>3.50</td>
<td>-0.14</td>
<td>50.7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mean</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Standard Dev.</td>
<td></td>
<td>20.2</td>
<td>22.3</td>
<td>18.6</td>
<td>3.9</td>
<td>1.98</td>
<td>0.29</td>
<td>17.7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Metamorphic</td>
<td>11</td>
<td>16.5</td>
<td>49.7</td>
<td>5.6</td>
<td>2.8b</td>
<td>0.02</td>
<td>28.6</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mean</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Standard Dev.</td>
<td></td>
<td>16.8</td>
<td>17.1</td>
<td>3.3</td>
<td>0.48</td>
<td>0.13</td>
<td>10.9</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*aPercent of total sample
bPercent of sand/silt/clay fraction

Table 6. Comparative clay mineralogy data for tills collected in granitic and metamorphic bedrock areas of the Brady Glacier region, Alaska.

<table>
<thead>
<tr>
<th>Type</th>
<th>n</th>
<th>% Illite</th>
<th>% Vermiculite</th>
<th>% Chlorite</th>
<th>% Quartz</th>
<th>% Expandables</th>
</tr>
</thead>
<tbody>
<tr>
<td>Granitic Till</td>
<td>5</td>
<td>15</td>
<td>24</td>
<td>32</td>
<td>7</td>
<td>8</td>
</tr>
<tr>
<td>Mean</td>
<td></td>
<td>13</td>
<td>6</td>
<td>6</td>
<td>5</td>
<td>7</td>
</tr>
<tr>
<td>Standard Dev.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Metamorphic Till</td>
<td>9</td>
<td>26</td>
<td>21</td>
<td>23</td>
<td>8</td>
<td>16</td>
</tr>
<tr>
<td>Mean</td>
<td></td>
<td>23</td>
<td>16</td>
<td>16</td>
<td>7</td>
<td>16</td>
</tr>
<tr>
<td>Standard Dev.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*aPercent of clay fraction ± 5%
Conclusion

The Brady Glacier is fairly typical of large valley glaciers. Unlike much larger ice caps and ice sheets, the till deposited by this glacier is not a homogeneous unit that can be readily identified and traced over a wide geographic area on the basis of its mechanical or mineralogic properties. On the contrary, Brady Glacier till exhibits a considerable range of data for grain-size composition, roundness of quartz grains, percentage and type of heavy minerals present, sorting, skewness, and clay mineralogy, even for unweathered tills of different ages. Heavy mineral data show a correlation to local bedrock. This precludes correlation of Brady till deposits with "type" deposits of known age, even within the Brady Glacier drainage basin. This conclusion may not hold for Wisconsin ice sheet deposits which probably contain material contributed from source areas outside the present Brady Glacier drainage area. However, data from the single Wisconsin-age till deposit known in the region is not greatly different from younger deposits.
Cirque basins are a dominant feature of the glacially-sculptured mountains throughout the entire Glacier Bay region. They range in width from only a few tens of meters to several kilometers, with cirque floor elevations from below sea level to the upper parts of the highest peaks. At higher elevations, most cirques are still being actively formed by small cirque glaciers, but lower mountainsides are greatly scalloped by abandoned cirques formed under different climatic conditions during and since Pleistocene glaciation. Since cirques are an important feature of the landscape in the Brady Glacier region, an analysis of cirque orientation, relief, and vertical distribution was made for 154 prominent, abandoned cirques on both sides of the glacier (Mt. Fairweather Topographic Sheets B: 2-4). These cirques have steep walls and flat floors enabling measurement of elevations from 1:63,360 scale maps to an accuracy of about ± 15 m. They occur in about equal proportions in a coastal zone of moderately to strongly metamorphosed rocks, and in a more interior zone of mostly granitic igneous rocks (Figure 3). The modern glaciation level, or maximum height of unglaciated mountains in this area, rises to the northeast from Cape Spencer from 900 to 1000 m elevation (Østrem, 1972).
Abandoned Cirque Orientation

Cirque orientation, or aspect, is defined as the long axis of the cirque taken in the direction in which the cirque "opens" (Embleton and King, 1975). These authors discuss factors which can influence cirque orientation: (1) lithology, (2) jointing, (3) pre-existing valleys, (4) faulting, (5) bedding or foliation, (6) maritime influence, (7) prevailing wind direction, and (8) solar insolation. Lithology and geologic structure are usually of minor importance (Derbyshire, 1968; King, 1974; Embleton and King, 1975). Patterns of cirque orientation, major valley systems, jointing, and bedding/foliation strike are illustrated separately for each bedrock area by rose diagrams in Figures 42 and 43. Since few faults have been mapped in the Brady Glacier region, a composite of fault orientation in the entire Glacier Bay area is included in both figures (for data sources, see figure captions). A correlation analysis of various sets of these data is summarized in Tables 7 through 9. It is important to note that this analysis only serves to quantitatively compare different patterns. A high degree of correlation does not necessarily imply a causal relationship and vice-versa.

Table 7 compares orientation data between granitic and metamorphic bedrock areas. To eliminate the effect of pre-existing valleys on cirque orientation in each area, non valley-head cirque patterns are also compared. As the low correlation coefficients indicate, the patterns of cirque development, jointing, and bedding/foliation are quite different in each bedrock area. Valley orientation, however, is similar in both areas. The correlation coefficient of 0.417 is significant.
Figure 42. Orientation data for the metamorphic-bedrock area of the Brady Glacier region for: (a) all cirques in this area, (b) only non valley-head cirques, (c) valley strike, (d) joint strike, (e) fault strike in the Glacier Bay region, and (f) foliation - bedding strike in these rocks. Sources: (a) - (c) U.S.G.S. Mt. Fairweather Quads, B:2-4, C:2-4; (d) air photo lineations; (e), (f) maps by Seitz (1959), Rossman (1963 a&b), Plafker (1967), and Mackevett et al. (1971).
Total Cirques
82
Non Valley-head
Cirques
52
Valley
(d) Joints
66
Faults, Glacier Bay Region
191
Foliation Strike
104
0—10%

Figure 43. Orientation data for the granitic bedrock area of the Brady Glacier region for: (a) all cirques in this area, (b) only non valley-head cirques, (c) valley strike, (d) joint strike, (e) fault strike in the Glacier Bay region, and (f) foliation strike. Sources: (a) - (c) U.S.G.S. Mt. Fairweather Quads. B:2-4, C:2-4; (d) air photo lineations; (e), (f) maps by Seitz (1959), Rossman (1963 a&b), Plafker (1967), and Mackevett et al. (1971).
Table 7. Correlation analysis of eighteen 20° bearing increments of the orientation data shown in Figures 42 and 43 between metamorphic and igneous rock zones of the Brady Glacier region.

<table>
<thead>
<tr>
<th>Correlation Coefficient (r)</th>
<th>Significance Level of r (Φ)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Het. rock cirques (72) to Ig. rock cirques (82)</td>
<td>0.087</td>
</tr>
<tr>
<td>2. Het. rock, non valley-head (N.V.H.) cirques (35) to Ig. N.V.H. cirques (52)</td>
<td>-0.107</td>
</tr>
<tr>
<td>3. Het. rock valleys (70) to Ig. rock valleys (42)</td>
<td>0.417</td>
</tr>
<tr>
<td>4. Het rock joints (125) to Ig. rock joints (66)</td>
<td>0.202</td>
</tr>
<tr>
<td>5. Het. rock bedding/foliation (125) to Ig. rock foliation (104)</td>
<td>-0.360</td>
</tr>
</tbody>
</table>

* Number in parentheses indicates the number of measurements defining the population of each class.

Sources: (1) cirque and valley orientation: U.S.G.S. Mt. Fairweather Quads., B:2-4, C:2-4
(3) Fault and bedding and/or foliation strike: mapped by Seitz (1959), Rossman (1963a, 1963b), Plafker (1967), and Hackavett et al. (1971)
(4) significance level values of r: Fisher and Yates (1953) where d.f. = 16
at a 90% level, that is, there is only one chance in ten that the similarity between valley strike patterns in the two areas is purely accidental. Therefore, valley strike is probably controlled by some factor or factors which are not dependent on the local lithology. This tends to support the conclusion of Twenhofel and Sainsbury (1958) that major valleys in southeastern Alaska are fault-controlled.

Table 8 compares cirque orientation and valley strike data shown in Figure 42 with patterns of jointing and bedding/foliation strike in the metamorphic bedrock zone, and with the regional pattern of fault strike. In these rocks the pattern of cirques correlates with the pattern of valley strike, but the orientation of non valley-head cirques does not. Although valley strike correlates with joint strike in this area, again the pattern of non valley-head cirques does not. This suggests that pre-existing valleys played an important role in determining cirque orientation. Jointing, faulting, and foliation/bedding have undoubtedly all had some effect on cirque formation, but the dominant directions of cirque orientation and the formation of cirques with the greatest relief.

Table 9 lists the results of a similar correlation between cirque orientation, joint patterns, and foliation in the granitic bedrock area, and regional fault strike, all illustrated in Figure 43. In these rocks no significant correlations between cirque orientation and patterns of lithologic parameters exist. However, again there is a significant correlation between dominant directions of cirque orientation and cirques with the greatest relief.
Table 8. Correlation analysis of eighteen 20° bearing increments of the orientation data shown in Figure 42 for the metamorphic rock zone of the Brady Glacier region.

<table>
<thead>
<tr>
<th>Correlation to</th>
<th>Correlation Coefficient Level of r</th>
</tr>
</thead>
<tbody>
<tr>
<td>(Class &quot;A&quot;)</td>
<td>(Class &quot;B&quot;)</td>
</tr>
<tr>
<td>1. All cirques (72) to valley strike (70)</td>
<td>0.537</td>
</tr>
<tr>
<td>2. Non valley-head (N.V.H.) cirques (35) to valley strike (70)</td>
<td>0.165</td>
</tr>
<tr>
<td>3. N.V.H. cirques (35) to joint strike (125)</td>
<td>-0.306</td>
</tr>
<tr>
<td>4. N.V.H. cirques (35) to 90° from joint strike (125)</td>
<td>0.284</td>
</tr>
<tr>
<td>5. Valley strike (70) to joint strike (125)</td>
<td>0.498</td>
</tr>
<tr>
<td>6. N.V.H. cirques (35) to regional fault strike (191)</td>
<td>-0.113</td>
</tr>
<tr>
<td>7. N.V.H. cirques (35) to 90° from regional fault strike (191)</td>
<td>0.051</td>
</tr>
<tr>
<td>8. Valley strike (70) to regional fault strike (191)</td>
<td>0.114</td>
</tr>
<tr>
<td>9. N.V.H. cirques (35) to bedding/foliation strike (125)</td>
<td>-0.234</td>
</tr>
<tr>
<td>10. N.V.H. cirques (35) to 90° from bedding/foliation strike (125)</td>
<td>0.238</td>
</tr>
<tr>
<td>11. Valley strike (70) to bedding/foliation strike (125)</td>
<td>-0.019</td>
</tr>
<tr>
<td>12. All cirques (72) to mean relief of the 3 largest cirques in each 20° bearing increment (n=14)</td>
<td>0.507</td>
</tr>
</tbody>
</table>

* Number in parentheses indicates the number of measurements defining the population of each class.
Sources:(c.f. Table 7)

Table 9. Correlation analysis of eighteen 20° bearing increments of the orientation data shown in Figure 43 for the igneous rock zone of the Brady Glacier region.

<table>
<thead>
<tr>
<th>Correlation to</th>
<th>Correlation Coefficient Level of r</th>
</tr>
</thead>
<tbody>
<tr>
<td>(Class &quot;A&quot;)</td>
<td>(Class &quot;B&quot;)</td>
</tr>
<tr>
<td>1. All cirques (82) to valley strike (42)</td>
<td>0.265</td>
</tr>
<tr>
<td>2. Non valley-head (N.V.H.) cirques (52) to valley strike (42)</td>
<td>0.151</td>
</tr>
<tr>
<td>3. N.V.H. cirques (52) to joint strike (66)</td>
<td>0.013</td>
</tr>
<tr>
<td>4. N.V.H. cirques (52) to 90° from joint strike (66)</td>
<td>-0.278</td>
</tr>
<tr>
<td>5. Valley strike (42) to joint strike (66)</td>
<td>-0.232</td>
</tr>
<tr>
<td>6. N.V.H. cirques (52) to regional fault strike (191)</td>
<td>0.165</td>
</tr>
<tr>
<td>7. N.V.H. cirques (52) to 90° from regional fault strike (191)</td>
<td>-0.026</td>
</tr>
<tr>
<td>8. Valley strike (42) to regional fault strike (191)</td>
<td>-0.186</td>
</tr>
<tr>
<td>9. N.V.H. cirques (52) to foliation strike (104)</td>
<td>0.307</td>
</tr>
<tr>
<td>10. N.V.H. cirques (52) to 90° from foliation strike (104)</td>
<td>-0.277</td>
</tr>
<tr>
<td>11. Valley strike (42) to foliation strike (104)</td>
<td>0.118</td>
</tr>
<tr>
<td>12. All cirques (82) to mean relief of the 3 largest cirques in each 20° bearing increment (n=13)</td>
<td>0.808</td>
</tr>
</tbody>
</table>

* Number in parentheses indicates the number of measurements defining the population of each class.
Sources:(c.f. Table 7)
In conclusion, cirque development in this region is undoubtedly affected to some degree by joints, faults, foliation, and bedding, but the dominant influence of any one of these lithologic parameters is not apparent from this analysis. The effect of pre-existing valleys may have been important in determining cirque aspect in the metamorphic rock area only. However, if only geologic structure determined the measured cirque aspect in each area, the patterns of cirque development should be nearly symmetrical about some axis. Figures 42 and 43 indicate that this is not the case. The vector sum of all cirque orientations in the Brady Glacier region has a value of 17.0°, 63.9° ± 10°. The vector sum of the non valley-head cirques has a similar value of 18.7°, 54.7° ± 10°. The strength of this northeastern trend must be due to factors other than geologic structure.

Maritime influence assures plentiful precipitation throughout southeastern Alaska, but the warming influence of the North Pacific Drift is diffused over the entire coastal region, and cannot explain local preference in cirque aspect. Wind direction, on the other hand, can be a major factor (Goldthwait, 1970; Evans, 1974).

In general, winds are from the southeasterly quadrant from the last of October to the middle of May, and are often accompanied by rain or drizzle at sea level. However, heaviest precipitation comes with the gales of Autumn and Winter which gradually set in with a southeasterly wind which shifts to southwesterly and increases in velocity as the storm passes (U.S. Coast and Geodetic Survey, 1969). A southerly displacement of these storm tracks during former glacial periods is probable (Flint, 1971; M.M. Miller, 1975), but would only have the effect of
shifting storm winds to a more southeasterly direction. Since snow accumulation tends to be greatest for those cirques on the lee side of ridges, sheltered from storm winds (Goldthwait, 1970), it follows that cirques with a northerly component of orientation should have been most favored for growth.

The importance of shading from the afternoon sun on cirque orientation is also well known (Derbyshire, 1968; Flint, 1971; King, 1974; Embleton and King, 1975). On June 21 in this region, the sun appears to move from due south at solar noon to set at a northwestern bearing of 314.5° nine hours later. These long days least affect cirques facing the northeastern quadrant, which is probably the major reason for the predominance of this orientation in the Brady Glacier region.

Local conditions, such as the shading of some cirques due to the configuration and great relief of topography or the funneling of winds down valleys, may have great effect on any particular cirque. Geologic structure is important on a larger scale. However, for the entire region the combined effects of wind and insolation were probably most important in causing the dominant northeastern orientation of abandoned cirques.

Cirque Relief

Mean cirque relief is nearly the same in both granitic and metamorphic rocks (Figure 44). As might be expected, the largest cirques tend to face in directions where cirque orientation is also strong.
Figure 44. Mean cirque relief in the Brady Glacier region, in each of the two major lithologies, and for the 20 largest cirques (by quadrant).

Figure 45. Mean cirque floor elevation in the Brady Glacier region, and in each of the two major lithologies (by quadrant).
(Tables 8 and 9, no. 12). Overall, mean relief of cirques in this region tends to be lowest for cirques facing the southeastern quadrant and highest for cirques facing the northwest, but this pattern varies with lithology (Figure 44). Mean relief of only the 20 largest cirques in each quadrant is greatest for those cirques facing northeast, but again it is least for those facing the southeastern quadrant (Figure 44). The development of the greatest relief in northerly-oriented cirques again suggests the importance of storm winds (which determine snow accumulation) and insolation (which determines snowmelt) on cirque development.

Vertical Distribution of Cirque Floors

The data in Figure 45 indicate that the mean cirque floor elevation in metamorphic rocks is 18.4% lower than that in granitic rocks. It is uncertain whether this: (1) reflects the rain shadow effect of the metamorphic hills, (2) is due to less burial and erosion of the coastal metamorphic rocks by Pleistocene icesheets, or (3) is a purely statistical effect due to a somewhat higher relief of the granitic rock terrain. From a regional viewpoint, the mean height of cirque floors is greatest for southwest-oriented cirques and lowest for northwest-oriented cirques.

Regionally, concentrations of cirque floor elevations occur at about 165, 335, and 410 to 505 m (Figure 46). This seems comparable to values further inland of 110, 340, 530, 750, and 960 m elevation for
Figure 46. A histogram of the occurrence of the floor elevation of 154 abandoned cirques in the Brady Glacier region. Modern glaciation limit lies between 900 and 1000 meters elevation (Ostrem, 1972). The peaks in this histogram may reflect stillstand positions of firm line during and since Late Wisconsin time.
concentrations of cirque floors in the Alaska-Canada Boundary Range which M.M. Miller (1961, 1975) suggested were related to periods of stillstand of firm line during and since Late Wisconsin time.

Cirque Morphometry, Climate, and Glaciation

The coastal zone of metamorphic rocks was probably least overriden by Pleistocene icesheets due to its proximity to the sea. Thus, discrete cirque glaciers must have existed over a longer time interval here than in the interior. The morphology of these cirque basins was probably less altered by eroding ice sheets subject to a different flow pattern than small cirque glaciers. Bearing this out, the low mean height of cirque floors and great relief of the northerly-oriented cirques in these rocks suggests that these cirques formed under near full-glacial conditions of the Pleistocene. During the height of glacial periods, the granitic rocks must have been nearly completely buried by ice. The strong northeasterly trend in cirque orientation in both bedrock areas, but particularly in the granitic rocks, must have developed in more recent time in response to a climatic regime more like that of the present. In conclusion, the complex pattern of cirque orientation, relief, and elevation in this region must be understood in terms of the effects of changing climate and the sheer size and extent of former icesheets superimposed on a landscape characterized by variable conditions of topography, lithology, and structure.
### APPENDIX A

**DESCRIPTION OF RADIOCARBON SAMPLE LOCATIONS**

<table>
<thead>
<tr>
<th>Date (B.P.)</th>
<th>Lab. No.</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>4680 ± 160</td>
<td>Y-9</td>
<td>Fragments, mostly from small deciduous trees, in very compact till 50 to 60 feet below till surface on east side of terminus of Reid Glacier (Preston, 1955).</td>
</tr>
<tr>
<td>2870 ± 80</td>
<td>DIC-460</td>
<td>(58°, 20' N; 136°, 30' W) Driftwood at the base of a 2 m section of interstratified gray clay, sand, and gravelly sand overlain by till. This exposure occurs in a small stream cut in the outermost of several concentric end moraines deposited by the Brady Glacier near southwestern Dundas Bay.</td>
</tr>
<tr>
<td>1960 ± 90</td>
<td>DIC-458</td>
<td>(58°, 25' N; 136°, 50' W) Horizontal log with the bark intact lying in forest duff overlying outwash. This forest was covered by 6 cm of fine brown sand, then about 1.5 m of varved gray clay, which was capped by several large boulders (residual till). The clays have been deformed into asymmetric folds with amplitudes up to 30 cm. This section is exposed by the Dixon River in the northeastern part of the Boussole Valley.</td>
</tr>
<tr>
<td>1960 ± 65</td>
<td>DIC-461</td>
<td>(58°, 22' N; 136°, 17' W) Stumps exposed in the intertidal zone of eastern Dundas Bay, 2.4 m above mean low tide. Raised marine deposits occur landward of this locality to an elevation of about 9 m where a sharp, sheer cliff rises to an elevation of about 20 m. A young (80 year-old) forest grows at the base of this cliff.</td>
</tr>
<tr>
<td>1780 ± 65</td>
<td>DIC-555</td>
<td>(58°, 20' N; 136°, 30' W) Driftwood fragment from glaciomarine outwash containing shells of <em>Macoma balthica</em> interstratified with intertidal sands and gray clay. Exposed 50 cm below ground surface in a small creek draining the east side of the lowland between Taylor and Dundas Bays.</td>
</tr>
<tr>
<td>Date (B.P.)</td>
<td>Lab. No.</td>
<td>Description</td>
</tr>
<tr>
<td>------------</td>
<td>----------</td>
<td>-------------</td>
</tr>
<tr>
<td>1750 ± 60</td>
<td>DIC-462</td>
<td>(58°, 24' N; 136°, 37' W) Fragment of a 59 year-old tree imbedded in an exposure of 1.5 m of compact gray till along the shoreline of an unnamed ice-dammed lake near the northwest arm of Dundas Bay.</td>
</tr>
<tr>
<td>1740 ± 100</td>
<td>DIC-285</td>
<td>(58°, 25' N; 136°, 54' W) Basal 1 cm of peat overlying a small glaciofluvial delta between two recessional moraines in the Boussole Valley.</td>
</tr>
<tr>
<td>1585 ± 85</td>
<td>DIC-286</td>
<td>(58°, 25' N; 136°, 58' W) Stumps rooted in coarse outwash gravel at 7 m elevation in the lower Palma Valley. These stumps, and three higher levels of stumps, were buried by fine alluvial sands during a period of outwash aggradation. Section exposed by the modern Palma River.</td>
</tr>
<tr>
<td>1435 ± 45</td>
<td>DIC-284</td>
<td>(58°, 23' N; 136°, 44' W) Wood fragments in a compact gray till exposed in a small stream cut on the eastern side of the Dixon Valley at an elevation of about 25 m. Large, glacially-transported boulders extend upslope another 17 m.</td>
</tr>
<tr>
<td>1230 ± 60</td>
<td>DIC-459</td>
<td>(58°, 25' N; 136°, 49' W) Log fragment in 2.5 m exposure of compact, gray till overlain by 0.6 m of varved clay. Locality is about 12 m above the Dixon River in a small stream cut on the north end of the ridge separating Dixon and Boussole Valleys.</td>
</tr>
<tr>
<td>1030 ± 60</td>
<td>DIC-554</td>
<td>(58°, 17' N; 136°, 35' W) Horizontal driftwood log in a raised marine beach deposit at an elevation of about 4 m on the southwestern shore of Taylor Bay.</td>
</tr>
<tr>
<td>685 ± 40</td>
<td>UN-14</td>
<td>(58°, 33' N; 136°, 43' W) Wood from the outside of a 2-foot diameter stump having 300 growth rings. The stump, transported by the Brady Glacier from an unknown but probably nearby location at a somewhat lower altitude, was partially imbedded in an upended position ca. 20 feet below the top of a lateral moraine at an elevation of about 1750 feet (Dorn, et al., 1962).</td>
</tr>
<tr>
<td>Date (B.P.)</td>
<td>Lab. No.</td>
<td>Description</td>
</tr>
<tr>
<td>------------</td>
<td>---------</td>
<td>-------------</td>
</tr>
<tr>
<td>433 ± 80</td>
<td>UJ-21</td>
<td>(59°, 26' N; 136°, 52' W) Wood from one of a group of upright stumps whose bases are buried in sand ca. 6 feet below grade in the Dixon River Valley (Dorn, et al., 1962).</td>
</tr>
<tr>
<td>340 ± 100</td>
<td>DIC-287</td>
<td>(58°, 23' N; 136°, 50' W) Driftwood in a gray clay matrix containing abundant marine shells near the contact with overlying brown alluvial sand. Exposed in small creek draining the northwest wall of the lower Dixon River Valley at an elevation of 3.6 m.</td>
</tr>
<tr>
<td>300 ± 105</td>
<td>DIC-556</td>
<td>(58°, 23' N; 136°, 46' W) Wood from the outside of a standing dead stump well out on the marshland adjacent to North Deception Lake. This tree was killed and partially buried by lake sediments when advancing Brady Glacier ice formed a much-enlarged North Deception Lake.</td>
</tr>
<tr>
<td>310 ± 70</td>
<td></td>
<td></td>
</tr>
<tr>
<td>260 ± 50</td>
<td>UW-15</td>
<td>(58°, 34' N; 136°, 30' W) Wood from the outside of a log about 2.8 feet in diameter, transported by a distributary tongue of the Brady Glacier from an unknown, but probably nearby point of growth at a somewhat lower elevation by advancing ice. The specimen was found on the surface of the ice near the end of the distributary tongue (Dorn, et al., 1962).</td>
</tr>
</tbody>
</table>

APPENDIX A (Continued)
# APPENDIX B

Till Grain Size Distribution, Sorting, Skewness, and Percent Heavy Minerals

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Age Type</th>
<th>Type</th>
<th>&gt;-150</th>
<th>Sand</th>
<th>Silt</th>
<th>Clay</th>
<th>Log_{10} Trask</th>
<th>Trask</th>
<th>Skewness</th>
<th>Minerals</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Wis.</td>
<td>Basal</td>
<td>46.8</td>
<td>16.3</td>
<td>76.1</td>
<td>7.5</td>
<td>2.29</td>
<td>-0.03</td>
<td>11.1</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>E. Neo.</td>
<td>Basal</td>
<td>62.1</td>
<td>45.0</td>
<td>48.4</td>
<td>6.6</td>
<td>3.74</td>
<td>0.13</td>
<td>31.7</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>&quot;</td>
<td>&quot;</td>
<td>20.0</td>
<td>50.6</td>
<td>45.1</td>
<td>4.4</td>
<td>2.29</td>
<td>0.13</td>
<td>31.7</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>&quot;</td>
<td>&quot;</td>
<td>56.8</td>
<td>65.5</td>
<td>32.5</td>
<td>2.0</td>
<td>3.60</td>
<td>-0.03</td>
<td>59.7</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>&quot;</td>
<td>&quot;</td>
<td>11.2</td>
<td>9.0</td>
<td>77.4</td>
<td>13.6</td>
<td>2.21</td>
<td>0.10</td>
<td>29.8</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>&quot;</td>
<td>&quot;</td>
<td>20.5</td>
<td>5.7</td>
<td>83.0</td>
<td>11.2</td>
<td>1.94</td>
<td>-0.03</td>
<td>61.0</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>E. Neo.</td>
<td>Ablation</td>
<td>56.5</td>
<td>61.9</td>
<td>34.3</td>
<td>3.8</td>
<td>6.84</td>
<td>-0.64</td>
<td>69.2</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>&quot;</td>
<td>&quot;</td>
<td>29.0</td>
<td>70.7</td>
<td>27.1</td>
<td>2.1</td>
<td>3.38</td>
<td>-0.17</td>
<td>26.0</td>
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<tr>
<td>9</td>
<td>&quot;</td>
<td>&quot;</td>
<td>37.4</td>
<td>65.7</td>
<td>30.4</td>
<td>3.9</td>
<td>2.92</td>
<td>-0.06</td>
<td>30.1</td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>&quot;</td>
<td>&quot;</td>
<td>35.8</td>
<td>69.6</td>
<td>27.3</td>
<td>3.1</td>
<td>2.91</td>
<td>-0.12</td>
<td>33.8</td>
<td></td>
</tr>
<tr>
<td>11</td>
<td>&quot;</td>
<td>&quot;</td>
<td>64.3</td>
<td>62.6</td>
<td>34.2</td>
<td>3.3</td>
<td>3.21</td>
<td>-0.06</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>&quot;</td>
<td>&quot;</td>
<td>36.2</td>
<td>64.9</td>
<td>31.6</td>
<td>3.4</td>
<td>2.60</td>
<td>-0.05</td>
<td>41.2</td>
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### APPENDIX C

#### Till Clay Mineralogy Data

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REFERENCES (Continued)


REFERENCES (Continued)


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PLATE 1.

Surficial Geology of the Brady Glacier Region, Alaska

OUTWASH

MORAINES:

Late Neoglacial

Early Neoglacial

WISCONSIN

ICE LIMITS:

WISCONSIN

ROCK GLACIER
Surficial Geology of the Brady Glacier Region, Alaska

- OUTWASH
- MORAINES:
  - Late Neoglacial
  - Early Neoglacial
- Wisconsin

ICE LIMITS:
- Wisconsin
- Late Neoglacial

ICE FLOW DIRECTION INDICATOR

ABANDONED MELTWATER CHANNEL

FORMER SEA CLIFF

ROCK GLACIER

TALUS CONE

ABANDONED CIRQUE