# Chronology of Late Holocene Glacier Recessions In the Cascade Range and Deposition of a Recent Esker in a Cirque Basin, North Cascade Range, Washington

by

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ABSTRACT

Recessional events over the past 800 years between 23 glaciers of varying size, elevation, and aspect in the North Cascades and on Mount Rainier were found to be broadly synchronous. Eight periods of recession were identified (early 1200s, the mid-1300s, the mid to late-1400s, between 1520 and 1580, the mid to late-1600s, between 1740 and 1780, between 1820 and 1860, and between 1880 and 1920), although the three earliest periods are poorly documented. Establishment of a more detailed chronology is dependent upon more accurate and consistent dating methods and the acquisition of more data.

Variable timing of late Neoglacial maximum downvalley positions of the glaciers studied indicates that glacier response to climatic signals is somewhat individual and complex. The spatial and temporal scales of glacier fluctuations are impossible to determine solely on the basis of moraine data. Observed twentieth-century rates of retreat and climatic data indicate, however, that some fluctuations covered distances of several hundred meters.

The unusual deposition of a 120 m long fine-grained esker in a rugged alpine environment is related to the presence of a large modern glacier that flowed down into an ancient Pleistocene cirque, the lobate flow of ice into the cirque, and the stagnation of ice in the cirque following rapid twentieth-century retreat of the Redoubt Glacier.

Longitudinal persistence of sedimentary units indicate that fluvioglacial deposition on the bed of a stream was the primary process responsible for the deposition and form of the esker. Of less importance was deposition of sediment directly from glacier ice and locally in water ponded in the tunnel. Deposition of the esker was rapid and episodic, and was separated by periods of erosion or non-deposition as indicated by the presence of two fining-upward sequences and a disconformity between sedimentary units. Deposition and tunnel widening, which caused the faulting of some units, was controlled by

either seasonal or shorter-term precipitation related fluctuations in discharge.

In addition to fluvioglacial deposition on the bed of the stream, the form of the esker was controlled by the original shape of the tunnel. Tunnel shape may have been controlled by structural features of the ice mass in an interlobate zone. This resulted in an esker that narrows from 36 m to 7 m downstream and has four perpendicular side ridges.

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# CHAPTER 1 : INTRODUCTION

Alpine glaciers cover over 270 $km^2$ of the North Cascade Mountains in Washington (Post et al., 1971). Major concentrations of glaciers occur in the Glacier Peak, Mount Baker, Mount Shuksan, Dome Peak, Eldorado Peak, and Mount Redoubt areas and on volcanic peaks in the central and southern Cascades, including Mount Rainier (Figure 1). Glaciers in the Cascades are important as indicators of climatic change, reservoirs of fresh water, and as influences on vegetation patterns and the evolution of landforms. This study focuses on late Holocene fluctuations of valley glaciers and glacial landform evolution.

No recent study has developed a regional chronology of late Holocene glacier fluctuations in the Pacific Northwest. The first part of this thesis establishes a moraine chronology of the beginning of recessional episodes over the last 800 years. The chronology is based on dating the timing of moraine stabilization on a series of moraines deposited by the Redoubt Glacier (Figure 2) and by comparing this chronology to similar data from 22 other glaciers. In order to determine the number, timing, and synchroneity of recessional events between the 23 glaciers, the influence of glacier size and aspect, and different dating methods were also examined. Finally, the spatial and temporal scales of the recessional events are also discussed.

Recent geomorphological research has emphasized the importance of process-form relationships as keys to the understanding of glacial landscape evolution (Price, 1973, p.17; Goldthwait, 1975, p.5; Sugden and John, 1976, p.3; Gjeissing, 1975). Such relationships may best be identified in modern glacial environments where active processes may be observed. The second part of this thesis focuses on the problem of esker genesis in a cirque basin near Mount Redoubt (Figure 2). Despite over 100 years of investigation, many questions remain concerning the formation and distribution of eskers. In this study, I attempt to determine why this esker is located in a somewhat unusual setting and what processes controlled its deposition and form.

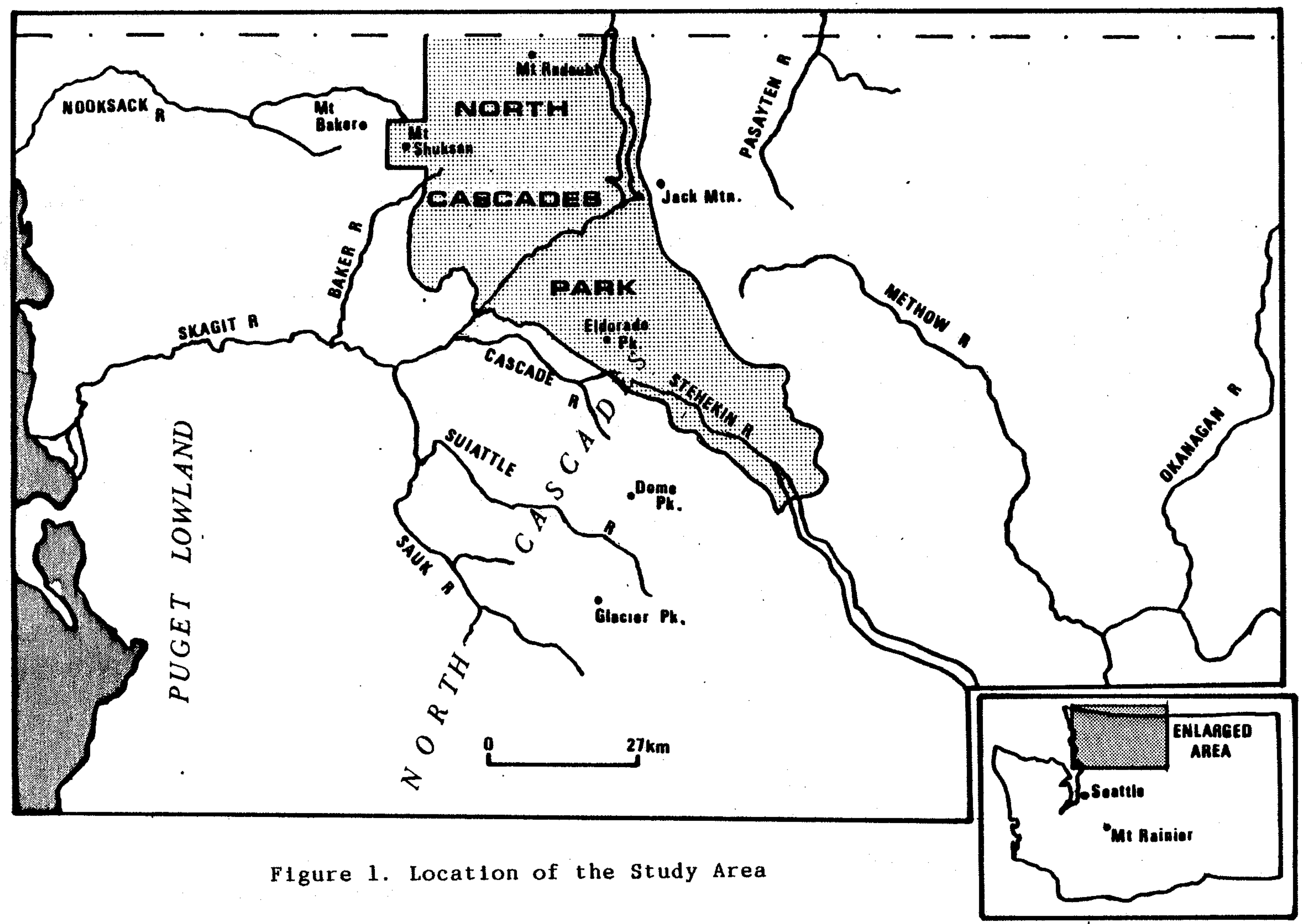



Figure 1. Location of the Study Area

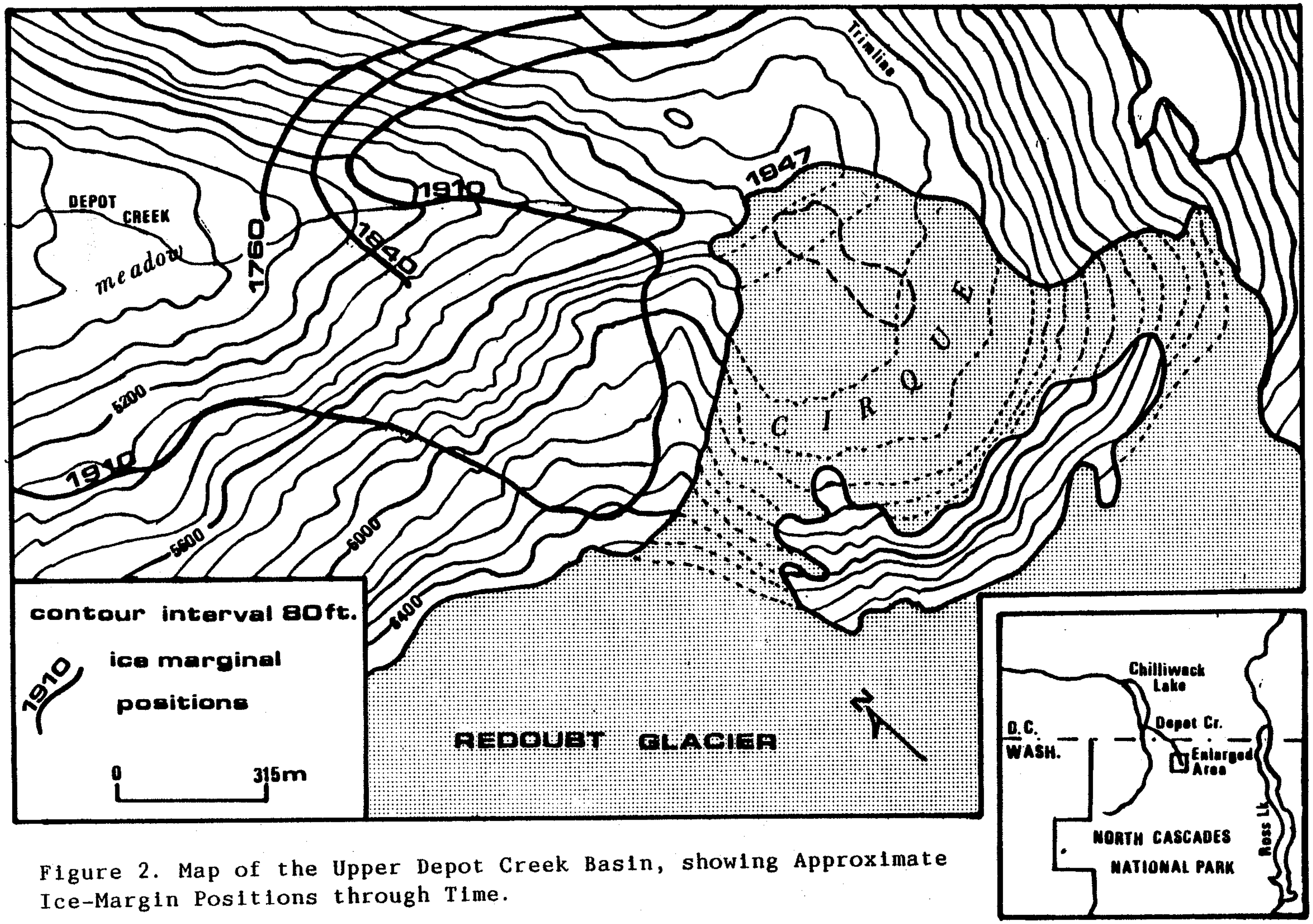



Figure 2. Map of the Upper Depot Creek Basin, showing Approximate Ice-Margin Positions through Time.

## CHAPTER 2 : DESCRIPTION OF THE STUDY AREA

### A. Location

The recently deglaciated terrain of the upper Depot Creek basin is located in the North Cascade Mountains of northwestern Washington (Figure 1). It lies within the borders of North Cascades National Park, at the headwaters of Depot Creek, approximately four kilometers south of the United States-Canada border (Figure 2). The study area extends from an elevation of approximately 1490 m in a small meadow to 1705 m in a cirque basin (Figure 2).

Physiographically, the North Cascades province is bordered on the north by the Fraser River in British Columbia, on the east by the Okanogan and Columbia Rivers, on the south near Snoqualmie Pass and on the west by the Puget Lowland (Easterbrook and Rahm, 1970). The North Cascades stand out as a distinct physiographic province because they were lifted higher and are more deeply dissected than the Cascade mountains farther to the south (Easterbrook and Rahm, 1970; Misch, 1977). Local relief in the North Cascades is as much as 2000 m, with peaks averaging 2500 m in height.

### B. Bedrock Geology

Metamorphism, folding, faulting, the intrusion of igneous rock bodies, and volcanic activity have been recurrent themes in the geologic history of the North Cascades (McKee, 1972, p.85-87). The first major recorded orogeny recorded occurred during the Cretaceous period, when the present structural units of the North Cascades were formed (Misch, 1977). The crystalline core of the range is the northwest trending Skagit Metamorphic Suite, which is composed of Cascade River Schist and Skagit Gneiss (also known as Custer Gneiss described by McTaggart and Thompson, 1976) (Misch, 1966). West of the Skagit Metamorphic Suite, across a major north to south trending fault, is the Shuksan Metamorphic Suite. The suite is composed of

Darrington Phyllite and Shuksan Greenschist, which were thrust 50 to 65 kilometers westward over Paleozoic strata along relatively flat glide planes (McKee, 1972, p.89; Misch, 1966). Rocks of the Yellow Aster Complex were squeezed up at the sole of the thrust sheet. These rocks date to the Proterozoic era and are the oldest rocks exposed in the North Cascades (Misch, 1977). The eastern border of the Skagit Metamorphic Suite is the Ross Lake fault (Staatz, et al., 1972, p.46). East of the fault lie steeply folded lower Cretaceous strata, which were covered by dominantly basaltic upper Paleozoic rocks thrust eastward along the Jack Mountain Thrust (Misch, 1966). Adjacent to the fault zone Cretaceous metamorphism has affected both younger (Jack Mountain Phyllite) and older (Elijah Ridge Schist) rocks (Misch, 1966).

Intrusion of the Chilliwack Composite Batholith during the mid-Tertiary Period and the deposition of extrusive igneous rocks of the Skagit Volcanic and Hannegan Volcanic Formations were the next major events that shaped the bedrock geology of the North Cascades. Rocks of the Chilliwack Composite Batholith entered the North Cascades in four intrusive phases (Misch, 1966). The four plutons vary in age, composition and structure, with the largest bodies composed of quartz diorite and granodiorite, and smaller ones being composed of quartz monzonite, diorite, alaskite, and gabbro (Staatz et al., 1972, p.35-36). Both predating and coeval with the intrusion of the Chilliwack Composite Batholith was the placement of the Hannegan and Skagit Volcanic Formations. These formations are composed of tuffs, volcanic breccias, and flows. The Skagit Volcanic Formation parallels the United States-Canada border and is in places over 1100 m thick (Staatz, et al., 1972, p.22).

Late-Tertiary and Quaternary events shaped the modern topography and geology of the North Cascades. Uparching began in the Pliocene Epoch and followed a north-south axis, which makes its structural patterns discordant with the northwest pattern of Cretaceous events (Misch, 1977). Estimates of total uparching near the 49th parallel have been as high as 2090 m (McKee, 1972, p.186). The strong uplift has resulted in the deep incision of river valleys. Only the Skagit River, however, predates the latest uplift since it cuts across the crystalline core of the range (Misch, 1966). Subduction of the Pacific Ocean plate beneath the North American plate has resulted in the building of volcanic cones, such as Mount Baker, Mount Rainier, and Glacier Peak, upon the pre-existing topography.

Three rock types representing both Cretaceous and Tertiary events are found in the upper Depot Creek basin. The west wall of the cirque is composed of quartz diorite and granodiorite from the Chilliwack Composite Batholith. On the south wall of the cirque the quartz diorite and granodiorite are overlain by Custer Gneiss. The east wall of the cirque is composed primarily of Custer Gneiss, although it is topped above 2240 m by rocks of the Skagit Volcanic Formation. Rocks of the Chilliwack Composite Batholith presumably underlie the cirque floor and the area between the cirque and the meadow, although glacial drift covers most of the cirque floor, the upper Depot Creek Valley, and the meadow (Figure 2).

## C. Climate

Primary factors influencing the climate of the North Cascades are latitude, proximity to the Pacific Ocean, the height of the North Cascade range, and semi-permanent high and low pressure cells located over the North Pacific Ocean (NOAA, 1979a). The position of the study area is within the belt of prevailing westerlies, which bring wave cyclones and moisture laden air from the Pacific Ocean onto the North American continent. The orographic effect increases precipitation at high altitudes when air from the Pacific cools as it rises over the North Cascades. The presence of semi-permanent pressure regions over the North Pacific imparts a strong seasonal component to precipitation in the Pacific Northwest. During the summer, precipitation is less because high pressure dominates the North Pacific. Circulation around the high pressure system causes a northwesterly flow of relatively cooler, drier air onto the continent. In late fall and winter the

Aleutian low pressure system dominates the North Pacific, bringing a predominantly southwesterly flow of warmer, moister air over the North Cascades. Cooling and condensation of this air is enhanced by the relatively cold land mass in winter.

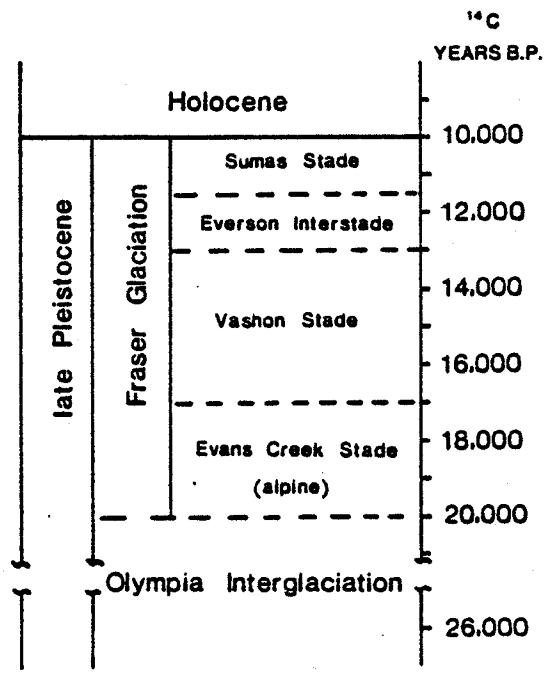
Northwestern Washington's climate is classified as marine with wet, mild winters and cool, dry summers (NOAA, 1979b). The closest weather station is located 25 km to the west at the Heather Meadows Recreation Area (elevation 1254 m). Mean annual precipitation at the Heather Meadows station was 2812 mm for the 1927-1951 period. The NOAA (1979b) estimates snowfall in the Western Cascades between 1194 m and 1493 m at between 10240 mm and 15360 mm annually. The mean annual temperature at the Heather Meadows station for the 1927-1951 period was 4.5°C. Finally, mean monthly temperature at elevations of 1642m in the western North Cascades is approximately -3.9°C in January and 7°C in July.

## D. Late Wisconsin Glacial History of the North Cascades

Quaternary glaciation has been the most important factor shaping the modern topography of the North Cascades (Post, et al., 1971). Erosional land forms dominate the landscape and attest to the repeated, intensive nature of glaciation of the range. In contrast to the Puget and Okanogan lowlands and portions of the southern and eastern North Cascades, relatively little is known about the glacial history of the range within and adjacent to the North Cascades National Park (Figure 1). This stems from the area's rugged, remote and pristine nature, dense vegetation cover in valley bottoms where most deposits are likely to be found, high rates of erosion, and the complicating influence the Cordilleran ice sheet had on geomorphology of the range north of 48° latitude.

Armstrong (et al., 1965) divided the late Wisconsin (also known as the Fraser Glaciation) into six geologic-climatic periods based on radiocarbon chronology, tephrochronology, palynology, and stratigraphic studies throughout the Pacific Northwest (Figure 3).

Figure 3. Late Pleistocene Glacial Sequence in the Central Puget Lowland. Dashed lines indicate approximate ages. From Heller (1980).



Climatic cooling following the Olympia Interglacial period led to the expansion of alpine glaciers in the North Cascades around 20,000 years B.P.. A limited amount of data has been gathered in the North Cascades concerning the fluctuations of early Fraser alpine glaciers. Presumably, most depositional evidence in the North Cascades was removed by the Cordilleran ice-sheet, which covered most of the region by 15,000 years B.P. (Figure 4). Therefore, reconstructions of early Fraser alpine glacier fluctuations are based on data gathered south of the Cordilleran ice-sheet limit, which assumes that fluctuations throughout the Cascade Range were synchronous.

Alpine glaciers on Mount Rainier reached their maximum extent around 18,000 years B.P. (Crandell, 1963). Less extensive advances south of the Cordilleran limit occurred between 16,000 and 13,000 years B.P., and around 12,000 and 11,000 years B.P. (Porter, Pierce and Hamilton, 1983). In the North Cascades, Clague (1980) believes that Chilliwack Lake is dammed by a moraine constructed by a valley glacier around 11,000 years B.P..

Although most moraines were destroyed by the advance of the Cordilleran ice sheet across the range around 15,000 years B.P. (Figure 4), erosional evidence of the downvalley extent of alpine glaciers has allowed the reconstruction of their areal extent and equilibrium line altitudes (ELAs) (Figure 5). The reconstructed ELAs of the Evans Creek alpine glaciers have been compared to those of modern glaciers to assess the relative degree of climatic change between the two periods (Waitt, 1977). Waitt (1977) estimated that ELAs were depressed 700m during maximum Evans Creek glaciation at 48° 30' north latitude.

Alpine glaciers headed at major divides in cirque basins. Nivation and rotational ice movement caused intense erosion of cirque floors, creating depressions that are now occupied by tarns such as Azure, Sulphide, Doubtful, Bear, Luna, Ouzel, and Upper Thornton lakes. Cirque headwall recession sculpted the resistant, higher rocks of the crystalline core of the range into arêtes and horns. In contrast to cirques being formed by modern glaciers, which are found

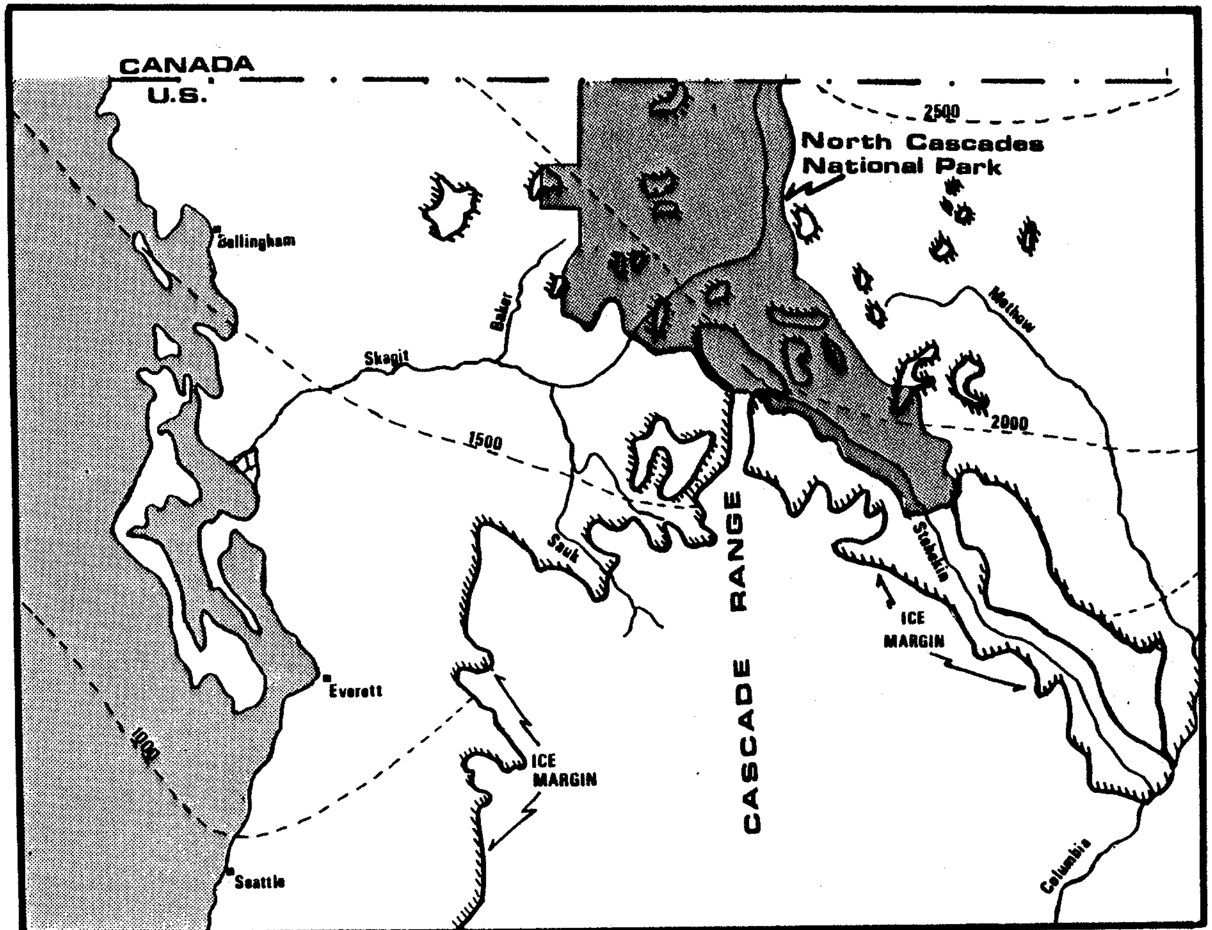



Figure 4. The Extent of the Cordilleran Ice Sheet in the Vicinity of North Cascades National Park around 15,000 years B.P.. Ice surface contours (dashed lines) are in meters. Modified from Waitt and Thorson (1983).

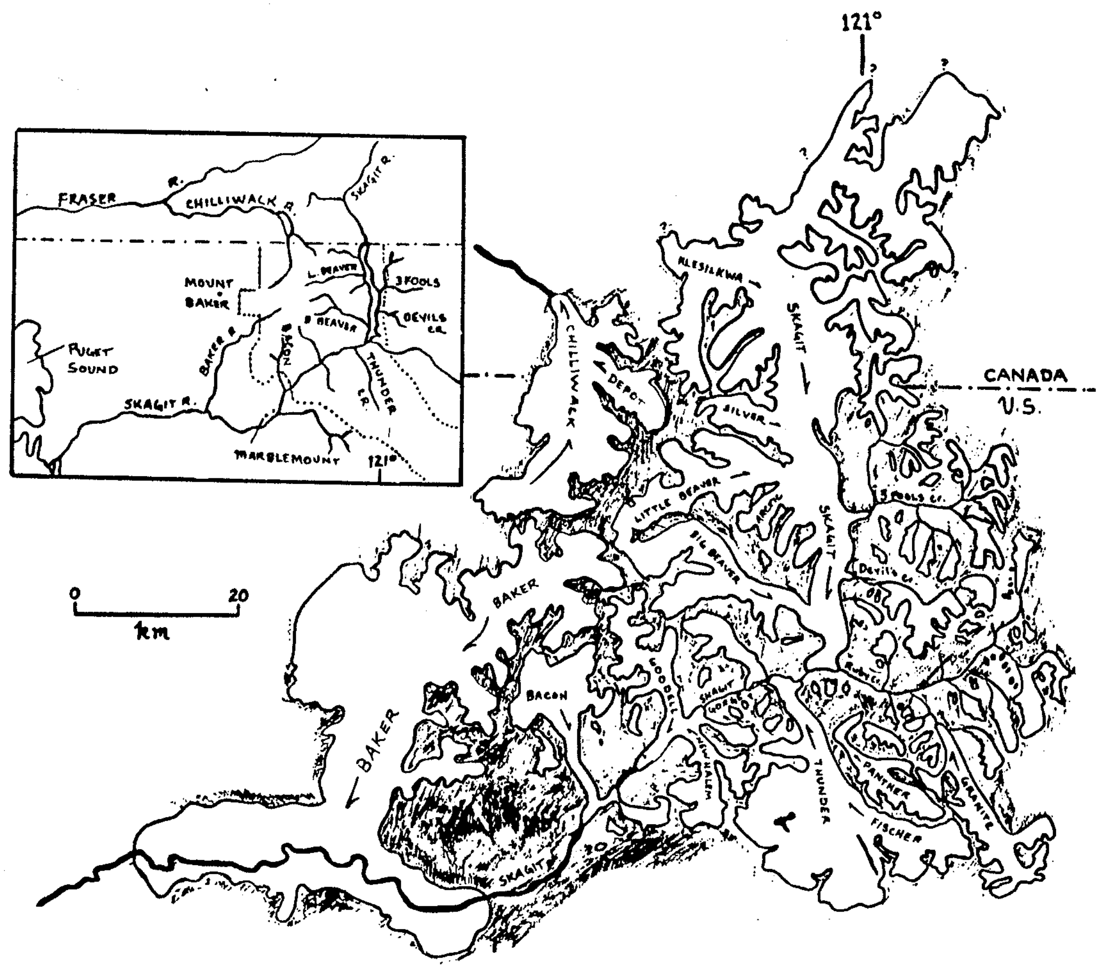



Figure 5. Areal Extent of Alpine Glaciers in the Vicinity of North Cascades National Park around 18,000 years B.P.. Modified from Waitt (1977) and Heller (1980).

primarily on protected north and northeast facing slopes, the ancient cirques are distributed symmetrically around most peaks. The symmetry of the ancient cirque aspects observed in the North Cascades suggests that a model for alpine glaciation proposed by Evans (1974) is applicable to the North Cascades. Evans (1974) suggested that as ELAs drop during increasingly intensive glaciation, glacial cover (and therefore cirque orientation) becomes more symmetrical as aspect becomes a less important factor.

In addition to the erosion of cirques, the alpine valley glaciers left many other imprints on the topography of the North Cascades. They straightened, broadened, and deepened river valleys, leaving them with U-shaped cross valley profiles. Smaller tributary valley glaciers did not carry enough ice to erode down to the level of the thicker trunk valley glaciers. Therefore, they were left as hanging valleys with cascading waterfalls at their mouths. Between their tributaries, the major valley glaciers truncated valley spurs and occasionally left lakes behind terminal moraines. Lakes such as Kachess, Cle Elum, Keechelus, and Wenatchee, however, are conspicuously absent north of the Cordilleran ice-sheet limit--with the exception of Chilliwack Lake (Figure 2) (Clague and Luternauer, 1983). Presumably the ice sheet destroyed such lakes within most of the North Cascades. Lacustrine deposits in the Skagit valley between Devils and May Creeks on Ross Lake suggest that a large lake occupied the valley for a considerable period (Riedel, 1985).

Because of the strong west-to-east precipitation gradient in the North Cascades, alpine glacial cover west of the Cascades crest was more extensive than on the drier east slope. The climatically controlled asymmetry in glacial cover is reflected in the contrasting geomorphology of the west versus east sides of the Skagit Valley along Ross Lake (Figure 5) (Waitt,1977). Tributary glaciers that entered the Skagit valley from the west had large enough accumulation areas to allow them to reach the Skagit, whereas those from the east did not (Waitt, 1977). As a result, streams now entering the Skagit valley from the west adjacent to Ross Lake cascade over hanging troughs (e.g.

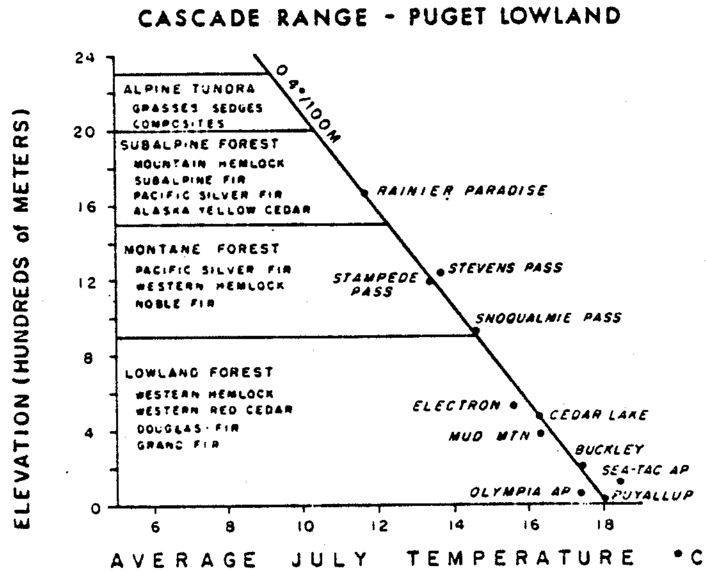
Arctic and Big Beaver creeks), while those from the east enter the Skagit Valley through narrow, water-cut canyons (e.g. Ruby, Devil's, and Lightning Creeks).

While alpine glaciers in the North Cascades were reaching their maximum downvalley extent around 18,000 years B.P., alpine glaciers in the coast mountains of British Columbia had coalesced to form piedmont glaciers and, eventually, an ice sheet (Clague, 1980). The ice sheet moved down the Georgia Depression and reached the international border around 18,000 years B.P. (Lowden and Blake, 1968; Armstrong 1977). By 14,500 to 14,000 years B.P. the ice sheet had moved down the Puget Lowland to a terminal position south of Olympia (Figure 4) (Waitt and Thorson, 1983). Distributary lobes of the ice sheet descended the upper Skagit Valley via Klesilkwa Pass (Waitt, 1977) and ascended valleys flowing to the north such as the Pasayten (Waitt, 1979). Another lobe of the ice sheet apparently ascended the lower Skagit valley at least as far as its junction with the Baker River valley (Heller, 1980) (Figure 1). As the ice sheet continued to inundate the range, it modified the pre-existing alpine glacial topography and lowered and broadened passes such as those at the headwaters of the Chilliwack, Methow, and Stehekin rivers (Waitt, 1979). It also enlarged the cross valley profiles and generally subdued the topography left by alpine glaciers below 2250 m because of its greater thickness (Waitt, 1972; Waitt and Thorson, 1983). Only the higher, more resistant massifs of the crystalline core of the range such as the Pickets, Mt. Redoubt-Mt. Spickard, Mt. Shuksan, and the Eldorado-Boston Peak areas have retained the ragged arête-horn topography of classical alpine glaciation. By 12,900 years B.P., the ice sheet had retreated north and east of Vancouver (Clague, 1980).

## E. Vegetation

Four broadly defined vegetation zones are present in the North Cascades (Figure 6). They are defined on the basis of a strong west to east precipitation gradient and altitude. From west to east,

Figure 6. Vegetational Zonation and Mean July Temperature Profiles for the Cascade Range - Puget Lowland. From Heusser (1983).



across the Cascade crest, they are the lowland forest, the montane forest, the subalpine forest, and the alpine tundra.

Vegetation within the study area can be divided into three zones based upon the age and character of the substrate, topography, and elevation. The first zone lies within the large meadow and is at the upper limit of the montane forest (Figures 2 and 6). Subalpine fir (Abies lasiocarpa) and mountain hemlock (Tsuga mertensiana) dominate better drained sites while a willow (Salix)--alder (Alnus)--sedge (Carex) community dominates the majority of the flat, poorly drained meadow. The area between the meadow and the cirque has been deglaciated since the turn of the century; its surficial geology is composed of outwash and glacial till. Subalpine fir, willow, alder and dwarf fireweed (Epilobium angustifolium) are reoccupying the area from the meadow; whitebark pine (Pinus albicaulis) and subalpine larch (Larix lyalli) are invading the upper portion of this zone from a pass to the east. The third vegetation zone surrounds the tarn in the cirque basin. It has been deglaciated within the past four decades, although considerable portions of the glacier remain at the base of the cirque walls today. The majority of this zone is void of vegetation; however, species of the Onagraceae, Cyperaceae, Compositae and Ericaciae families occupy sites close to streams and along the lakeshore in fine-textured soils.

## CHAPTER 3 : CHRONOLOGY OF LATE NEOGLACIAL GLACIER RECESSIONS IN THE NORTH CASCADES AND ON MOUNT RAINIER

### A. Previous investigations

Climatic warming at the end of the Pleistocene may have caused the disappearance of alpine glaciers throughout the North Cascades, except for glaciers on higher volcanic peaks such as Mount Baker and Glacier Peak (Matthes, 1939). Geologic and paleobotanical evidence suggest that the Hypsithermal Interval (Deevy and Flint, 1957) lasted from 10,000 to 6,000 years B.P. (Denton and Porter, 1970). Due to a lack of synchrony in events worldwide (Grove, 1979), however, the boundaries and exact timing of the Hypsithermal Interval are not well established.

Evidence suggests that alpine glaciers were still active during the Hypsithermal Interval in the North Cascades. At Glacier Peak, Beget (1981) found evidence of glacial advances around 8,300 and 6,700 years B.P.. Waitt, Yount and Davis (1980) dated the advance of an alpine glacier in the Enchantment Basin at between 8,000 and 7,500 years B.P.. On Mount Baker, Easterbrook and Burke (1972) dated advances at 6630 and 6435 years B.P..

Climatic cooling after 6,000 years B.P., during the Neoglacial period, appears to have a time parallel boundary. The Neoglacial was defined by Denton and Porter (1970) as an interval of "...rebirth or renewed growth, and all subsequent fluctuations, of glaciers after the time of maximum Hypsithermal shrinkage." In the Alps, Heuberger (1974) identifies six periods of glacial advance within the last six millennia. Denton and Karlen (1973) identify three distinct phases of alpine glacial activity in Swedish Lapland and South Alaska and Yukon Territory that occurred at 5,800 to 4,900 years B.P., 3,300 to 2,400 years B.P., and 350 to 150 years B.P.. In the North American Cordillera, Porter and Denton (1967) identified three periods of glacial resurgence, including one around 4,600 years B.P., another between 2,800 to 2,600 years B.P., and one again within the past 800 years (Figure 3).

Very few Early and Middle Neoglacial moraine surfaces have been dated in the North Cascades, presumably because late Neoglacial advances were more extensive and destroyed evidence of them. On Mount Baker, Easterbrook and Burke (1972) used tephrochronology and radiometric methods to date advances of the Boulder Glacier at 5,965, and 5,815 years B.P.. Meier (1964) used a radiocarbon sample from a buried tree to date an advance of the South Cascade Glacier at 4,700 years B.P.. Stuiver et al. (1960) found evidence of an early Neoglacial advance at 5,260 years B.P. near Mount Garibaldi, although no moraine was left by the glacier.

At Mount Rainier, in the Central Cascades (Figure 2), Crandell and Miller (1964) and Crandell (1965) radiometrically dated advances at 3,500 $\pm$ 250 and 2,040 $\pm$ 200 years B.P.. They termed the Neoglacial period the Winthrop Creek Glaciation and identified the Burroughs Mountain and Gorda stades within this period.

Late Neoglacial advances have been the most extensive of all Holocene advances in the North Cascades and on Mount Rainier. Therefore, moraines deposited during this period are well preserved and have been investigated by a number of authors (Table 1). Chronologies based on dating the timing of moraine stabilization have been reported by Porter and Denton, 1967; Leonard, 1974; Burbank, 1981; and Heikkinen, 1984.

Porter and Denton (1967) summarized data from the entire North American Cordillera and suggested there were five periods of glacier fluctuation in the past 700 years. Further, they suggested that the most recent fluctuations of glaciers occurred while they were in advanced (downvalley) positions (Figure 7). The four most recent periods were based on the dating of moraine stabilization using lichenometry, dedrochronolgy, and tephrochronology, while an earlier period (circa 600 years B.P.) was identified on the basis of indirect climatic evidence. They note that the curve is not in agreement with data from the Pacific Northwest between 1800 and 500 years B.P., when

Table 1. Previous Studies of Glacier Fluctuations in the Pacific Northwest.

| | |
|---|---|
| Mount Rainier: | Harrison, 1956; Sigafoos and Hendricks, 1961, 1962; Crandell and Miller, 1964; Burbank, 1981. |
| Mount Baker: | Long, 1953; Harrison, 1961; Easterbrook and Burke, 1971, 1972; Burke, 1972; Fuller, 1980; Heikkinen, 1984. |
| Mount Shuksan: | Leonard, 1974; Oliver and Zaszosky, 1985. |
| Dome Peak Area: | Miller,1969. |

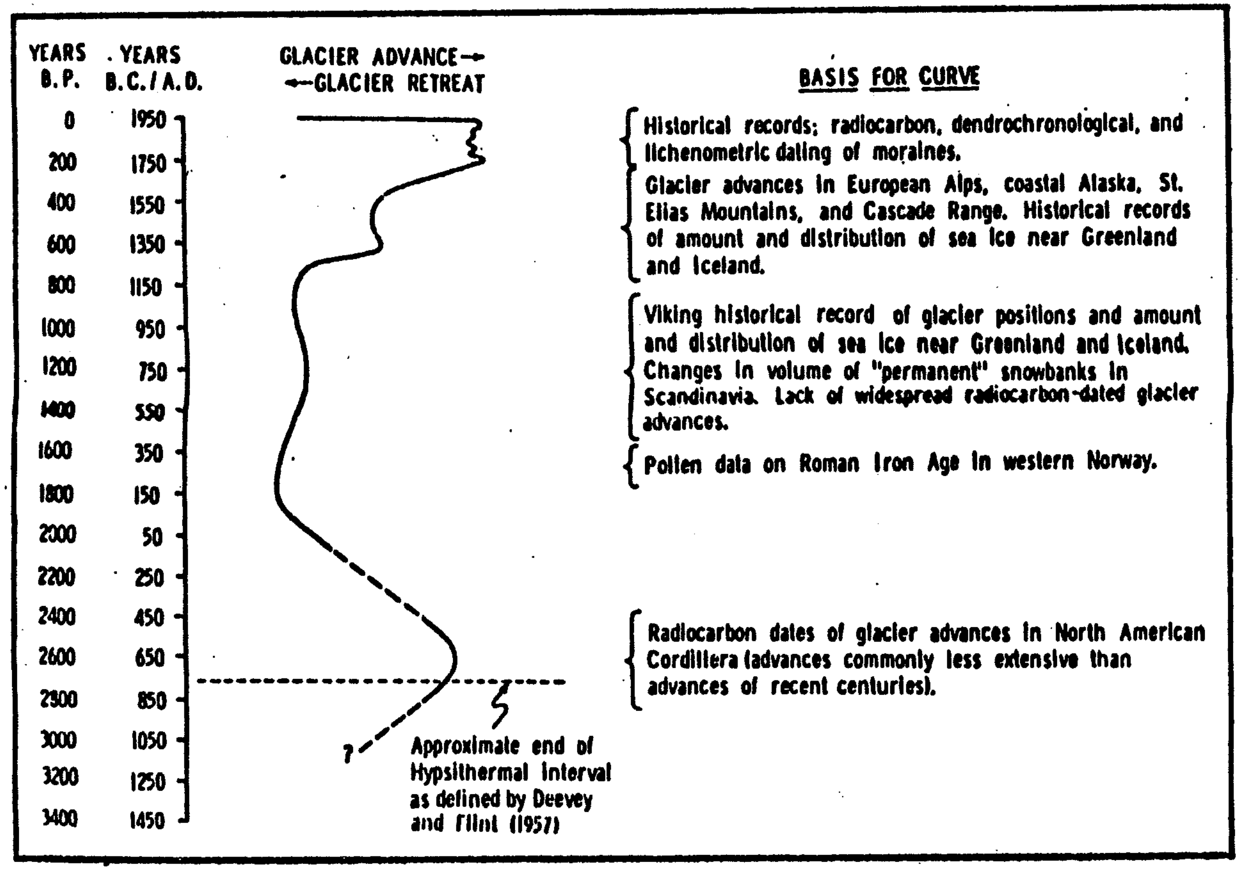



Figure 7. Generalized Curve of Glacier Fluctuations during the Neoglacial Interval. From Porter and Denton (1967).

glaciers were advancing. This suggests the possibility that there were six or more periods of fluctuation in the last 700 years. Finally, Porter and Denton (1967) noted that a lack of synchroneity in the timing of late Neoglacial maximum advances on Mount Rainier was probably due to "varied behavior patterns" of the glaciers.

Leonard (1974) summarized the late Neoglacial fluctuations of seven glaciers on Mount Shuksan, Mount Baker, Mount Garibaldi, and the Dome Peak area (Figure 1). He suggested there were five-to-six periods of "culminating glacial advances" within the past 700 years. They occurred during the 1200s, the mid to late 1400s, the mid 1500s to early 1600s, the early 1700s, the early 1800s, and the late 1800s to early 1900s. Leonard (1974) also suggests that chronologies established using dedrochronologic methods on Mount Rainier by Sigafoos and Hendricks (1961, 1972) and Crandell and Miller (1964) were not synchronous with data he summarized farther to the north.

Burbank (1981) used lichenometry as a primary dating method to establish a chronology of glacier fluctuations at Mount Rainier. He found agreement in the chronologies of five glaciers and identified 15 "periods of recession" (1328-1363, 1519-1528, 1552-1556, 1613-1623, 1640-1666, 1690-1695, 1720, 1750, 1768-1777, 1823-1830, 1857-1863, 1880-1885, 1902-1903, 1912-1915, 1923-1924). Burbank (1981) also noted that in the past 200 years, when lichenometric, rather than dendrochronologic, methods were used to date moraines, the five glaciers had more synchronous chronologies.

Heikkinen (1984) used dendrochronology to determine the timing of moraine stabilization of a series of moraines deposited by the Coleman-Rainbow Glacier complex on Mount Baker. The chronology of fluctuations he established at Mount Baker were in agreement with events over the last two centuries at Mount Rainier determined by Burbank (1981).

B. Methodology

1. **Determination of the timing of stabilization of moraines deposited by the Redoubt Glacier.**

The dating of Neoglacial fluctuations of glaciers throughout the world has been accomplished primarily by determining the time of lateral and end moraine stabilization. Stabilization begins when a glacier melts back from a moraine. Therefore, determining the time of moraine stabilization gives a minimum date for the beginning of a glacial recession.

Moraines can mark the maximum extent of a particular glacial advance (terminal moraine), but they also may record stillstands of a retreating glacier (recessional moraine). Therefore, a series of nested moraines may represent either a retreat series of a single glacial episode (one terminal moraine and a series of recessional moraines), a number of distinct glacial episodes (only terminal moraines), or a complex assemblage of both. This variability makes establishment of a chronology based on moraines difficult.

Another inherent problem in establishing a chronology based on moraines is that a particular advance may overrun and destroy evidence of all previous fluctuations - a process termed "obliterative overlap" (Gibbons et al., 1984). Obliterative overlap results in an incomplete record of glacier fluctuations based on moraines, with any single set of moraines being progressively younger upvalley. In the North Cascades and on Mount Rainier, the only moraines to survive obliterative overlap are those deposited in downvalley positions, excluding recessional moraines from the most recent advance. Records of glacier fluctuations based on moraines are also incomplete because of the destruction of moraines by fluvial, freeze-thaw, and slope processes, particularly in rugged alpine landscapes.

Tephrochronology, lichenometry, dendrochronology, and radiocarbon dating are specific methods used to determine absolute minimum and maximum time of Neoglacial moraine stabilization and deposition in the

Pacific Northwest. Relative age differences between moraines have been determined by noting the degree of soil development, weathering, and topographical relationships.

No volcanic ash or buried wood suitable for radiocarbon dating was found in the study area. This left lichenometry, dendrochronology, and relative dating techniques for determining the time of stabilization of moraines deposited by the Redoubt Glacier. Both of the botanical dating methods give absolute minimum dates of moraine stabilization. Before lichenometric and dendrochronologic sampling was undertaken, end moraines were mapped on a 1:13,750 scale map using airphotos and field observations.

a. Dendrochronology.

Lawrence (1950) outlined the procedures for using dendrochronology to date geomorphic surfaces and he identified three major problems with the method:

(1) It is impossible to obtain a core at the level of initial growth. Sigafoos and Hendricks (1962) estimated this error could be 10 years on a 60 cm diameter tree, and 50 years on a 120 cm tree.

(2) It may be impossible to sample every tree on a given surface, which introduces a potential error factor in the subjective determination of which trees are the oldest. Often, the largest or tallest trees are not the oldest.

(3) An unknown amount of time may have elapsed before germination of a tree on a newly exposed surface. Ecesis time depends upon environmental conditions such as seed availability, slope, soil texture and nutrient availability, and disturbance mechanisms. In studies using dendrochronology to determine the timing of Neoglacial moraine stabilization in the North Cascades, researchers have added from five (Sigafoos and Hendricks, 1961) to fifty (Harrison, 1956) years to the age of a tree to account for ecesis time. Burbank (1981) suggested that ecesis may even

take as long as 100 years when a seed source is 50 m away. Most estimates of ecesis time in the North Cascades range from 15 to 35 years (Long, 1953; Miller, 1969; Leonard, 1974; Heikkinen, 1984).

Another problem associated with the use of dendrochronology is the possibility that a stand of trees does not represent the first generation of growth on a given site. The presence of large dead trees, for example, may indicate that a stand did not represent the first generation of growth.

Field methods used in this study to obtain cores generally followed the procedures outlined by Lawrence (1950). Further, methodology was designed to minimize previously mentioned problems with dendrochronologic dating. Stands were examined for uniformity in age structure and the presence of dead trees to assure they represented the first generation of growth on a given site. A Swedish increment borer was used to extract cores, which were taken as close to the ground as possible all the way to the pith. Trees were selected primarily on the basis of size; however, a number of trees were sampled on each surface to verify that the oldest tree was sampled. Diameters at breast height (DBH), species type, and coring distance above the ground were recorded. Cores were mounted, sanded and stained to facilitate accurate counting of rings in the laboratory.

**b. Lichenometry.**

Lichens have been used to date the time of stabilization of moraines in the past several centuries throughout the world. Beschel (1950, 1961) pioneered this use of lichens. The application of lichenometry to date Neoglacial glacier fluctuations is based upon two critical assumptions. The first is that lichens cannot survive transportation by glacier ice and that growth is initiated at the time of moraine stabilization. The second is that after an initial period of exponential growth, lichens grow at a linear rate for several centuries. It is during this linear growth phase that lichen size can be related to the age of a substrate.

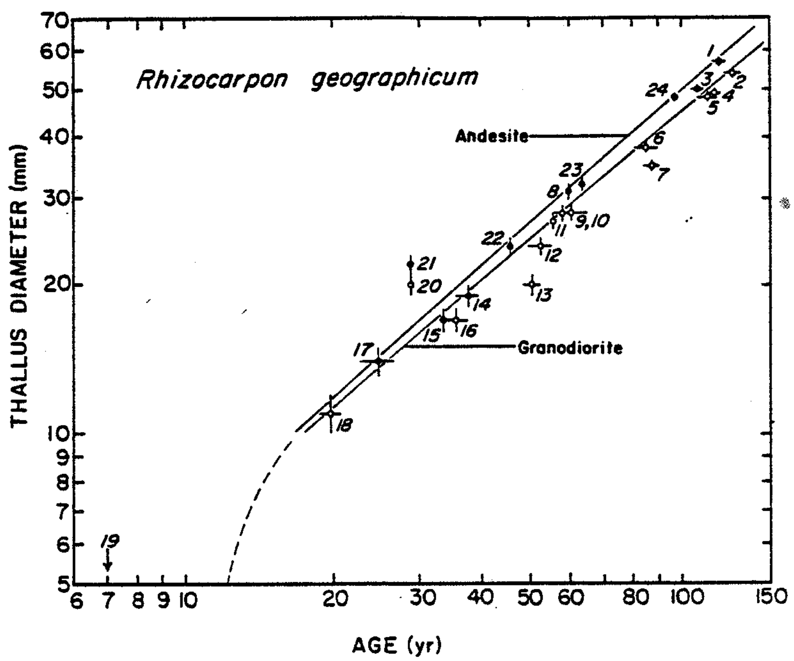
Many questions have been raised concerning the accuracy of dates determined using lichenometry. Jochimsen (1973) argued that lichen growth varies between individuals and is so complex that simple correlations between age of a substrate and lichen size are impossible. A number of factors have been identified as being critical to lichen growth rates. These include: substrate composition, shading, elevation, length of growing season, substrate stability, moisture availability, and factors inherent in lichen population dynamics such as colonization, competition and mortality (Porter, 1981; Innes, 1984). Also, perennial snowbanks may kill lichens, resulting in anomalously young populations.

Despite the questions concerning its accuracy, lichenometric dating continues to be used worldwide. Burbank (1981) and Porter (1981) have noted that recent research (e.g., Benedict, 1967; Denton and Karlen, 1973, 1977; Karlen, 1976) has established the usefulness of lichenometry in dating alpine glacial deposits. The accuracy of lichenometry in these studies was dependent upon construction of accurate growth curves.

Rhizocarpon geographicum is the species of lichen most commonly used to determine the timing of Neoglacial moraine stabilization throughout the world. Growth curves for Rhizocarpon in the Pacific Northwest have been constructed by Miller (1969), Leonard (1974), and Porter (1981). Only Porter's (1981) curve established at Mount Rainier is based on accurately dated control points. Porter (1981) used moraines dated with maps and airphotos and man-made structures to fix substrate age and establish control points (Figure 8). Leonard (1974) and Miller (1969) used dendrochronologic data to fix the age of substrates and construct their curves. Considering the previously mentioned problems with dendrochronologic dating, the accuracy of these curves is suspect.

The distance between Mount Rainier and the upper Depot Creek basin makes the application of Porter's (1981) curve questionable. Fortunately the altitude of the study area falls into the range for

Figure 8. Rhizocarpon geographicum Growth Curves at Mount Rainier. From Porter (1981).



which the curve was calibrated (1195 - 1655 m). Further, the curve is constructed for lichens growing on granodiorite, which is the most common rock type composing the moraines deposited by the Redoubt Glacier.

Maps published by Daly (1912) and Tittman and Walcott (1913) made it possible to test the accuracy of Porter's (1981) curve in the upper Depot Creek basin. A moraine associated with the 1912 1913 position of the Redoubt Glacier at 1616 m (Figure 2) was also dated using Porter's (1981) curve and a mean lichen size of 32 mm (Tables 2 and 3). The growth curve predicted the age of the moraine to be 67 years (circa 1918 ± 2), which is only a few years less than the actual age.

Lichen sampling in the study area generally followed procedures outlined by Porter (1981) and Innes (1985). Measurements were made of the maximum diameter of circular thalli of Rhizocarpon geographicum on each moraine. The ten largest thalli measured were used to compute a mean thallus size for a given moraine. Circular thalli were used to prevent measurement of two overlapping individuals. Lichens were sampled only on the distal side of moraines, away from streams and ponded water, and on granodiorite substrates to remove as much microenvironmental influence on growth rate as possible. Relief within the study area where moraines were deposited is minimal and sample sites were located well away from valley walls, so that the influence of elevation and disturbance on lichen growth rates was believed to be minimal.

Considering the uncertainties associated with dendrochronologic dating of Neoglacial moraines, Burbank (1981) suggested that lichenometric dating provided a more consistent chronology of Neoglacial activity on Mount Rainier because of shorter ecesis time and well-defined growth curves. Because of these factors and the apparent accuracy of Porter's (1981) curve in the study area, lichenometry was used as the principal dating method in this study.

Table 2. Lichenometry Sample Data for Rhizocarpon geographicum. Figures are the ten largest thallus diameters for the areas sampled, in centimeters. The final row represents the means for each area.

| Moraine #1 | Inter-Morainal Area | Moraine #2 | Moraine #3 |
|---|---|---|---|
| 3.09 | 5.40 | 5.08 | 8.59 |
| 3.09 | 4.04 | 6.03 | 8.26 |
| 2.54 | 3.05 | 6.75 | 8.26 |
| 3.81 | 5.08 | 7.62 | 7.31 |
| 3.05 | 3.89 | 7.62 | 8.89 |
| 3.09 | 3.96 | 4.85 | 8.89 |
| 3.07 | 3.89 | 5.00 | 8.72 |
| 4.14 | 4.04 | 5.64 | 8.10 |
| 3.18 | 4.67 | 5.23 | 8.52 |
| 3.25 | 4.52 | 5.72 | 8.31 |
| 3.23 | 4.25 | 5.95 | 8.31 |

Table 3. Extrapolated Ages of Rhizocarpon geographicum. From Porter (1981).

| Thallus diameter (mm) | Equivalent age (yr) (andesite substrate) | Equivalent age (yr) (granodiorite substrate) |
|---|---|---|
| 10 | 16 ± 2 | 17 ± 2 |
| 12 | 20 ± 2 | 21.5 ± 2 |
| 14 | 24 ± 2 | 25.5 ± 2 |
| 16 | 28 ± 2 | 30 ± 2 |
| 18 | 32 ± 2 | 34.5 ± 2 |
| 20 | 36 ± 2 | 39 ± 2 |
| 22 | 40 ± 2 | 43.5 ± 2 |
| 24 | 44 ± 2 | 48 ± 2 |
| 26 | 48.5 ± 2 | 53 ± 2 |
| 28 | 52.5 ± 2 | 58 ± 2 |
| 30 | 57 ± 2 | 62.5 ± 2 |
| 32 | 61 ± 2 | 67 ± 2 |
| 34 | 65.5 ± 2 | 72.5 ± 2 |
| 36 | 70 ± 2 | 77.5 ± 2 |
| 38 | 74 ± 2 | 82.5 ± 2 |
| 40 | 78.5 ± 2 | 87.5 ± 2 |
| 42 | 83 ± 2 | 92.5 ± 2 |
| 44 | 87.5 ± 2 | 97.5 ± 2 |
| 46 | 92 ± 2 | 103 ± 2 |
| 48 | 97 ± 2 | 108 ± 2 |
| 50 | 102 ± 2 | 113 ± 2 |
| 55 | 112 ± 2 | 127 ± 2 |
| 60 | 124 ± 2 | 140 ± 2 |
| Extrapolated ages | | |
| 65 | 135 ± 3 | 155 ± 3 |
| 70 | 147 ± 3 | 168 ± 3 |
| 75 | 158 ± 3 | 182 ± 3 |
| 80 | 170 ± 4 | 196 ± 4 |
| 85 | 182 ± 4 | 210 ± 4 |
| 90 | 195 ± 4 | 225 ± 4 |
| 95 | 206 ± 5 | 240 ± 5 |
| 100 | 217 ± 5 | 255 ± 5 |

## 2. Establishing a Chronology

Establishing a chronology of late Neoglacial glacier fluctuations over the past 800 years was based on four general methods. These methods were used together to identify the number, timing, and synchroneity of ice-margin recession.

The number of recessional events over the past 800 years was determined using several methods. The first method involved comparison of the timing of stabilization of moraines deposited by the Redoubt Glacier with data from 22 other valley glaciers in the North Cascades and on Mount Rainier (Table 4). Most of the data from the 22 other glaciers were obtained by dendrochronologic methods. The methods employed by the various researchers were generally consistent, although adjustment of ages to account for ecesis time and unsampled portions of trees varied.

Dates of moraine stabilization from the 23 glaciers were plotted on frequency distribution diagrams to facilitate qualitative assessment of the number, timing, and synchroneity of events. Because of possible variations in the response rates of the glaciers and the accuracy of the botanical dating methods, the timing of stabilization for a given moraine was assigned to a 20 year interval between 1200 and 1980 A.D.. Further, only one date per glacier was plotted in any single 20 year interval. If a date fell between two intervals it was assigned to the earlier interval because both botanical dating methods provide minimum dates.

The spatial scale of the recessions was examined by noting modern rates of retreat and by comparing the chronology to climatic data from Graumlich and Brubaker (1985).

After the data were plotted, the number, timing, and synchroneity of general periods of recession were determined by noting: (1) the number of moraines in any given 20 year interval; (2) the timing of late Neoglacial maximum events; (3) changes in the number of moraines between 20 year intervals; (4) whether or not different glaciers deposited moraines in adjacent periods.

Table 4. Characteristics of Glaciers in the North Cascades and on Mount Rainier. Data from Post et al. (1971), Mueller (1977), Denton (1975), and other studies listed in Table 1.

| GLACIER | LOCATION | MAX.ADVANCE (A.D.) | SIZE (km2) | ASPECT | MEAN EL. (meters) |
|---|---|---|---|---|---|
| Carbon | MRainier | 1200-1220 | 9.8 | N | 2065 |
| Chickamin | DomePk | 1200's | 4.7 | N | 2040 |
| N.Mowich | MRainier | 1320-1340 | 8.3 | NW | 2380 |
| Cowlitz | MRainier | 1360-1380 | 12.4 | SE | 2485 |
| Price | MShuksan | 1440-1460 | 1.6 | NW | 1950 |
| LeConte | DomePk | 1520-1540 | 2 | N | 2010 |
| Boulder | MBaker | 1520-1540 | 3.4 | E | 1920 |
| Coleman | MBaker | 1520-1540 | 5.2 | NW | 2070 |
| S.Tahoma | MRainier | 1520-1540 | N/A | SW | 2395 |
| S.Cascade | DomePk | 1560-1580 | 2.9 | N | 1860 |
| Tahoma | MRainier | 1620-1640 | 14.8 | SW | 2940 |
| Winthrop | MRainier | 1660-1680 | 10 | N | 2916 |
| Ohanapecosh | MRain. | 1740-1760 | 1.6 | SE | N/A |
| Emmons | MRainier | 1740-1760 | 14.8 | NE | 2930 |
| Redoubt | MRedoubt | 1760-1780 | 2.5 | N | 2100 |
| Nooksack | MShuksan | 1780-1800 | 2.9 | N | 1650 |
| Nisqually | MRainier | 1820-1840 | 7.3 | S | 2885 |
| VanTrump | MRainier | 1840-1860 | N/A | N/A | N/A |
| Puyallup | MRainier | 1840-1860 | N/A | W | 2317 |
| Dana | DomePk | N/A | 2.9 | N | 1980 |
| Deming | MBaker | N/A | 4.5 | SW | 2070 |
| Easton | MBaker | N/A | 5 | S | 2160 |
| Rainbow | MBaker | N/A | 5.6 | NE | 1710 |

N/A : data not available

Identification of the number and timing of the general periods of recession was facilitated by comparing the moraine data to a record of temperature obtained from tree rings by Graumlich and Brubaker (1985). Comparison between the climatic data and the moraine chronology was limited to the past 370 years.

A method used to establish the accuracy of the chronology was to compare data obtained from the two botanical dating methods and from different characteristics of the glaciers that may have influenced their response rates. Frequency distribution diagrams were plotted for north (315° to 45°) versus south (135° to 225°) facing glaciers, between glaciers of different sizes, and for moraines dated by dendrochronologic versus lichenometric methods. Data for these comparisons were obtained from studies that determined the time of moraine stabilization (Table 1), topographic maps, and inventories of glaciers and their characteristics compiled by Post et al. (1971), Denton (1975), and Mueller (1977) (Table 4).

Finally, a probability model was used to estimate the range in the number of periods of recession. The model was developed to act as a guide to the number of glacial episodes (cycle of advance and retreat) when only the number of surviving moraines is known. Gibbons (et al., 1984) specified that the model is applicable when:

(1) the ranks attained by moraines as a result of the relative distances of glacial advances are random over time;

(2) moraines of a succession are restricted to the same glacial pathway;

(3) the only moraines to survive obliterative overlap are those deposited at distances greater than any subsequent moraine; and,

(4) no overall trend in the extent of glaciations can be inferred.

The model is based on the probability equation

$$P(n/N) = 1/N \sum_{N=n-1}^{N-1} P(n-1/N)$$

where n equals the number of surviving moraines, N equals the number of glacial episodes, and $N \geq n \geq 2$. For example, the probability that two moraines would survive in four glacial episodes is:

$$P(2/4) = 1/4 \sum_{N=1}^{3} P(1/N)$$

$$= 1/4\ [P(1/1) + P(1/2) + P(1/3)]$$

$$= 1/4\ (1 + 1/2 + 1/3\ )$$

$$= 11/24$$

$$= .4583$$

The model is applicable to the problem of determining the number of major recessional events during the 620 years before 1820 because there is no known trend in the extent of glacial advances during this period. Variable timing for maximum downvalley extent of late Neoglacial advances by the 23 glaciers studied supports this conclusion. The model is not applicable to events after 1820 because moraines deposited during this period represent retreatal series from the last major advance of the early 1800s.

The mean number of surviving moraines from 21 glaciers was used to estimate the number of glacial episodes, rather than the number of surviving moraines from any one valley, for two reasons. First, high rates of moraine destruction and the process of obliterative overlap resulted in incomplete records in some valleys. Second, the moraine data indicates that fluctuations of glaciers in the North Cascades and on Mount Rainier over the past 700 years were synchronous, so that the mean number of moraines would represent the number of glacial episodes.

## C. Results and discussion

### 1. Fluctuations of the Redoubt Glacier

The Redoubt Glacier is classified as an alpine, temperate glacier with a surface form similar to that of both cirque and valley glaciers (Post et al., 1971). Although its terminus is presently above 1800 m, the Redoubt Glacier has advanced to elevations below 1600 m several times in the past 700 years. Three moraines were identified and examined in the upper Depot Creek basin. The moraines are located in the large meadow and along the east side of Depot Creek, where they are continuous from a trimline at 1791 m to Depot Creek (Figure 2). They are composed primarily of well rounded boulders from the Chilliwack Composite Batholith, with occasional boulders from the Skagit Volcanic and Custer Gneiss Formations.

The most extensive late Neoglacial position of the Redoubt Glacier is marked by a moraine at the southern edge of the large meadow, at an elevation of approximately 1510 m (Figure 2). This moraine represents the maximum downvalley extent of the Redoubt Glacier during Neoglacial time because no moraines were found beyond it. Fine-grained sediment has been weathered from the 1.75 m high moraine, leaving an open-work accumulation of well-rounded boulders. The time of moraine stabilization was estimated to be 210 ± 4 years B.P.(1775 AD), based on a mean thallus size of 8.3 cm from the ten largest thalli of Rhizocarpon geographicum found on the moraine (Tables 2 and 3). Dendrochronologic dating of the surface immediately upvalley of the moraine suggests this surface is at least 111 years old (Table 5). Considering the previously mentioned problems with the accuracy of dendrochronologic dating methods, however, this date could be too young.

A moraine 2.5 m high stands approximately 40 m behind the 1775 moraine, and is located at the southeastern edge of the meadow (Figure 2). As the moraine nears the trimline it appears to merge with its older counterpart. The upper end of the 1775 moraine, however, was

Table 5. Dendrochronologic Sample Data.

| Sample #/ Location | Species | DBH (cm) | Coring Distance Above Ground (cm) | # rings |
|---|---|---|---|---|
| 1 / IM | SAF | 37.1 | 29 | 45 |
| 2 / IM | SAF | 73.0 | 31 | 60 |
| 3 / M2 | SAF | 135.0 | 100 | 100 |
| 4 / M2 | MH | 120.0 | 45 | 92 |
| 5 / M3 | MH | 160.0 | 45 | 93 |
| 6 / M3 | SAF | 138.0 | 49 | 111 |
| 7 / M3 | SAF | 140.0 | 50 | 96 |

Location (cf. Table 2):

IM : inter-morainal area

M2 : moraine #2

M3 : moraine #3

Species:

MH : Mountain Hemlock

SAF : Subalpine Fir

destroyed by a recent rockfall that separates the two moraines near the meadow. Like the 1775 moraine, this moraine is composed of an open-work accumulation of boulders, although it does not appear as weathered. The minimum age of the moraine was determined to be 140 $\pm$ 2 years (1845 AD) using a mean lichen size of 5.9 cm (Tables 2 and 3). The oldest tree found on the moraine is a 100 year old subalpine fir (Table 5). Assuming that the lichen date is accurate, the age of this tree suggests that the ecesis time was shorter than for the trees behind the earlier moraine and more of the initial growth rings were sampled.

An intermorainal surface upvalley from the 1845 moraine was also dated. Using lichenometry this surface was estimated to be 93 $\pm$ 2 years old (Tables 2 and 3), while the oldest tree cored on the same surface was a 60 year old subalpine fir (Table 5). If the lichenometrically derived date is accurate, the dendrochronologic date also underestimates the age of this surface.

The most recent advance of the Redoubt Glacier is marked by a moraine approximately 30 m upvalley from the 1845 moraine at an elevation of 1552 m. The till in the moraine has a fresh, unweathered appearance and a greater fine-grained component than either of the older moraines. The moraine is approximately 1 m high and can be traced from Depot Creek to the trimline, (Figure 2). Maps published by Tittman and Walcott (1913) and Daly (1912) were used to fix the age of the moraine at approximately 75 years B.P..

## 2. Chronology of late Neoglacial glacier recessions in the North Cascades and on Mount Rainier

Fluctuations of the Redoubt Glacier over the past 200 years are, in general, synchronous with those of other glaciers in the North Cascades and on Mount Rainier (Figure 9). The chronology of recessional events since 1820 of all glaciers studied is well documented, while events between 1500 and 1820 are less so, and those prior to 1500 are very poorly documented (Figure 9).

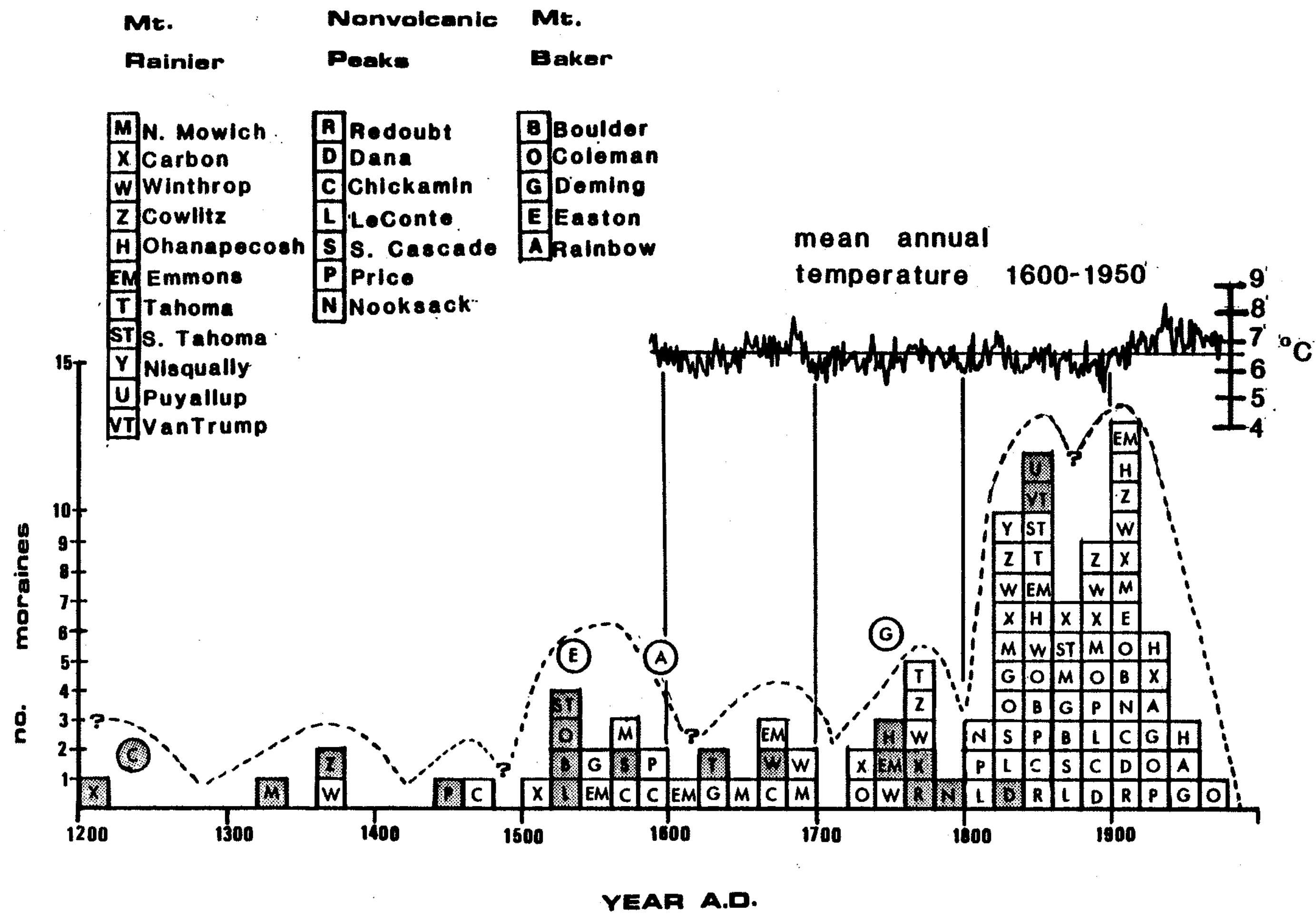



Figure 9. Timing of Moraine Stabilization of 23 Glaciers in the North Cascades and on Mount Rainier over the past 800 Years. Each square represents one moraine; patterned squares indicate late Neoglacial maximum downvalley extents; circles indicate approximate dates. Inset of mean annual temperature from tree rings from Graumlich and Brubaker (1985).

Periods of glacial activity prior to 1820 are poorly documented because more recent advances have destroyed evidence of them. Approximately one-third of the 23 glaciers studied reached late Neoglacial maximum extents after 1720 (Figure 9). Most of the other glaciers were at or near late Neoglacial maxima in the late 1700s and early 1800s (Miller, 1969; Sigafoos and Hendricks, 1962; Leonard, 1974; Burbank, 1981).

Eight periods of recession are identified in the past 800 years through examination of the timing of moraine stabilization (Figure 9) and comparison of this chronology with climatic data (Graumlich and Brubaker, 1985).

Poorly documented recessional phases occurred during the early-1200s, mid-1300s, and mid to late-1400s (Figure 9). Five moraines deposited between 1200 and 1500 survived subsequent advances because they represent late Neoglacial maximum downvalley extents of the Chickamin (Miller, 1969), Carbon (Sigafoos and Hendricks, 1962), North Mowich (Burbank, 1981), Cowlitz (Crandell and Miller, 1964), and Price (Leonard, 1974) Glaciers.

In contrast to the preceding 300 years, events between 1500 and 1820 are better documented (Figure 9), and three periods of recession are identified. Distinguishing the boundaries and exact timing of the first two periods is problematical, however, because moraines were deposited and stabilized in every 20-year interval between 1500 and 1700.

Between 1500 and 1600 13 glaciers deposited moraines, with the most synchronous period of moraine stabilization occurring between 1520 and 1540 (Figure 9). Cooler temperatures during the early 1600s (Graumlich and Brubaker, 1985) may have caused glaciers to readvance during this period. At Mount Rainier, Burbank (1981) suggested there were two periods of recession, from 1519 to 1528 and from 1552 to 1576. His two separate periods are considered as one here because several glaciers melted back from moraines between 1528 and 1552 and because none of the glaciers deposited moraines in both periods.

Evidence of cooler temperatures in the early 1600s and warmer temperatures in the late 1600s (Graumlich and Brubaker, 1985) was used to distinguish a period of recession in the mid-to-late 1600s. During this period six glaciers melted back from moraines, two from late Neoglacial maxima (Figure 9). The most synchronous interval of recession occurred between 1660 and 1680, when three moraines were deposited and stabilized. Multiple moraines were deposited by the Winthrop and North Mowich Glaciers, which indicates wastage was not continuous (Figure 9). This period of recession is presumed to have ended around 1700 because no moraines were deposited between 1700 and 1720 and because temperatures cooled between 1700 and 1760 (Graumlich and Brubaker, 1985).

A sixth period of glacier wastage occurred between 1720 and 1800, when 12 of 23 glaciers melted back from moraines (Figure 9). This period is distinguished from those after and before it because four of the glaciers that melted back from moraines also abandoned moraines in the previous and succeeding periods. Recessional responses of the glaciers that deposited moraines during this period were most synchronous between 1760 and 1780 (Figure 9). Five of the glaciers deposited moraines at their late Neoglacial maximum downvalley positions.

A broad, well documented period of recession has occurred since 1800, when all 23 glaciers deposited and melted back from moraines (Figure 9). Tree-ring data (Graumlich and Brubaker, 1985) and the timing of moraine stabilization (Figure 9), however, indicate that recession was punctuated by a period of cooler temperatures and, possibly, glacier expansion around 1870. Therefore, this broad period of recession is divided into two periods.

Highly synchronous periods of moraine abandonment occurred between 1820 and 1860, and 1880 and 1920, when 20 and 16 glaciers, respectively, melted back from moraines (Figure 9). Burbank (1981) recognized five periods of glacial recession between 1800 and 1920. These five periods fall within the range of the 1820-1860 and 1880-1920 periods identified in this study.

Differences in the timing of late Neoglacial maxima of 23 glaciers in the Pacific Northwest probably reflects the sensitivities of individual glaciers to variations in climate. The lack of a more synchronous chronology probably reflects both the individual responses of glaciers and differences between and uncertainties inherent in the two botanical dating methods. For this reason chronologies established using the two dating methods were compared, as were chronologies between glaciers of different, size and aspect.

Advances of glaciers since 1720 were generally the most extensive during the past 800 years. The process of obliterative overlap associated with the recent advances, however, has not been complete in every valley, which allowed use of a probability model to estimate the number of glacial episodes based on the number of surviving moraines. The mean number of surviving moraines below 21 glaciers in the Pacific Northwest that stabilized between 1200 and 1820 is 2.24 (Figure 9). The highest probability that approximately 2 moraines would survive obliterative overlap is associated with 2-3 glacial episodes (Table 6). The probability that two moraines would survive, however, remains relatively high until 10 to 15 glacial episodes occur (Table 6). Based on these relationships, the mean number of surviving moraines in 21 valleys indicates there were probably between 2 and 10 glacial episodes (including advances and recessions) between 1200 and 1820. This range in the number of glacial episodes is generally in agreement with the eight previously identified periods of recession.

A lack of data, particularly for moraines dated by lichenometry, precluded determining if either of the two botanical dating methods produced a more synchronous chronology (Figure 10). Some general conclusions concerning the utility of these methods, however, may be drawn.

First, dendrochronology allows dating of moraines that are older than 400 years, although sampling problems and variable ecesis times make interpretation of dendrologic data difficult, especially for older, larger trees (Sigafoos and Hendricks, 1962). Lichenometric determination of the timing of moraine stabilization for periods greater than 400 years is not possible because these moraines often

Table 6. Probabilities of n Surviving Moraines Given N Glacial Episodes using Probability Model described in Chapter 3.

| N | n=1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 |
|---|---|---|---|---|---|---|---|---|---|---|
| 1 | 1.0000 | ---- | ---- | ---- | ---- | ---- | ---- | ---- | ---- | ---- |
| 2 | 0.5000 | 0.5000 | ---- | ---- | ---- | ---- | ---- | ---- | ---- | ---- |
| 3 | 0.3333 | 0.5000 | 0.1667 | ---- | ---- | ---- | ---- | ---- | ---- | ---- |
| 4 | 0.2500 | 0.4583 | 0.2500 | 0.0417 | ---- | ---- | ---- | ---- | ---- | ---- |
| 5 | 0.2000 | 0.4167 | 0.2917 | 0.0833 | 0.0083 | ---- | ---- | ---- | ---- | ---- |
| 6 | 0.1667 | 0.3806 | 0.3125 | 0.1181 | 0.0208 | 0.0014 | ---- | ---- | ---- | ---- |
| 7 | 0.1429 | 0.3500 | 0.3222 | 0.1458 | 0.0347 | 0.0042 | 0.0002 | ---- | ---- | ---- |
| 8 | 0.1250 | 0.3241 | 0.3257 | 0.1679 | 0.0486 | 0.0080 | 0.0007 | 0.0000 | ---- | ---- |
| 9 | 0.1111 | 0.3020 | 0.3255 | 0.1854 | 0.0619 | 0.0125 | 0.0015 | 0.0001 | 0.0000 | ---- |
| 10 | 0.1000 | 0.2829 | 0.3232 | 0.1994 | 0.0742 | 0.0174 | 0.0026 | 0.0002 | 0.0000 | 0.0000 |
| 15 | 0.0667 | 0.2168 | 0.2999 | 0.2378 | 0.1221 | 0.0433 | 0.0110 | 0.0021 | 0.0003 | 0.0000 |
| 20 | 0.0500 | 0.1774 | 0.2748 | 0.2508 | 0.1527 | 0.0664 | 0.0215 | 0.0053 | 0.0010 | 0.0002 |

*Note:* Probabilities of survival of moraines in a sequence assuming ranking of moraine distance from source area is random through time. $N$ = 10 is the number of glacial-interglacial oscillations in the last 0.9 m.y.

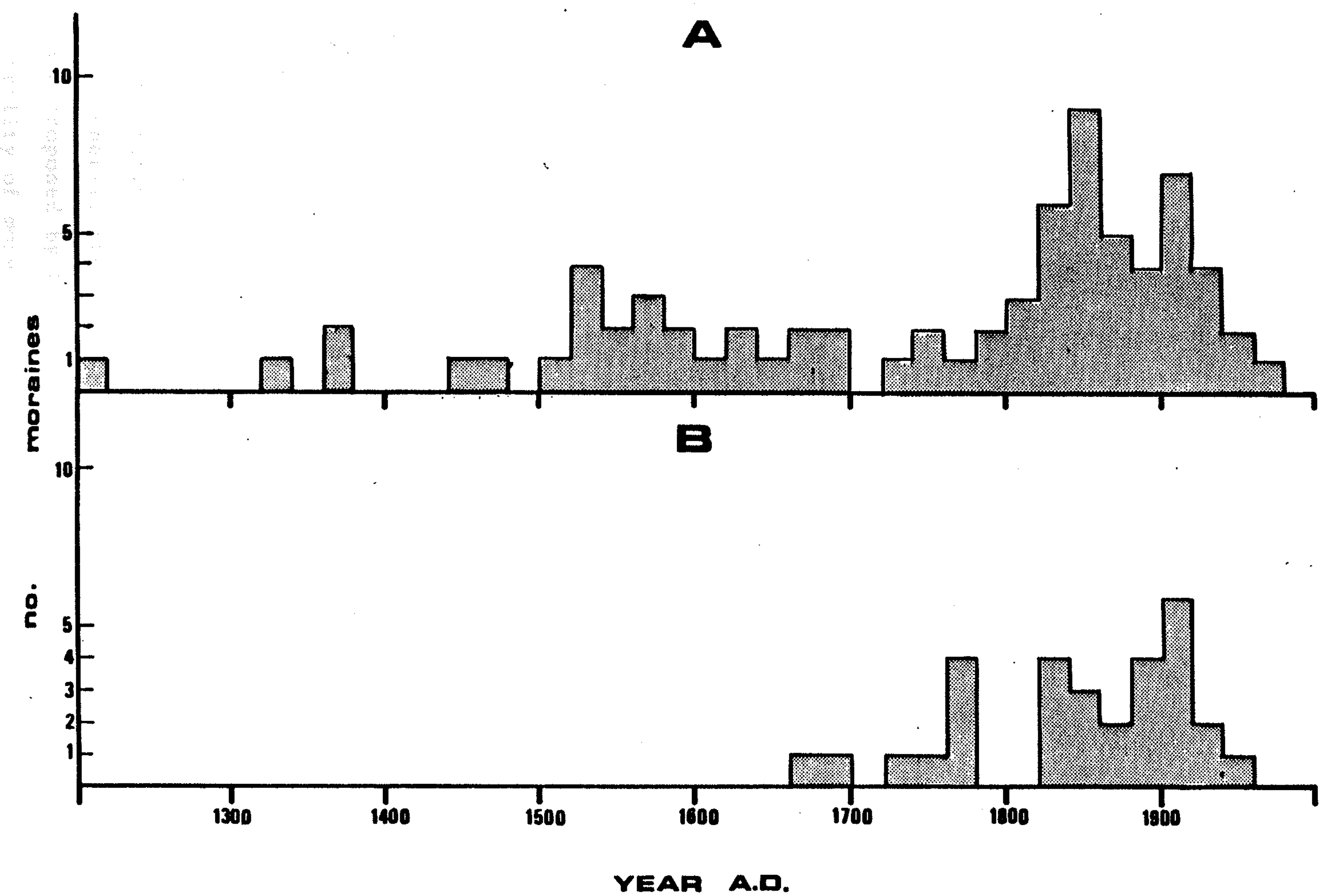



Figure 10. Timing of Moraine Stabilization as determined by (A) dendrochronologic and (B) lichenometric methods.

lie below the elevation that Porters (1981) curve is calibrated for (Burbank, 1981), and because there is a lack of reliable control data for that segment of the growth curve.

Second, when lichenometric dates are determined using accurate growth curves and within the period of linear growth rates, they provide a more accurate measure of a substrates minimum age than dendrochronologic dates. Comparisons made between the two methods by Burbank (1981) and in this study suggest that dendrochronology underestimates the age of a moraine. Further, the underestimation of age by dendrochronologic methods appears to be greater as substrates become older.

Comparisons were also made between the chronologies of north versus south facing glaciers (Figure 11). No discernable difference, however, could be made concerning the synchroneity or timing of these two classes of glaciers.

Glaciers of varying size appear to have responded synchronously during recessional events over the past 800 years, although this comparison is limited to events since 1800 by a lack of data (Figure 12). Larger glaciers in general reached late Neoglacial maxima earlier than their smaller counterparts (Table 4), which suggests that glacier size affects the timing and magnitude of glacier response to climatic forcing.

Identification of approximately eight major periods of glacial recession in this study during the last 800 years contrasts with the 15 periods identified by Burbank (1981) on Mount Rainier. Burbank's (1981) 15 periods of recession all fall within the eight broader periods identified in this study. The difference in the number of periods reflects either smaller temporal scale fluctuations (ie., retreatal series) or a bias in Burbank's data toward larger glaciers on Mount Rainier.

In general, the results of this study agree with the five to six periods proposed by Porter and Denton (1967) and Leonard (1974). The availability of more data in this study, however, allowed definition of three more periods during the 1300s, 1600s, and 1800s not

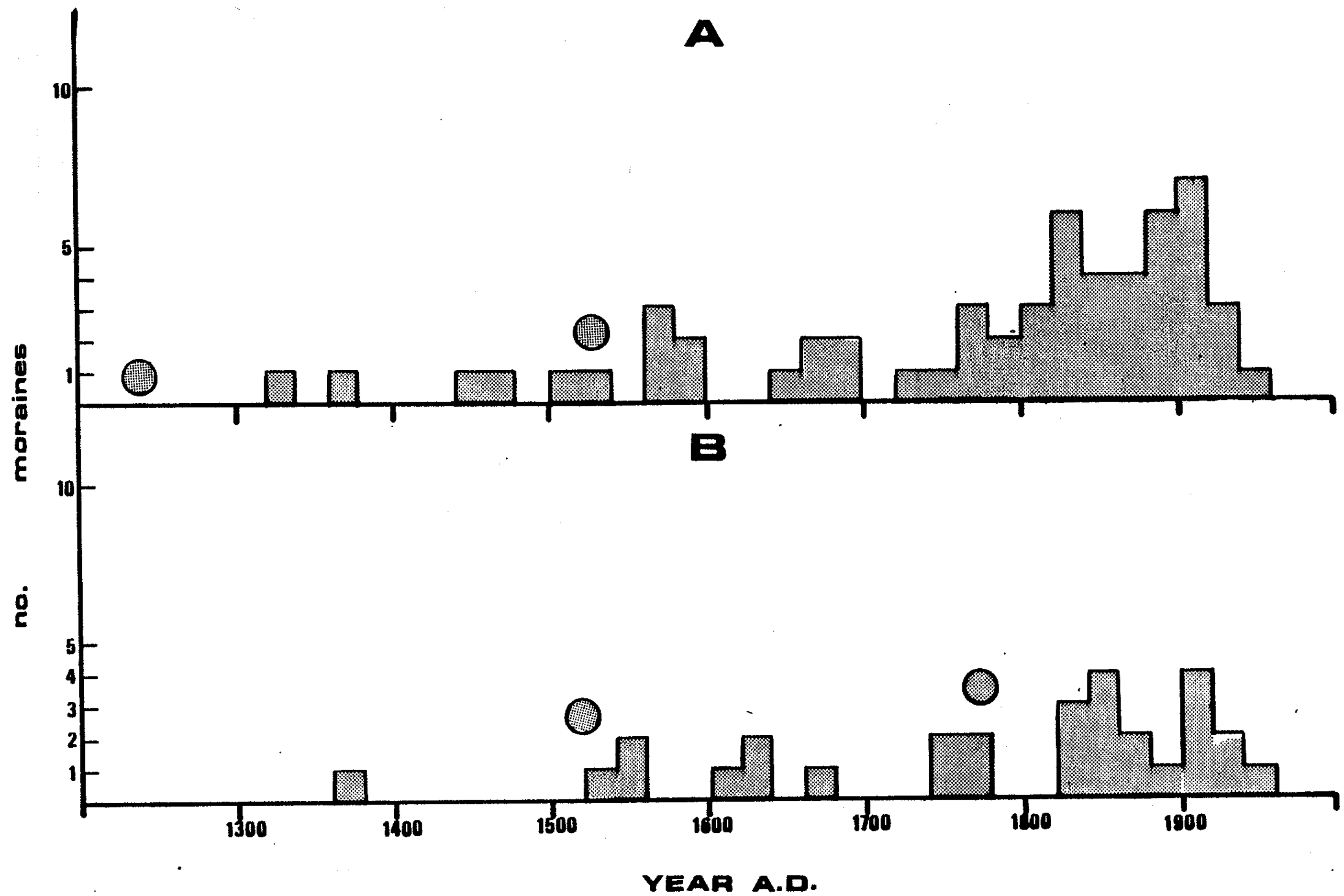



Figure 11. Timing of Moraine Stabilization of (A) North and (B) South facing Glaciers. Circles indicate approximate dates.

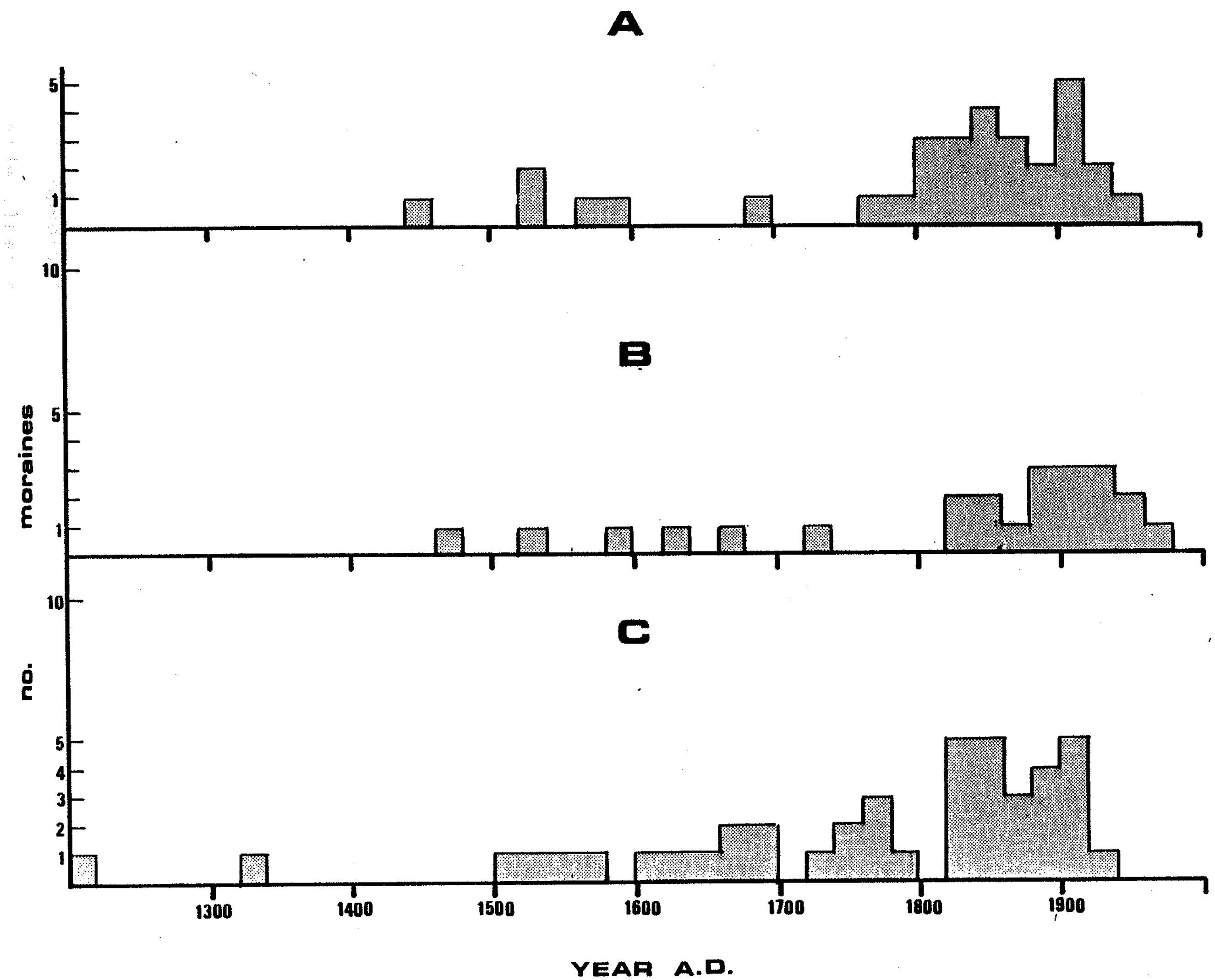



Figure 12. Timing of Moraine Stabilization of Glaciers of Different Sizes: (A) 1.6-3.4 km$^2$; (B) 3.4-5.6 km$^2$; and (C) > 7.3 km$^2$.

recognized by Leonard (1974) (Figure 9). In contrast to Leonard's (1974) study, glacial fluctuations on Mount Rainier were found to be generally synchronous with those in the North Cascades (Figure 9). More accurate lichenometric dating of moraines on Mount Rainier by Burbank (1981), which were used in this study but not in Leonard's (1974), are probably responsible for a more synchronous chronology between the two regions.

The precise spatial and temporal scales of late Neoglacial glacier recessions in the North Cascades and on Mount Rainier are impossible to determine because moraines provide a limited amount of information concerning glacier fluctuations. Porter and Denton (1967) suggested that fluctuations during the last 800 years were superimposed on a spatially and temporally broader scale fluctuation beginning around 800 years ago and ending within the last century (Figure 7). In this case, the variable timing of late Neoglacial maxima reflects individual responses of glaciers to relatively smaller-scale climatic signals.

On the other hand, rapid rates of retreat of glaciers during the twentieth century and evidence of warmer episodes comparable to those associated with rapid modern retreat (Graumlich and Brubaker, 1985) (Figure 9) suggest that glaciers may have retreated hundreds of meters several times over the last 800 years. Glaciers must then have advanced well downvalley again to deposit moraines that survived destruction by subsequent advances. Whether or not these recessions are comparable to the wide-spread retreat of glaciers during the last century is impossible to determine solely on the basis of moraines. Evidence indicates, however, that twentieth-century retreat has ended and that some glaciers in the North Cascades and on Mount Rainier began advancing in the 1950s (Meier, 1965; Post et al., 1971). Therefore, observations of glacier activity within the next century may provide clues as to the scale of recessions during the past 800 years.

## 3. Summary

The results of this study indicate that eight major periods of glacier wastage occurred in the North Cascades and on Mount Rainier over the past 800 years. Obliterative overlap and a general lack of data for periods prior to 1500 made it difficult to determine the exact number and timing of these earlier periods. Nonetheless, glaciers melted back from moraines during the early 1200s, mid-1300s, mid to late-1400s, between 1520 and 1580, mid to late-1600s, between 1740 and 1780, between 1820 and 1860, and between 1880 and 1920. Determining the duration of these periods was problematical because moraines stabilized in every 20-year interval except one between 1500 and 1980. The timing of these periods is in agreement with climatic data. Although the responses of glaciers were in general synchronous, variable timing of late Neoglacial maxima indicates response is also somewhat individual. In contrast to earlier studies, the fluctuations of glaciers on non-volcanic peaks were found to be synchronous with those of larger glaciers on higher volcanic peaks such as Mount Rainier and Mount Baker.

Lichenometric dating of moraine stabilization seems to provide a more accurate chronology than dendrochronology, although comparison of these two methods was limited due to a lack of data. Finally, a lack of data also inhibited identification of the influence of glacier size and aspect on the chronology. Larger glaciers, however, generally reached late Neoglacial maxima earlier than their smaller counterparts.

The exact spatial and temporal scale of glacier recessions is impossible to determine, although evidence suggests that they may have retreated hundreds of meters during warmer episodes.

## 4. Implications of this study

The identification of eight periods of glacier wastage and the rapid retreat of glaciers since 1900 indicates that glaciers in the

North Cascades and on Mount Rainier can be useful indicators of climatic change. Further, knowledge of the fluctuations of glaciers in the Pacific Northwest can provide valuable information to hydrologists, resource managers, land use planners, botanists, archaeologists, and climatologists.

Before more detailed relationships between climatic change and glacier variations can be established, however, more valley glaciers in the Pacific Northwest must be studied. Such areas in the North Cascades include Glacier Peak and the McAllister Glacier. Further, more accurate, consistent dating methods will need to be developed.

# CHAPTER 4 : ESKER GENESIS IN THE UPPER DEPOT CREEK VALLEY

## A. <u>Previous investigations</u>

Banerjee and McDonald (1975) defined an esker as "...a linear accumulation of gravelly and/or sandy stratified sediment that was deposited by a stream confined on both sides by glacier ice." In general, eskers are deposited during late stages of deglaciation when meltwater flow stabilizes in tunnels, ice movement is minimal, and a considerable volume of meltwater is available to erode, transport, and deposit sediment.

Despite over 100 years of investigation, many questions remain concerning the distribution of eskers and the processes responsible for their deposition and form. There are two reasons why the origin of eskers is still a subject of controversy. First, most eskers form in tunnels beneath glacier ice that are, for the most part, inaccessible. Second, as Flint (1928) noted, eskers are polygenetic; they form under highly variable sedimentological conditions in the ice-contact environment and therefore exhibit wide variations in form, structure, and distribution. The overall variability of eskers has made it impossible for any single theory to explain all eskers. Further, the characteristics of eskers have led to their confusion with other linear accumulations of stratified sand and gravel, such as crevasse fills and moraines--particularly when eskers are defined in a morphologic, rather than morphogenetic, sense (Lundqvist, 1979).

During over 100 years of investigation, two main models of esker genesis have been proposed and subsequently developed. The first was by Hummel (1874), who suggested that eskers form along the beds of subglacial streams (Lewis, 1949). In 1884, Shaler initially proposed a second model that related esker deposition to deltaic settings.

Shaler's (1884) model was adopted and developed by DeGeer (1897), who observed downstream changes in grain size from sand and gravel to laminated silt and clay in eskers. More recently, Banerjee and McDonald (1975) related this model of deposition to a segmented esker ridge morphology.

Hummel's (1874) theory has become the most widely accepted and discussed model of esker formation. Mannerfelt (1945) suggested a variation of this idea when he defined subglacially engorged eskers, which form in topographic settings where water flows from a hill slope back under glacier ice. This type of esker is typically shorter and straighter than other eskers (Sugden and John, 1976, p.331). Saunderson (1975) and Banerjee and McDonald (1975) identified two variations on Hummel's general model, which were based on whether or not meltwater completely filled a tunnel (closed conduit) or flowed under atmospheric pressure (open channel).

Studies by Shreve (1972; 1985), Rothlisberger (1972), and Hooke (1984) focused on features of glacial hydrology such as ice thickness, ice surface and bed gradients, fluctuations in water discharge, and topography. Through theoretical calculations and field observations they suggested how these factors interact to influence tunnel shape, stability and drainage patterns beneath glaciers. They concluded that excess energy from meltwater discharge can maintain an open channel under active ice conditions (Hooke, 1984); hydraulic gradient and ice thickness determines whether or not meltwater channels are braided, straight or meandering (Shreve, 1985); channelized movement of meltwater at the base of a glacier is more stable than sheetlike drainage; and, given enough time, the network of subglacial meltwater tunnels assumes an arborescent pattern (Shreve, 1972).

Studies of eskers in modern glacial environments are rare in comparison to studies of much larger Pleistocene eskers. Nonetheless, these studies have added a great deal to our understanding of process-form relationships in the genesis of eskers. Price (1966) studied eskers forming near the margin of the Casement Glacier in Glacier Bay, Alaska. Using air photos and field evidence he proved that eskers could be deposited in englacial and supraglacial positions over several tens of meters of ice and still retain the classical esker form upon subsequent ice melt-out. He also noted, however, that preservation of eskers deposited in these positions was dependent on the amount of sediment and the manner of ice disintegration. Price (1966) also observed that supraglacially and englacially deposited eskers often had ice cores, which caused the splitting of a single esker ridge in two as they melted. Also at Glacier Bay, Price (1966) and Mickelson (1971) observed that esker complexes (also known as esker nets and braided eskers) formed where intricate, interconnected drainage networks developed in large, stagnant ice masses in the lee of nunataks. Lundqvist (1979) noted that esker nets also form where subglacially engorged eskers merge in the bottom of valleys. Mickelson (1971) also showed how topography controlled the sedimentology of an esker by influencing stream power near the margin of the Burroughs Glacier in southeastern Alaska. Where water flowed downhill the esker is coarse grained, whereas water that flowed uphill under hydrostatic pressure into an ice-marginal lake formed a fine-grained esker composed of laminated silt and sand. Stokes (1958) studied an esker ridge in Norway that formed in a subglacial tunnel as englacial and supraglacial debris melted out. The tunnel was also occupied by a meltwater stream, but Stokes concluded that the sediment composing the esker was deposited directly from glacier ice. In Iceland, Howarth (1971) observed an esker ridge forming as adjacent outwash surfaces were lowered by buried ice melt-out. He related the presence of the tunnel, which the esker ridge had filled, to lines of weakness in the ice associated with splaying crevasses.

Although most eskers appear to be deposited in subglacial tunnels either at the beds of streams or in standing water, a largely unexplained feature of eskers is their distribution. Sugden and John (1976, p. 330) have suggested that eskers represent a fraction of total subglacial meltwater conduits because erosion or non-deposition is more common than deposition in meltwater conduits.

A number of complex, inter-related factors are probably responsible for the distribution of eskers in glaciated landscapes. These might include preservation, the availability of sediment, the structure of an ice mass, subglacial topography, and the unstable, episodic nature of deposition in meltwater conduits.

Conditions that lead to deposition are related to the ponding of water, bedrock topography, the availability of sediment, changes in the morphology of conduits, fluctuations in stream power, and the migration of meltwater to different drainage paths.

Church and Gilbert (1975) recognized five seasonally controlled periods of runoff in glacial meltwater streams. They are: winter, spring breakup, nival flood late summer, and freezeback. Smith (1985) notes that high sediment transport rates occur during the nival flood and late summer periods. Deposition is generally associated with the falling limb of the seasonal hydrograph during the late summer period.

Precipitation events and diurnal fluctuations in the rate of ice and snow melt also influence the rates of sediment transport and deposition (Church and Gilbert, 1975). Meier et al. (1971) and Tangborn et al. (1977) showed that precipitation events controlled peaks in discharge in the late summer and freezeback periods that equalled those of the nival flood at the South Cascade Glacier in Washington.

In general, most eskers are deposited in topographic lows beneath temperate, warm-based glaciers (Banerjee and McDonald, 1975; Lundqvist, 1979). Eskers deposited on steep slopes (i.e. subglacially engorged eskers) and in englacial and supraglacial positions are less likely to be preserved because of the meltout of supporting ice cores and destruction by slope processes.

Many researchers have related the presence of eskers to structural features of glaciers, which control meltwater drainage patterns (Stenborg, 1968; Shreve, 1972). These features include interlobate zones, shear planes, crevasses, and medial moraines.

The availability of sediments is probably also an important factor controlling the distribution of eskers. Stokes (1958) and Howarth (1971) related the deposition of eskers to large concentrations of englacial and supraglacial debris. Mickelson (1971), Pessl and Fredrick (1981), and Gustavson and Boothroyd (1982) suggested that subglacial till is the primary source of sediment for fluvioglacial deposition. Pessl and Fredrick (1981) also noted the importance of supraglacial debris from valley walls as an important debris source for valley glaciers. Finally, Mickelson (1971) observed that sediment composing eskers deposited in the lee of nunataks was derived locally from till surfaces on the stoss sides of the nunataks.

Studies of eskers in modern glacial environments have identified process-form relationships in the deposition of eskers. Questions have been raised, however, concerning the application of evidence from relatively small eskers deposited in modern glacial environments to larger Pleistocene eskers (Shaw, 1972; Banerjee and McDonald, 1975).

The unusual occurrence of a 120 m long, fine-grained esker in a cirque basin in the North Cascade Mountains of Washington provides an excellent opportunity to study esker genesis. The deglaciation of the region is well documented by seven aerial photographs taken at various times since 1947, and the somewhat unusual existence of this esker in an alpine setting indicates that an uncommon combination of factors were responsible for its genesis.

## B. Methodology

Evidence gathered in the field during the summer of 1985, examination of air photos and topographic maps, comparisons with other cirques in the North Cascades, laboratory analyses of sediment from the esker, and a review of pertinent literature were used to evaluate the nature of esker formation in the upper Depot Creek Basin.

Information gathered from air photos taken in 1947, 1950, 1956, 1958, 1966, 1967 and 1978 and maps published in 1906, 1913, 1947 and 1955 were used to identify features of the Redoubt Glacier that may have influenced formation of the esker. These features include surficial accumulations of debris, ice-flow directions, the timing of stagnation, ice surface features such as moulins and crevasses, the location of meltwater channels, and the sequence of deglaciation of the cirque.

A majority of the 35 days in the field was spent constructing a 1:1,600 topographic map of the upper portion of the study area. A suitable base map was unavailable because ice occupied the cirque when the 1:62,500 U.S. Geological Survey topographic map was made in 1947. The 1947 map (Challenger Quadrangle) and an airphoto taken in 1978 were used to make a base map and fix the location of a lake within the cirque. The recently deglaciated terrain adjacent to the lake was mapped at a 2 m contour interval using a Brunton Compass mounted on a tripod and a 50 m tape. The elevations of points within the cirque were calculated using trigonometry. An elevation of 1701 m for Ouzel Lake was used as the base elevation for calculating the elevation of other points and was taken from a Pictour alpine guide photograph taken in 1974.

After completion of the base map, the surficial geology and geomorphology of the cirque floor were surveyed to identify features deposited during deglaciation that may have been related to deposition of the esker. Landforms such as kames, meltwater channels, moraines, and kame terraces were mapped and described while traversing the cirque floor on foot.

Detailed morphologic and sedimentologic analyses of the esker were undertaken once the base map and field descriptions of the geology and geomorphology of the cirque were completed. Morphologic features of the esker measured included plan shape, height, width, angle of side slopes, length, and profile. Excavations were made in five places in the esker to examine its internal structure and sedimentology. Distinction of different sedimentary units were made in the field on the basis of sedimentary structure, stratigraphic position, and texture. Lateral, vertical, and longitudinal relationships between sedimentary units were recorded in detailed sketches and with photographs. Samples were taken from different units within each pit. The original shape of the esker was restored by backfilling the excavations, which resulted in only a small loss of crest height.

Fifteen samples were returned to the laboratory for analysis of lithology, sorting, roundness and grain-size distribution. The greater than 2 phi and less than 4 phi fractions were removed by dry and wet sieving, respectively. The remaining fractions were sorted at 0.5 phi intervals in a sonic sifter. Cumulative frequency graphs were plotted and graphical-moment statistics were calculated on an Apple IIe computer with a program developed by Knox and Bartlein (1976).

Field distinction of sedimentary units and longitudinal, vertical, and lateral sedimentological relationships were then quantitatively checked using properties of the samples identified in the lab.

Processes responsible for the deposition and form of the esker were identified from morphologic features of the esker, textural and structural properties of the sedimentary units within the esker, and spatial sedimentological relationships. Interpretations were based on both empirical and experimental studies of fluvial (Simons and Richardson, 1962; Harms and Fahnestock, 1965; Visher, 1965; Jopling, 1966; Jopling and Walker, 1968), proglacial fluvial (Church and Gilbert, 1975), and fluvioglacial systems (Stokes, 1958; Price, 1966; Shreve, 1972; Shaw, 1972; Saunderson, 1975, 1977, 1980; Banerjee and McDonald, 1975; and Ringrose, 1980).

Identification of factors responsible for the deposition and location of the esker in this unusual alpine setting were made by direct comparison of the Redoubt cirque with other cirques in the North Cascades.

## C. Results and Discussion

### 1. Deglaciation of the upper Depot Creek basin

Since 1910 the Redoubt Glacier has retreated approximately 1.3 km up the Depot Creek valley. Between 1910 and 1950 retreat was along a single margin, following a pattern similar to that of most alpine valley glaciers (Sugden and John, 1976). The absence of recessional moraines and the consistent slope of that portion of the valley suggest frontal retreat was steady, at a rate of about 15 m/year. The major geomorphic features and surfical deposits of this area are characterized by a few outcrops of glacially scoured bedrock, local accumulations of till up to 5 m thick, and ice-marginal channels.

Deglaciation of the cirque basin after 1950 contrasted sharply with the frontal style of deglaciation of the valley below. Around 1950 the ice margin reached the open, northern end of the cirque (Figure 2, Plate 1), while exposure of the headwall separated ice within the cirque from its accumulation area. Once ice stagnated in the cirque, ice surface lowering accelerated. Between 1950 and 1985 approximately 35 m of ice has melted from above the present elevation of Ouzel Lake.

The style of deglaciation of the cirque basin is analogous to, but on a smaller scale than, a model based on the twentieth-century deglaciation of Glacier Bay, Alaska developed by Goldthwait and Mickelson (1982). The model is characterized by temporally arranged nunatak, channel, esker, and lacustrine phases of deglaciation. As the Redoubt Glacier retreated up the Depot Creek valley, ice entering the cirque thinned until the headwall of the cirque became exposed (nunatak phase, circa 1910-1930). Exposure of the headwall restricted the flow of ice into the cirque (Plate 2) and resulted in the stagnation of ice. During the late 1940s, as the ice margin reached the edge of the cirque, frontal deglaciation slowed and a series of meltwater channels were cut into thick accumulations of till (Figure 13) (channel phase). As areal deglaciation proceeded, a number of eskers and kames were deposited during the late 1950s (Figure 13) (esker phase) (Plate 2). Finally, the emergence of Ouzel Lake in the late 1950s and early 1960s led to the collapse of a large portion of the glacier over the lake (Plate 3) (lacustrine phase).

The timing of the collapse coincides with deposition of the esker and probably controlled deposition of two kames located along the shoreline of Ouzel Lake (Figure 13) by opening passages from the surface of the ice to its bed. Whether or not this event directly

Plate 1. 1947 Vertical Airphoto of Cirque Basin. Position of bedrock divide shown by arrow. (Photo by U.S.Forest Service)

Plate 2. 1956 Vertical Airphoto of Cirque Basin. (Photo by U.S.Forest Service)

Plate 3. 1968 Oblique Airphoto of the Cirque from the Northwest, lookinf Southeast. (Photo by Post)

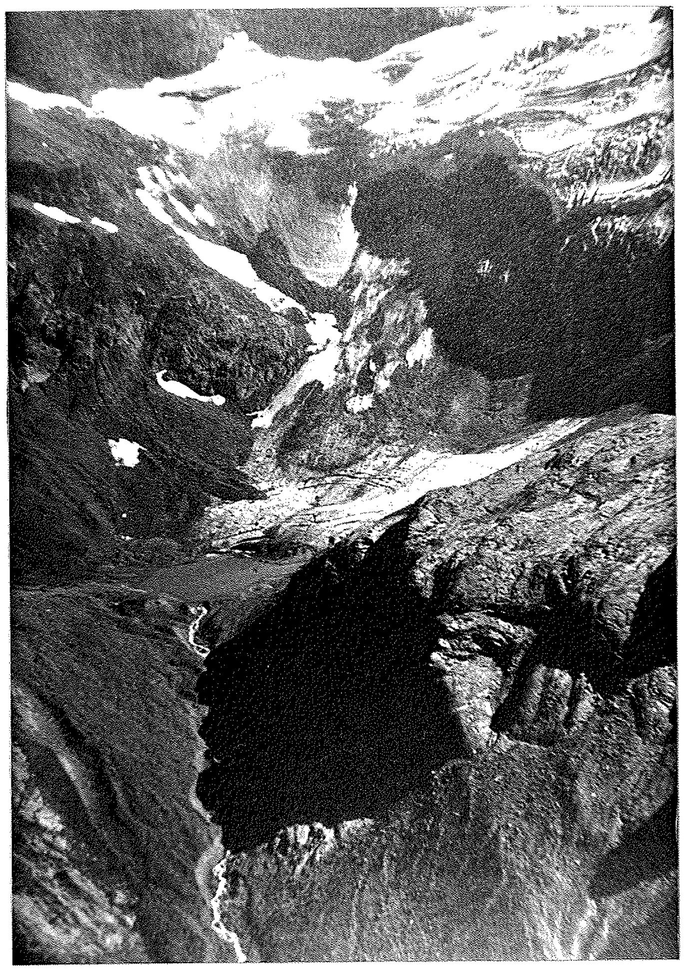
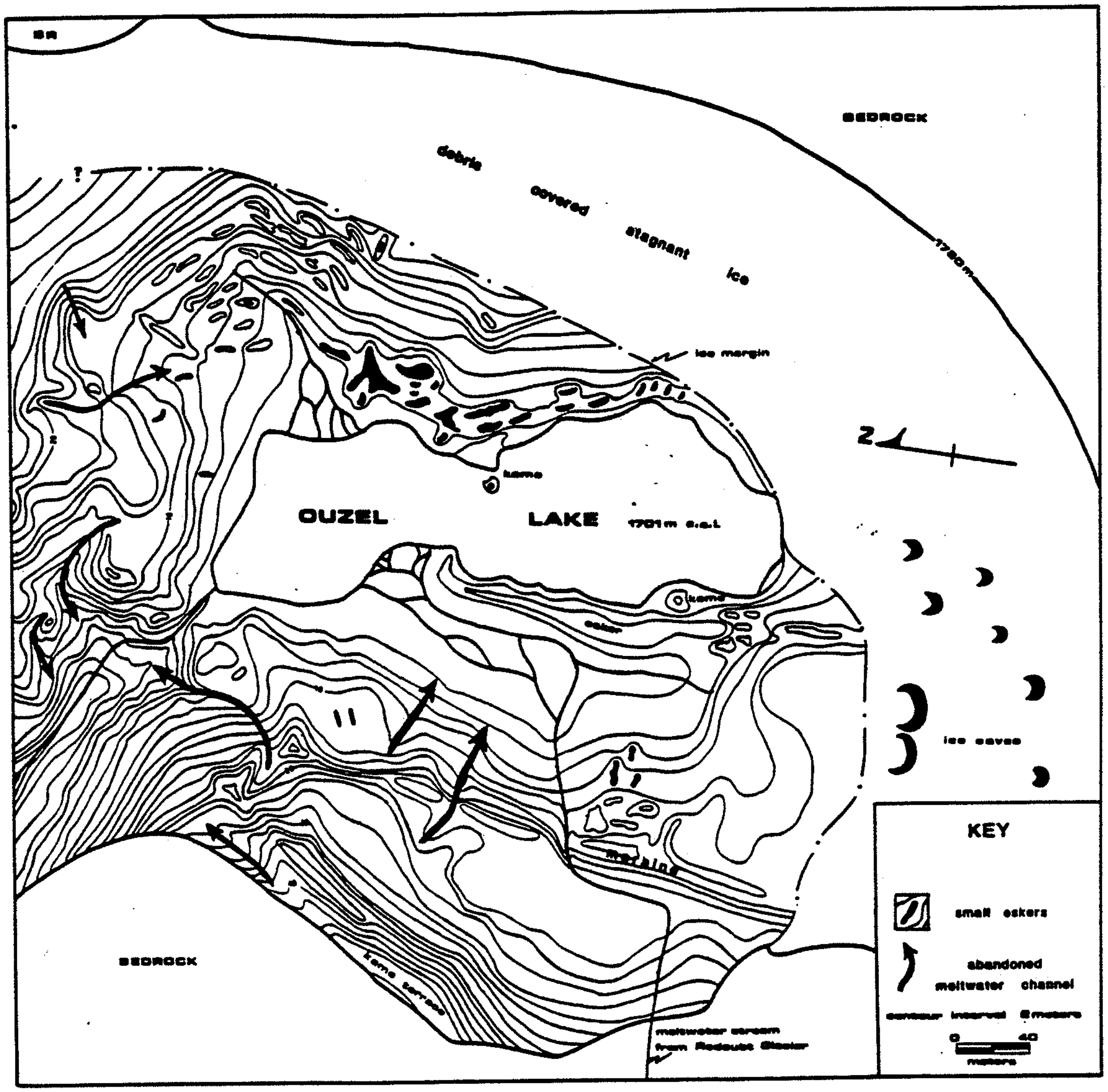


Figure 13. Map of the Redoubt Cirque made in 1985, showing the Position of the Fine-Grained Esker and other Features formed during Deglaciation.



influenced deposition of the esker is impossible to determine because the esker was deposited beneath the surface of the ice. It is possible, however, that the collapse of ice affected depositional processes, opened the tunnel, or influenced tunnel morphology.

## 2. Esker Genesis

The conditions that led to the stagnation of ice in the cirque are not uncommon in the North Cascades, although the landforms deposited during stagnation are. Similar situations where large, Pleistocene-age cirques lie below modern glaciers are found throughout the North Cascades. Such areas include the Sulphide, Luna, Azure and upper Thornton Lake areas. Differences between these areas and the Redoubt cirque were examined to identify factors that may have influenced deposition of the large esker and other fluvioglacial landforms. These factors include size, elevation, amount of modern glacial cover above the cirque, the manner of ice flow into the cirque, and the movement of meltwater during deglaciation.

In part because of its northerly aspect and high elevation (1700 m), the Redoubt cirque is much larger than most of the others. Smaller cirques in the North Cascades typically have floors covered by lakes and steep slopes ascending from shorelines. In contrast, a good portion of the Redoubt cirque floor is not occupied by a lake (Figure 13, Plate 4), leaving room for fluvioglacial landforms to be deposited and preserved.

Another feature of the Redoubt Glacier that distinguishes it from others in the area is its size. It covers 2.7 $km^2$ on the northeast side of Mount Redoubt. As one of the largest glaciers on a nonvolcanic peak in the North Cascades (Post _et al._, 1971), the Redoubt Glacier has delivered large quantities of ice and sediment to the cirque. Because of its morphology, the cirque has trapped sediment eroded by the Redoubt Glacier.

The nature of ice flow into the Redoubt cirque also appears to be uncommon. A bedrock divide (Plate 1) caused the Redoubt Glacier to

Plate 4. 1985 Photograph of the Fine-Grained Esker looking East from the West Wall of the Cirque.

flow into the cirque below in two separate lobes. During the most recent advance of the Redoubt Glacier, the two lobes flowed parallel to each other over the headwall of the cirque, creating a medial moraine between them. Continued activity of the two lobes above the cirque deposited a series of medial moraines on the surface of the ice below during the late 1940s and 1950s (Plates 1 and 2), which mark the position of the interlobate zone. The orientation and position of the interlobate zone in the cirque, as represented by the medial moraines, coincides with that of the large, fine-grained esker. Further, the location of the interlobate zone coincides with that of a series of ice caves that were observed in the last portion of the stagnant glacier against the cirque headwall (Figure 13).

The nature of ice flow into the cirque and its influence on the structure of the ice played a major role in the deposition of the fine-grained esker by controlling meltwater movement beneath the stagnant ice. Either the interlobate zone represented an area of structural weakness in the ice that was exploited by meltwater drainage, or tunnels existed before meltwater reached the interlobate zone. The latter possibility is more probable since tunnels not occupied by the stream are continuous to the headwall of the cirque (Figure 13). Previous studies have related both the location of meltwater channels and eskers to structural weaknesses in glacier ice such as crevasses (Stenborg, 1968; Shreve, 1972), shear planes (Stokes, 1958; Howarth, 1971), and interlobate zones (Stoelting, 1978; Punkari, 1980, 1982, 1985; Thome, 1986). Most of these studies, however, focused on events of a much larger scale.

Most of the meltwater that deposited the esker appears to have originated from the main body of the Redoubt Glacier (i.e., west lobe) (Figure 2, Plates 1 and 2). Around 1956 this stream flowed along the edge of the cirque, depositing a kame terrace along its western wall (Figure 13, plate 2). During subsequent deglaciation of the cirque, the stream migrated from north to south beneath the stagnant glacier, leaving meltwater channels as it flowed toward Ouzel Lake (Figure 13). Sometime during the late 1950s or early 1960s the stream entered the interlobate zone and turned abruptly to the north (Plate 4).

Meltwater entering the interlobate area eroded through a moraine and flowed across a till surface deposited by the west lobe of the Redoubt Glacier (Figure 13, Plate 4). Comparisons between pebble lithologies of the till surface and the esker indicate that the till surface was the primary source of sediment for the esker. A majority of the pebbles in both the till and the esker were composed of granodiorite from the Chilliwack Composite Batholith, which underlies the west lobe of the Redoubt Glacier.

Deposition of the 120 m long, 6 m high esker occurred shortly after meltwater entered the interlobate zone. Air photos taken in 1966 and 1967 by Post show that the distal portion of the esker was emerging from a completely sediment-filled tunnel. Important morphologic features of the esker include a predominantly single, sharp ridge crest, downstream narrowing of the esker from 36 m to less than 1 m, and four side ridges perpendicular to its main axis.

The side slopes of the esker are near the angle of repose of unconsolidated sand and range from 29° to 37°. The sharp ridge crest indicates that the tunnel had a low width/depth ratio (Price, 1973, p.41). Shreve (1985) suggested that tunnel shape is controlled by ice thickness and surface gradient and subglacial topography. Morphologic features of the esker, however, such as downstream narrowing and side ridges , indicate that tunnel shape was primarily controlled by structural features of the ice.

Most of the length of the esker has a single ridge crest, although the crest is split at the upstream end of the esker (Figure 14). The east ridge is composed primarily of poorly sorted sand and gravel, while the west ridge has well-sorted fine-grain components. Price (1966) observed that eskers deposited in englacial and supraglacial positions frequently had ice cores, which caused the splitting of single ridges into two as they melted. An alternative explanation for the split ridges observed in the fine-grained esker is that they represent separate stream channels. Stokes (1958) observed that tunnels frequently split, meander, and form oxbow loops.

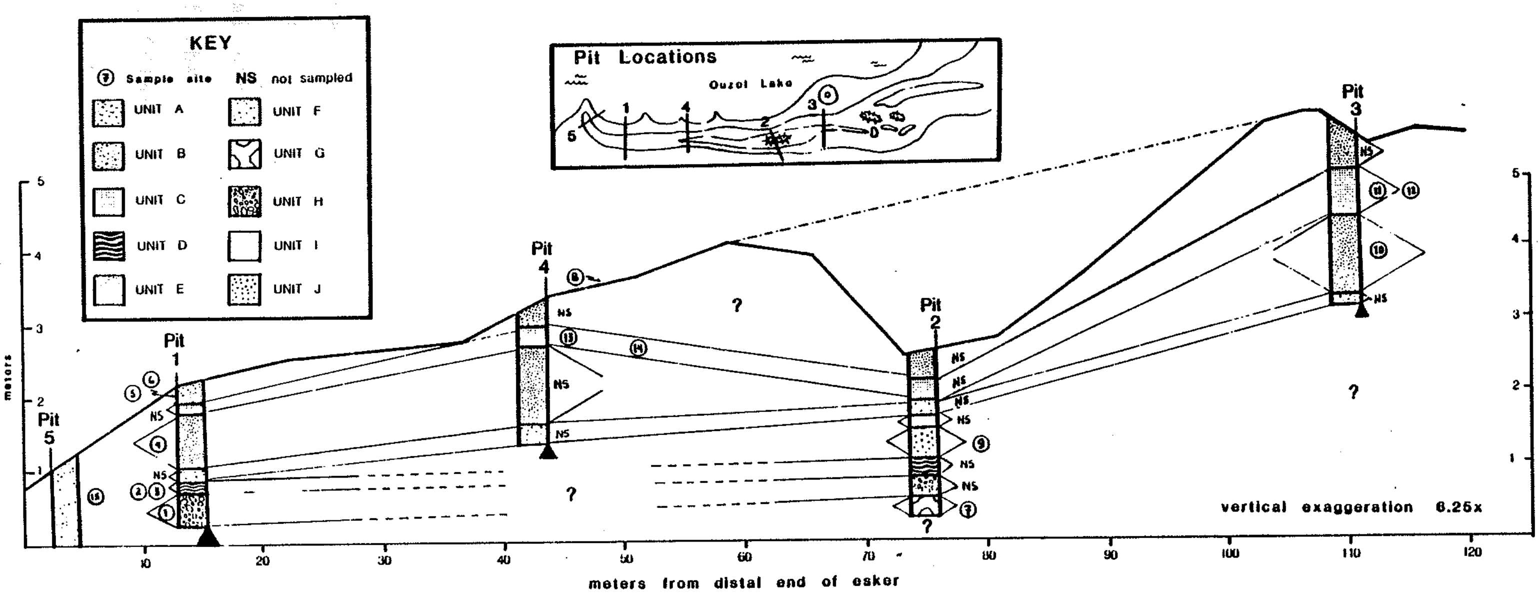



Figure 14. Longitudinal Profile, Stratigraphy, Sample Sites, and Correlation between Sedimentary Units in the Fine-Grained Esker. The units are described in Table 7 and sample data are in Table 8. The horizontal base-line represents the base of the esker, which is coincident with the stream gradient adjacent to the esker.

Differences between the sedimentology of the two parallel ridges indicate that a split channel, and not an ice core, created the split crest.

Structural control of the tunnel's shape may also explain the presence of four side ridges on the east side of the esker (Figure 13). All four ridges trend perpendicular to the main axis of the esker and extend into Ouzel Lake. The most distal ridge extends across the bottom of the lake and forms the downstream end of the esker. Side ridges are infrequently discussed features of eskers in previous studies. Lundqvist (1979) suggested they may represent crevasses intersecting subglacial tunnels. An alternative explanation is that side ridges represent junctions between tributary and primary meltwater streams. The former explanation, however, best explains the side ridges observed in this study because of the angle of their intersection with the esker and they are all composed of the same sediment that forms the crest of the esker.

Process-form relationships in the development of the esker are difficult to identify for two reasons. First, several different processes operating in the ice-contact environment deposited sediment and influenced the internal structure of the esker. Second, the form of the esker may have been controlled primarily by structural features of the ice before depositional processes were active. Nonetheless, a number of depositional processes were identified by examining the internal sedimentology and structure of the esker. They include deposition in lacustrine, fluvioglacial, and ice-contact settings.

Deposition of sediment directly from the glacier is evident in a number of sedimentary units. Unit A (Table 7) is supraglacial diamicton that is distributed unevenly on the surface of the esker. It was the final sedimentary unit deposited and becomes thicker at the upstream end of the esker, where the esker merges with an area of hummocky debris (Plate 5). Units B, D, E, F and G (Table 7) also have clasts in otherwise moderate to well sorted textures that were most likely deposited from the roof of the tunnel.

Another ice-contact process that influenced the form of the esker

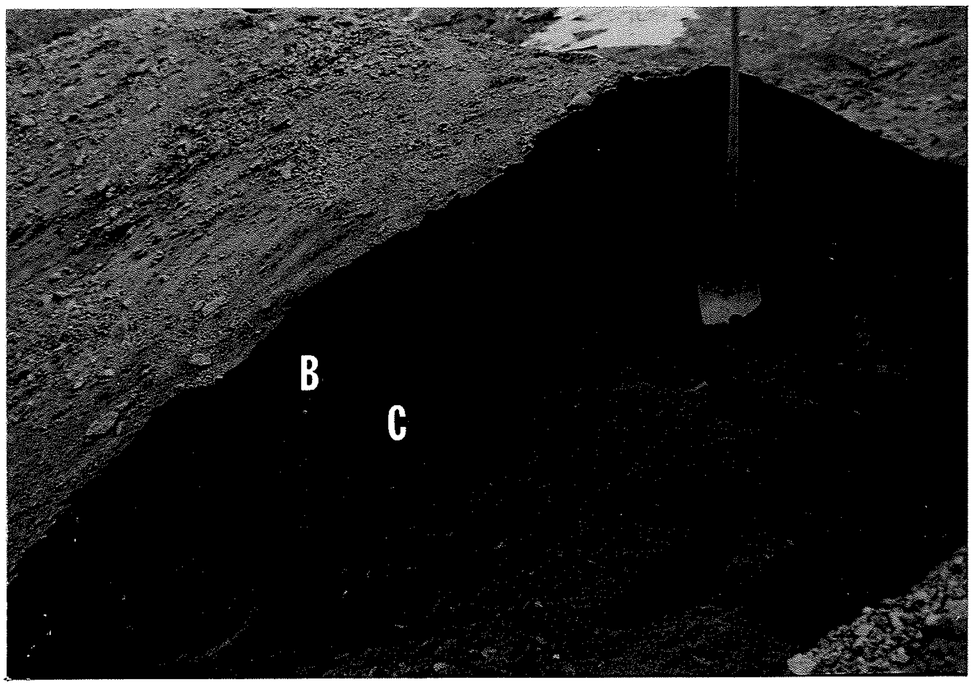
Table 7. Description, Distribution, and Interpretation of the Origin of Sedimentary Units in the Fine-Grained Esker.

| Unit | Description/Location | Origin |
|---|---|---|
| A | Very poorly sorted with subrounded to very angular clasts as large as 40 cm. Thickness 5 - 20 cm on surface of esker. Sample 14. | Supraglacial diamicton |
| B | Moderately to poorly sorted with nearly equal proportions of medium and coarse sand. Forms nose and side-ridges of esker. Thickness 5 cm - 2 m. Samples 5, 6, and 15. | Fluvioglacial |
| C | Very poorly sorted (Pit 4) to very well sorted fine and very fine sand. Faulted in pits 1 and 3. Thickness increases up-stream. Samples 8, 11, 12, and 13. | Lacustrine/ Fluvioglacial |
| D | Moderately sorted, ripple laminated fine sand. Faulted on west side of esker. Thickness approx. 15 cm. Samples 2 and 3. | Lacustrine/ Fluvioglacial |
| E | Moderately to poorly sorted coarse sand. Volumetrically most extensive facies. Thick-ness approx. 1m. Samples 4 and 10. | Fluvioglacial |

Table 7 continued.

| | | |
|---|---|---|
| F | Poorly sorted sand and gravel observed in all pits except 5. Thickness 10 - 20 cm. Sample 1. | Fluvioglacial |
| G | Boulders approx. 50cm B-axis in diameter supported in silt matrix. Observed only in pit 2. Thickness unknown. Sample 7 silt matrix. | Ice - contact (boulders)/ Lacustrine (silt) |
| H | Well sorted gravel observed at at base of pits 1 and 2. Not sampled. | Fluvioglacial |
| I | Faulted silt observed only in pit 2. Thickness approx. 10 cm. Not sampled. | Lacustrine |
| J | Muddy sand observed only in pit 2. Thickness approx. 50 cm. Sample 9 | Lacustrine/ Fluvioglacial |

Plate 5. Sedimentary Units B and C in Pit 1 on the East Side of the Esker.



was deformation of facies by ice wall melting. Two types of disturbance were noted in the esker. Unit C is offset along near-vertical normal faults near the center of the esker in pits 1 and 3 (Figures 15 and 16, Plates 5 and 6). Unit D in pit 2, on the other hand, shows no evidence of faulting, but curves abruptly toward the base of the esker in pits 1 and 2 (Plate 7, Figure 15). The lack of faulting in unit D indicates that it was frozen at the time of ice-wall meltout (Plates 7 and 8). Similarly, unit C in pit 3 (Figure 16) was also frozen at the time it was faulted, since a block of this unit is isolated in coarser material.

Deposition of some units in the esker occurred in a lacustrine environment. Units G, I, and J all have significant silt sized components in their grain size distributions (Tables 7 and 8, Figures 14 and 17). All of these units were found only in pit 2, which indicates that ponding of water in the tunnel was local. Ponding may have been controlled by changes in the morphology of the tunnel or channel. Ouzel Lake does not appear to have been a major factor controlling deposition although it is located adjacent to the esker (Figure 13) and was present when the esker was deposited.

Most of the esker was deposited in a fluvioglacial environment within the tunnel as indicated by the longitudinal persistence of sedimentary units and the linear esker ridge morphology (Banerjee and McDonald, 1975; Saunderson, 1975). It is possible that a linear ridge morphology could have formed in a deltaic setting by the coalescence of a series of deltas. It would be unlikely, however, for sedimentary units to persist through the length of such an esker ridge. Units B, C, E, and F are traceable through the length of the esker, while units D and H were observed in pits 1 and 2 (Figures 14, 15, and 17).

A lack of lateral variation in the sedimentology of the esker and a fairly steep longitudinal crest profile indicate that the stream occupied a single, stable channel, with the exception of the split crest near pit 2 (Visher, 1965; Shreve, 1972; Banerjee and McDonald, 1975). The presence of two fining upward sequences (Figure 14) and a disconformity between units C and E in pit 3 (Figure 16) indicate that

Figure 15. Cross-Section of the Esker at Pit 1, looking upstream and showing the field of view in Plates 5 and 7.

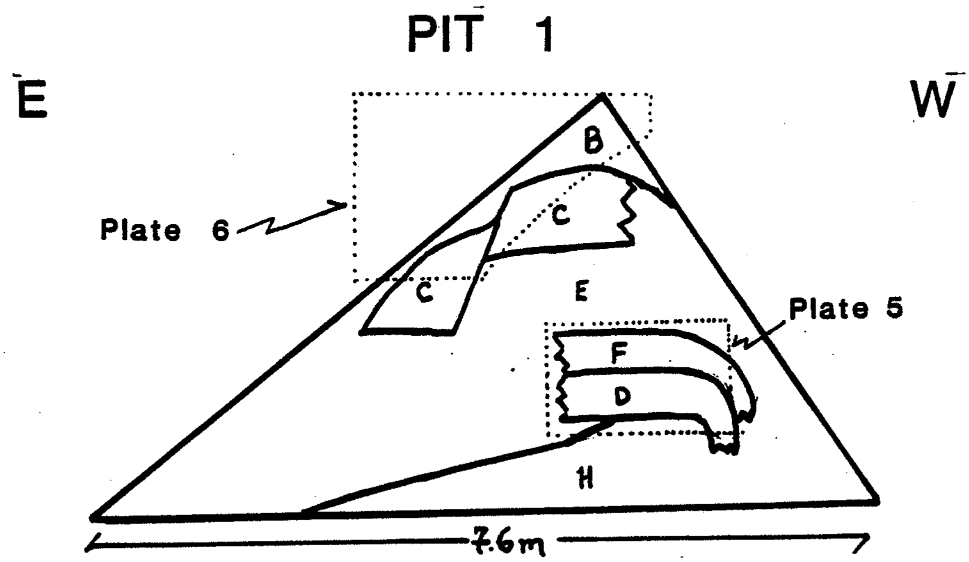


Figure 16. Cross-Section of the Esker at Pit 3, looking upstream.

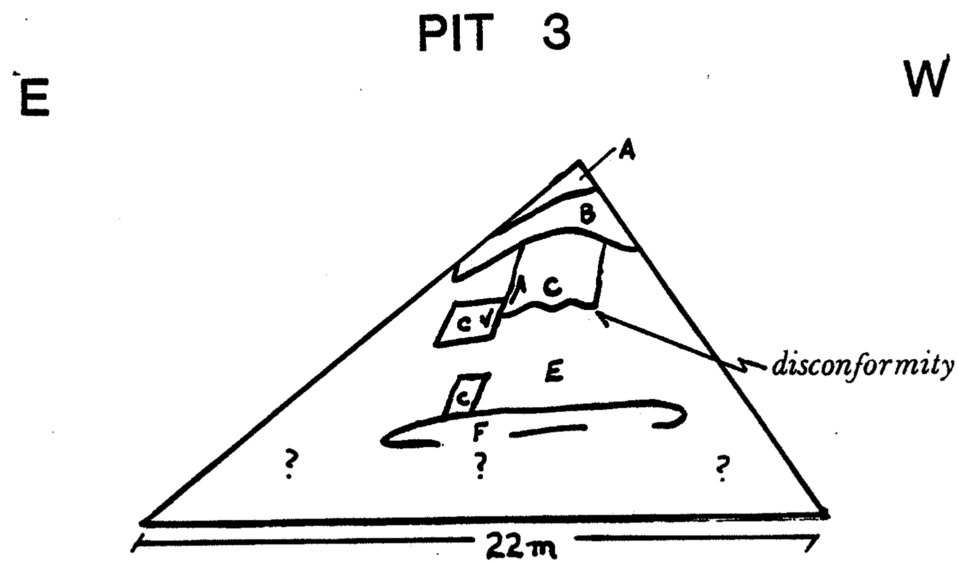


Plate 6. Sedimentary Units B, C, F, H, and I in Pit 2 on the East Side of the Esker.

Plate 7. Sedimentary Units D, F, and H in Pit 1 on the West Side of the Esker.

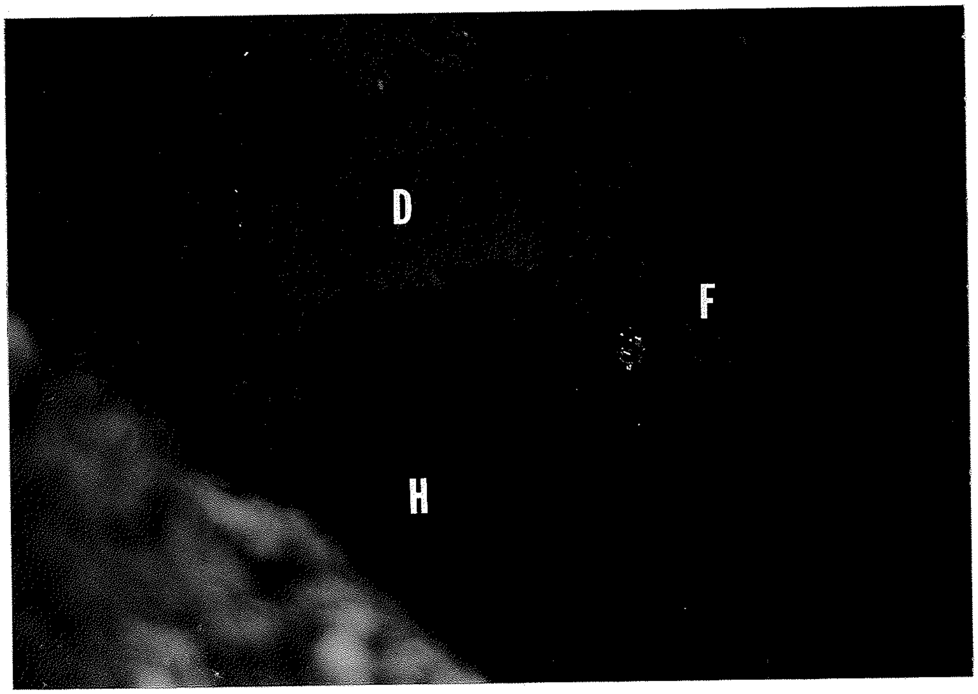


Plate 8. Sedimentary Units B, C, and E in Pit 4 at the Crest of the Esker.

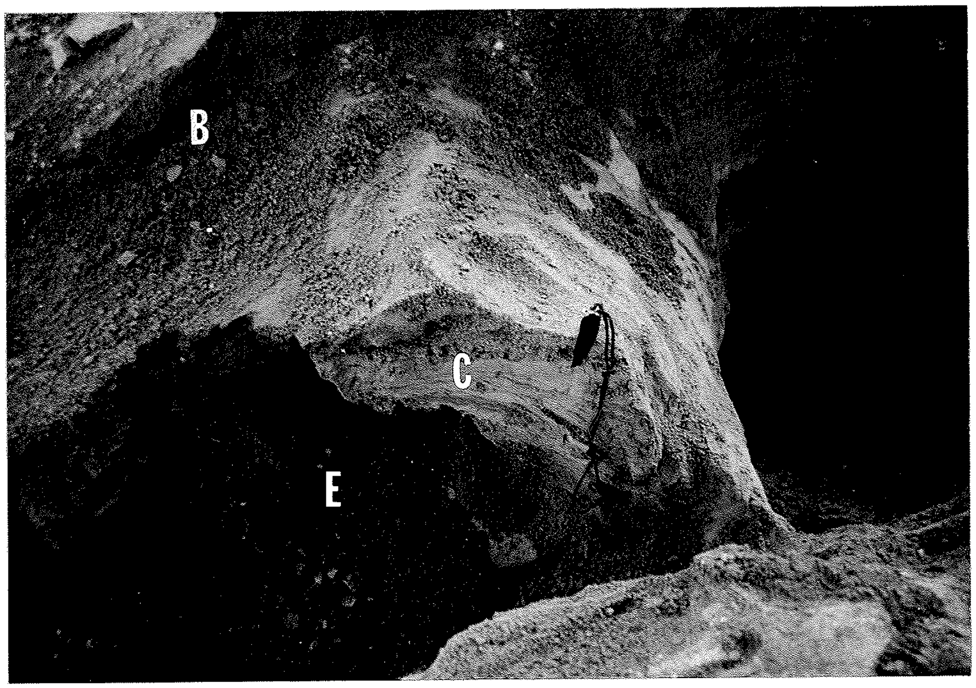


Table 8. Textural Data of Sediment Samples taken from the Fine-Grained Esker.

| SAMPLE NO. | PIT NO. | SORT. COEF. | MEAN GRAIN SIZE (mm) | <- - - - VCS | CS | PERCENT MS | FS | - - - VFS | -> SI | SKEWNESS |
|---|---|---|---|---|---|---|---|---|---|---|
| 1 | 1 | 1.15 | .49 | 18 | 30 | 31.0 | 13 | 3 | 0.8 | 0.20 |
| 2 | 1 | .84 | .18 | 1 | 2 | 30.0 | 43 | 15 | 4.0 | -0.31 |
| 3 | 1 | .89 | .13 | 0.8 | 1 | 12.0 | 44 | 29 | 5.0 | -0.50 |
| 4 | 1 | 1.24 | .78 | 44 | 28 | 15.0 | 6 | 2 | 0.9 | 0.57 |
| 5 | 1 | .7 | .37 | 4 | 21 | 57.0 | 13 | 3 | 1.0 | 0.02 |
| 6 | 1 | 1.4 | .41 | 19 | 28 | 28.0 | 14 | 6 | 3.0 | --- |
| 7 | 2 | 2.72 | .07 | 0 | 0 | 0.3 | 3 | 29 | 68.0 | -0.39 |
| 8 | 2-4 | -.2 | .14 | 0 | 0 | 1.0 | 14 | 49 | 27.0 | 0.47 |
| 9 | 2 | .10 | .13 | 0 | 0 | 0.6 | 29 | 49 | 12.0 | -0.52 |
| 10 | 3 | 1.06 | .27 | 3 | 15 | 41.0 | 25 | 12 | 3.0 | -0.02 |
| 11 | 3 | .66 | .11 | 0 | 0.3 | 0.8 | 38 | 48 | 3.0 | -0.73 |
| 12 | 3 | .69 | .20 | 0.9 | 3 | 27.0 | 52 | 13 | 2.0 | -0.39 |
| 13 | 4 | 3.48 | .01 | 0 | 0 | 0.0 | 1 | 14 | 84.0 | -0.94 |
| 14 | 4 | 3.61 | .07 | 9.7 | 10 | 12.0 | 13 | 14 | 41.0 | 0.07 |
| 15 | 5 | .97 | .50 | 16 | 34 | 36.0 | 8 | 1 | 0.4 | 0.20 |

Percentages :

VCS : very coarse sand

CS : coarse sand

MS : medium sand

FS : fine sand

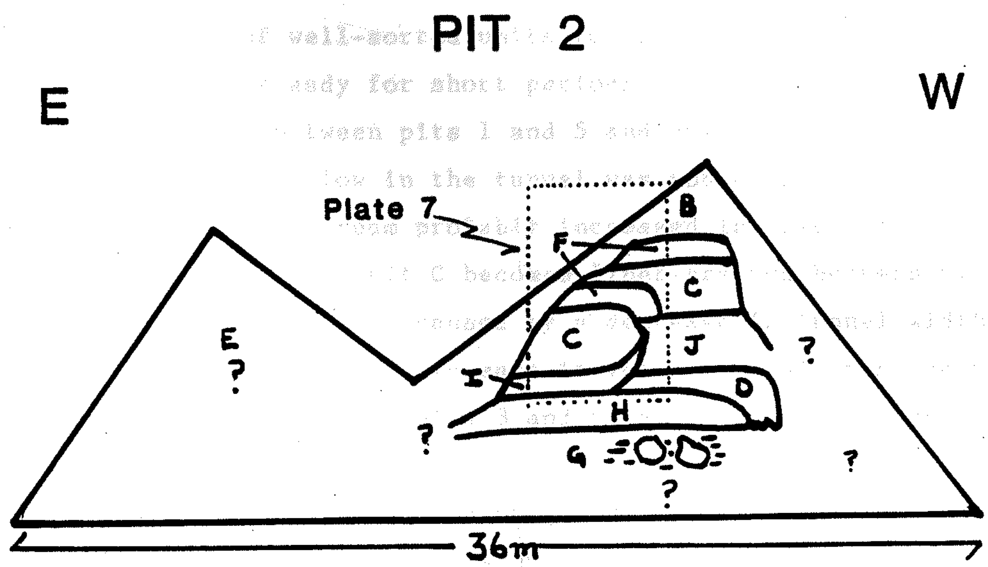
VFS : very fine sand

SI : silt

Note 1) Sample 8 was taken between pits 2 and 4 in unit C.

2) Mean grain size was determined graphically.

Figure 17. Cross-Section of the Esker at Pit 2, looking upstream and showing the field of view in Plate 6.



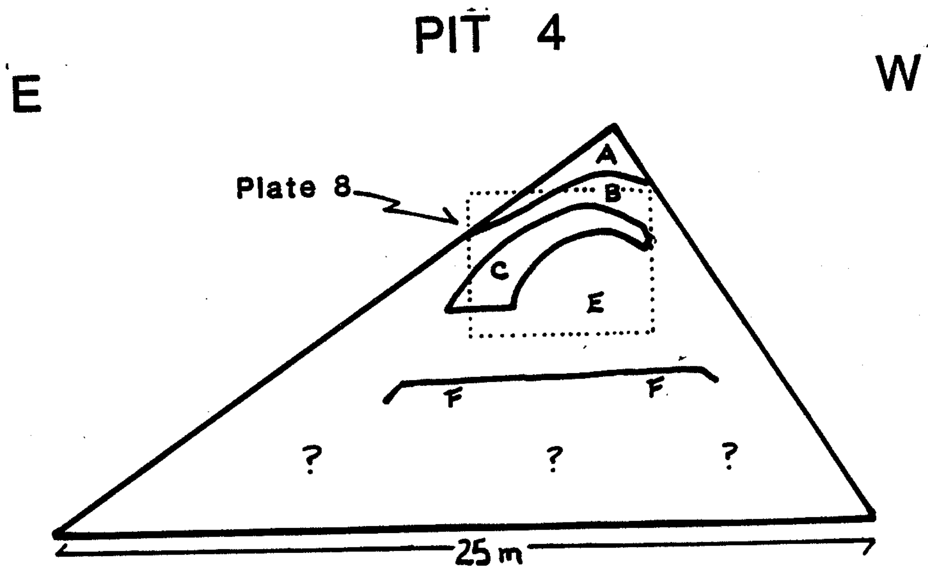
meltwater discharge through the tunnel was unsteady during deposition of most of the esker. The disconformity suggests that deposition was interrupted by a period of erosion after deposition of unit E. The two fining upward sequences probably represent separate episodes of deposition that occurred as discharge decreased, and may have separated by periods of erosion or non-deposition.

The presence of well-sorted units (C and D), however, indicates flow may have been steady for short periods. Downstream coarsening of sediments in unit B between pits 1 and 5 and unit E between pits 3 and 1 (Table 8) indicates flow in the tunnel was non-uniform. Velocity and competence of the stream probably increased in response to tunnel narrowing. In contrast, unit C becomes finer-grained between pits 3 and 4, which was most likely caused by a decrease in tunnel width from 25 to 22 m and an associated decrease in velocity. Alternatively, the grading of sediments between pits 3 and 4 may have been caused by local ponding of water near pit 4.

Units B, E, F, and H are coarse-grained and moderately to poorly sorted and form the core of the esker (Table 7, Figures 14-18, plates 5-8). The coarse texture and poor sorting of these units indicate they were deposited quickly (Saunderson, 1975) from both suspended and bed loads of the stream (Harms and Fahnestock, 1965). Minimum velocities at the time of deposition of units B, E, and F were calculated using the mean grain size of the units and entrainment velocity curves reported by Sundborg (1956). The values ranged from 0.45 m/sec for unit B, to 0.5 m/sec for unit F, and 0.55 m/sec for unit E.

Ripple laminations in unit D were used to reconstruct hydraulic parameters at the time of deposition of unit D. Mean grain size values from unit D (samples 2 and 3, Table 8) were used to estimate entrainment velocity at 0.35 m/sec (Sundborg, 1956). Jopling (1966) suggested that surface velocities are typically 2.25 times threshold velocities when suspended sediment concentrations are high, which is indicated by the presence of type D ripple laminations (Jopling and Walker, 1968). Surface velocity was then multiplied by 0.8 to

Figure 18. Cross-Section of the Esker at Pit 4, looking upstream and showing the field of view in Plate 8.



estimate mean velocity at 0.63 m/sec (Saunderson, 1980). A bedform stability diagram constructed by Simons and Richardson (1962) and the mean velocity were used to estimate depth at the time of deposition at 0.36 m. A Froude number value of 0.34 was calculated using the depth and mean velocity values, which indicates deposition of unit D occurred in the lower flow regime.

Unit C is also a fine-grained unit, but the absence of bedforms prevented estimation of hydraulic parameters at the time of its deposition. The massive structure and textural properties of unit C, however, indicate aggradation was steady and rapid (Banerjee and McDonald, 1975; Harms and Fahnestock, 1965).

Rapid deposition of the esker within the tunnel in a fluvial environment was most likely controlled by fluctuations in discharge. Whether or not fluctuations in discharge were controlled seasonally or by smaller-scale precipitation is impossible to determine. Meier et al. (1971) and Tangborn et al. (1977) showed that precipitation events in a basin of similar size as that covered by the Redoubt Glacier can cause peaks in discharge comparable to those of the nival flood period of a seasonal hydrograph. In addition to controlling deposition of the esker, fluctuations in discharge may also have influenced faulting of certain units by controlling tunnel widening.

## 3. Inferred sequence of deposition

After initial occupation of the tunnel in the late 1950s and early 1960s, deposition of units H, D, J and I (Figure 14) probably occurred during the late summer of the first melt season. Mean velocity at the time of deposition of unit D was near 0.63 m/sec in a stream approximately 0.36 m deep. Localized ponding of water occurred farther back in the tunnel and influenced deposition of the silt component of unit G. Tunnel widening during the early part of the next melt season, or in response to increases in discharge, caused the deformation of previously deposited units, which were probably frozen. During subsequent episodes deposition, units F, E, and C were deposited as runoff decreased. During deposition of unit E meltwater was diverted to the east through a side channel near pit 2. A second period of faulting was followed by deposition of unit B and formation of the distal end of the esker and side ridges. The distal end of the esker completely filled the narrow downstream tunnel and was not removed by erosion because the meltwater stream migrated to its present position west of the esker. As deglaciation of the cirque proceeded, supraglacial debris (unit A) was deposited on the surface of the esker as ice melted toward the cirque headwall.

## 4. Summary

The unusual existence of the esker in an alpine setting is related to the presence of a large modern glacier over an ancient Pleistocene cirque. Stagnation of ice in the cirque following rapid 20th century retreat of the Redoubt Glacier created conditions suitable for deposition of the esker. The pre-stagnation flow of ice into the cirque controlled the location and, possibly, the formation of the esker.

Several processes were responsible for deposition of the esker and controlled its form. Of primary importance was fluvioglacial deposition within a subglacial tunnel, which produced the linear esker ridge morphology. Localized deposition occurred in ponded water within the tunnel and directly from ice. Fluctuations in runoff controlled deposition of most sedimentary units in the esker. In addition to fluvioglacial deposition of a meltwater stream along its entire course, the form of the esker was controlled by the original shape of the tunnel, which was probably controlled by structural features of the ice created by the lobate flow of ice into the cirque.

## 5. Implications of this study

Questions have been raised concerning the applicability of data gathered in relatively small scale modern glacial environments to the greater problem of genesis of Pleistocene glacial landforms (Shaw, 1972; Banerjee and McDonald, 1975). Studies in modern glacial environments, however, have contributed to our understanding of process-form relationships in the genesis of eskers (Lewis, 1949; Stokes, 1958; Price, 1966; Mickelson, 1971; Howarth, 1971; Gustavson and Boothroyd, 1982).

The results of this study underscore the usefulness of studies in modern glacial environments. Evidence gathered in this study indicates that: (1) methods used to identify process-form relationships in Pleistocene eskers (e.g. Banerjee and McDonald, 1975; Saunderson, 1975) can also be used to identify similar relationships in the development of smaller eskers in modern glacial environments. This similarity may indicate genetic relationships between the eskers of different scales. (2) a variety of depositional processes interact in the ice-contact environment to produce eskers - reaffirming the belief that eskers are polygenetic; (3) structural features of an ice mass, such as interlobate zones can control the deposition and location of an esker; (4) tunnel shape is complex and can be controlled by structural features of glaciers; and, (5) split ridge crests can be produced by separation of channels.

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