THE FRASER GLACIATION IN THE CASCADE MOUNTAINS,
SOUTHWESTERN BRITISH COLUMBIA

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A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF
THE REQUIREMENTS FOR THE DEGREE OF
MASTER OF SCIENCE

in
THE FACULTY OF GRADUATE STUDIES
DEPARTMENT OF GEOGRAPHY

We accept this thesis as conforming
to the required standard

THE UNIVERSITY OF BRITISH COLUMBIA
April 1995
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Date April 24, 1995
The objective of this study is to reconstruct the history of glaciation from the start of Fraser (Late Wisconsinan) Glaciation to the end of deglaciation, for three areas in the Cascade Mountains. The Cascade Mountains are located between the Coast Mountains and the Interior Plateau in southwestern British Columbia. The Coast Mountains were glaciated by mountain glaciation followed by frontal retreat, whereas the Interior Plateau underwent ice sheet glaciation followed by downwasting and stagnation. The Cascades were supposed to have undergone a style of glaciation transitional between these two.

Terrain mapping on air photographs followed by field checking was used to locate surficial materials and landforms indicative of glaciation style and pattern. All three study areas were glaciated by mixed mountain and ice sheet glaciation. At the start of Fraser Glaciation, alpine and valley glaciers formed around higher summits as occurred in the Coast Mountains. At the glacial maximum the entire area was covered by the Cordilleran Ice Sheet. Deglaciation was largely by continuous downvalley retreat of active glaciers, contrasting with downwasting and stagnation in the Interior Plateau, and frontal retreat in the Coast Mountains. The scarcity of fresh moraines in the cirques suggests that, unlike in the Coast Mountains, most cirque glaciers were not active at the end of glaciation. Only the highest north facing cirques remained above the local snowline throughout
deglaciation and, as a result, glaciers in these valleys remained active and retreated up valley.
The pattern of glaciation in the Cascade Mountains was similar to that of other areas which underwent mixed mountain and ice sheet glaciation, such as the Presidential Range in New Hampshire, the Green Mountains in Vermont, mountain ranges in west central Maine and the Insular Mountains on Vancouver Island. However, deglaciation in all areas was complex and depended strongly on local conditions. For this reason local patterns cannot be predicted easily on the basis of glaciation style.
The value of an understanding of glaciation style to improve the accuracy of terrain mapping was also investigated. It was found that the model developed for the Cascade Mountains was of some use in predicting the presence of fine-textured material in valley bottoms and for the prediction of glaciofluvial material overlying till. However fine-textured sediments were not found in all valleys which were predicted to contain them. The model appears to be most useful as an indicator of where to concentrate field checking in order to locate fine-textured sediments.
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ACKNOWLEDGMENTS

Many thanks to my supervisor June Ryder for her support and guidance throughout this project. Her insights and ideas on many aspects of this work were invaluable. Michael Church reviewed this document and provided may useful suggestions for its improvement. Numerous fellow students and friends provided support and encouragement. Assistance in the field was provided by Carrie Brown, Liz Leboe and Brian Waddington. Funding was provided by British Columbia Ministry of Forests, Cattermole Timber and a National Research Council of Canada matching funds grant.
CHAPTER 1 INTRODUCTION

The Cascade Mountains, in Canada, occupy a small area between
the Coast Mountains and the Thompson Plateau (Figure 1.1). The
objective of this study is to determine the style and pattern
of glaciation for three study areas in the northern Cascade
Mountains during the Late Wisconsinan Fraser Glaciation. Areas
investigated for this thesis are centred around Mt. Stoyoma,
Anderson River and Mt. Outram in the northern, central, and
southern parts of the Canadian Cascade Mountains (Figure 1.1).
A further objective is to investigate whether or not an
understanding of the glacial history of the Cascade Mountains
can be used to improve the accuracy of terrain mapping in the
study areas.

The style of glaciation is determined by the type of ice
mass present, whether ice flow direction is controlled by
underlying topography or ice surface slope, and whether ice is
actively flowing or stationary. The pattern of glaciation is
the spatial distribution of ice and ice flow direction.

From descriptions of glaciation in several localities
(Sutherland and Gordon 1993; Clague 1989; Waitt and Thorson
1983; Fulton 1967; Flint 1971, 1951; Davis and Mathews 1944)
three general styles of glaciation can be defined: 1. mountain
 glaciation, where ice is local and topographically controlled;
2. ice sheet glaciation, where the ice source is often
regional and ice flow is controlled by ice surface gradient,
not local topography; and 3. mixed mountain and ice sheet
 glaciation, where both local and regional ice sources exist at
Figure 1.1: Location map for the Cascade Mountains, showing the location of the three study areas.
different times throughout the glacial cycle. Ice flow is controlled by topography during mountain glaciation stages and by ice surface gradient during the ice sheet stages. These styles with examples from British Columbia and elsewhere will be discussed further in Chapter 2.

Three styles of deglaciation are described from several localities in North America, (Clague and Evans 1994; Clague 1989, 1984; Koteff 1974; Fulton 1967). The first is frontal retreat, which occurs when a glacier remains active throughout deglaciation and the ice thins while the snout moves back in the direction from which it advanced (Figure 1.2). This is most common with mountain glaciation, but also occurs at the periphery of ice sheets. The second style is frontal

Figure 1.2: Sketch depicting frontal retreat. a). Maximum extent of glacier. b). Glacier thins as snout retreats, ice is active. c). Continued retreat of active ice and deposition of glaciofluvial outwash. d). Continued retreat of active ice.
retreat with stagnation of the glacier snout (Figure 1.3). This occurs with both mountain and ice sheet glaciation. Finally, the third style occurs when ice becomes cut off from its source; it eventually loses plasticity and ceases to flow and downwasting and stagnation occurs. (Figure 1.4).

![Figure 1.3: Sketch depicting frontal retreat with stagnation of the snout. a) Maximum extent of glacier. b) Glacier thins as snout retreats, ice is active. c) Snout stagnates as glacier retreats. Active ice is present behind the snout. d) Continued retreat; ice blocks become isolated from stagnant front. Glaciofluvial outwash and ice contact deposits formed.](image)

Each style of glaciation produces a characteristic suite of landforms and deposits which will be discussed further in chapter 3. In particular the style of deglaciation, has the largest influence on the type of surficial material deposited.

The most recent glaciation during which the Cordilleran Ice Sheet covered much of British Columbia was the Fraser Glaciation. It has been characterized as mountain glaciation.
Figure 1.4: Sketch depicting downwasting and stagnation. a) Maximum extent of glacier. b) Ice mass thins in place, surface gradient is still maintained and local flow is present. c) Continued thinning, surface gradient is lost and ice is stagnant. d) Final stages of retreat, ice separates into blocks of dead ice. Ice contact glaciofluvial and glaciolacustrine material deposited.

and frontal retreat in mountain ranges such as the Coast Mountains and ice sheet glaciation with downwasting and stagnation in the lower areas such as the Thompson Plateau (Clague 1989). Existing evidence (Saunders et al. 1987; Mathews 1968; Holland 1964) suggests that elements of both of these styles are present in the Cascade Mountains. Thus the existing models of glaciation in British Columbia may not fully describe glaciation in the Cascade Mountains.

Fraser Glaciation started roughly 29,000 years BP with the formation and expansion of alpine glaciers in mountainous areas throughout the Canadian Cordillera. These continued to grow, and coalesced to form ice caps. Continued growth led to
the formation of the Cordilleran Ice Sheet at the Fraser Maximum roughly 14,500 years BP. This covered all but the highest summits in the Coast, Columbia and Rocky Mountains, and all lowlying areas between mountain ranges (Figure 1.5) (Ryder et al. 1991; Flint 1971; Fulton 1971; Davis and Mathews 1944).

Deglaciation began shortly after the ice sheet reached its maximum extent with thinning of the ice sheet (Figure 1.6). The ice mass over the Interior separated from that over the mountains due to the emergence from beneath the ice of intervening ridges. After this, continued thinning of Coast and Columbia Mountain ice led to separation into valley glaciers and frontal retreat continued until glaciers receded to roughly their present day extent (Clague 1989). In the Interior, ice downwasted and upland areas became ice free before low lying areas. Continued thinning led to widespread stagnation (Ryder et al. 1991; Tipper 1976, 1971; Fulton 1967). By 11,500 years BP (Ryder et al. 1991) glaciers had retreated to roughly their present day extent. Further details of the Fraser Glaciation in southern British Columbia will be discussed in Chapter 2.

Snowline rises from southwest to northeast across the Coast Mountains (Evans 1989). At Brittain River, on the coast, contemporary glaciation level is 1900 m while at Yalakom River on the east side of the mountains it is 2700 m. This 800 m rise in glaciation level over 150 km is due to a steep decline in snowfall eastward. As a result of this gradient areas to
Figure 1.5: The four phases of glaciation (after Clague 1989; Ryder 1982; Tipper 1976; Davis and Mathews 1944). a) Growth of alpine glaciers. b) Valley glaciers begin to coalesce. c) Piedmont complexes form around the Coast and Columbia Mountains. d) An ice sheet is formed over the Interior Plateau.
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the east of the Coast Mountains, such as the Cascade Mountains and Thompson Plateau are much drier and less likely to support glaciation.

The landforms of the Coast Mountains are dominated by glacial erosion. Horns, aretes, and deep steep walled troughs are common. The highest summits were nunataks during Fraser Glaciation and as a result are extremely steep and sharp. This area contains the highest peaks in British Columbia, with many summits over 3000 m, the highest being 4016 m Mt. Waddington. As a result of the large amount of winter snowfall and cool summer temperatures there are currently thousands of glaciers in the Coast Mountains (Evans 1989) including extensive ice fields around Mt. Silverthrone, Mt. Waddington, and the head of the Lillooet River (Ryder 1981; Holland 1964).

The Thompson Plateau is dominated by a rolling upland surface dissected by deep broad valleys. Most of the Thompson Plateau lies between 1200 m and 1500 m with a few summits to 2250 m (Holland 1964). As a result of the dry climate and low elevations, snowline is currently above all summits. Even during Fraser Glaciation there were no significant alpine glaciers within the southern interior.

The Cascade Mountains have been divided into four ranges: Okanagan, Hozameen, Skagit, and an unnamed northern range (Figure 1.1). In British Columbia they are lower than the Coast Mountains; the highest peaks are between 1800 m and 2400 m. Throughout the Cascade Mountains there are scattered remnants of a Late Miocene erosion surface (Ryder 1981;
Mathews 1968), which has been dissected by cirques and troughs. The higher summits show the effects of alpine glaciation, while lower peaks are rounded and clearly were overridden by ice (Holland, 1964).

As the Canadian Cascades lie east of the Coast Mountains, they receive less precipitation, and as a result of the rise in snowline across the Coast Mountains, snowline is currently above most peaks in the northern Cascades. In British Columbia and northern Washington only a few small glaciers remain in the Skagit and Hozameen Ranges, which are largely south of the Coast Mountains. These ranges are located where valleys cut through the Cascade Mountains from the coast, allowing moist coastal air masses to penetrate further into the range (Porter 1977). In Washington glaciers are extensive only on the large volcanoes.

In the north and east the Cascades grade into the Thompson Plateau. In these areas summits are low and rounded. The highest summits are less than 2000 m and total relief is between 900 m and 1400 m. In the south and west peaks are sharper. Summits are up to 2400 m in the Skagit area and relief is 1700 m to 1800 m.

The glacial history of the Cascade Mountains has been studied mostly in Washington, near the periphery of the Cordilleran Ice Sheet (Heller 1980; Porter 1976; Waitt 1977, 1975; Crandell 1963; Mackin 1941). Daly (1912) made observations of glaciation along the 49th parallel as part of a bedrock mapping project. Matthews (1968) studied glaciation
of the Lightening Lakes area in southern British Columbia and Saunders et al. (1987) studied deglaciation in the Chilliwack Valley. No studies exist for the area north of the Skagit River except that of Ryder (1981) in the northernmost Cascade Mountains.

The Washington studies show that, as in the Coast Mountains, alpine glaciers formed at the start of the Fraser Glaciation, but, unlike in the Coast Mountains these retreated before the Fraser Maximum. Then, at the Fraser Maximum the Cordilleran Ice Sheet overtopped most summits as in the Interior Plateau (Ryder 1981; Waitt 1977, 1975). Alpine glaciers were generally not active during deglaciation (Waitt 1977, 1975; Mackin 1941) and mountain valleys were ice free while Cordilleran Ice remained in main valleys.

The Chilliwack valley glacier located further north was likely confluent with the Cordilleran Ice Sheet (Saunders et al. 1987), but it began to retreat up valley while active Cordilleran ice remained in the Fraser valley. As a result a lake was dammed in the lower Chilliwack valley and a large sandur prograded to the west as the valley glacier retreated upvalley. This study demonstrates that drainage in one valley in the Cascade Mountains was disrupted by Cordilleran Ice pushing up into the lower reaches of the valley. It is possible that drainage in other Cascade valleys was also dammed by Cordilleran Ice, resulting in the formation of lakes in valleys which were deglaciated by frontal retreat of Cascade Mountain alpine glaciers.
The Canadian Cascade Mountains thus provide an opportunity to study patterns of glaciation which may differ somewhat from the models of mountain glaciation followed by frontal retreat and ice sheet glaciation with downwasting and stagnation, which have been developed for the Coast Mountains and Interior Plateau respectively. The areas studied are closer to the centre of the Cordilleran Ice Sheet than the Washington sites and thus may have experienced different patterns of glaciation. This study will add more detail to the understanding of Fraser Glaciation in British Columbia.

The three study areas are spread over a distance of 80 km from near the northern limit of the Cascade Mountains to the Skagit River (Figure 1.1). Each area encompasses between 100 and 200 km². The study areas were selected to encompass some of the topographic variability present within the British Columbia Cascades. The northern and southern areas were known to contain lateral moraines of the Cordilleran Ice Sheet (Dr. J.M. Ryder, personal communication 1992) and thus one ice marginal position was already known in each area. The study boundaries were delimited to include the high summits, large troughs and plateau areas near the moraines. The central area, at the head of Anderson River, was selected to include several well developed horns and large troughs, which are rare in other parts of the Cascades. Details of each area will be discussed in chapters 4 to 6.

The style of glaciation was determined by study of erosional and depositional features created during the last
glaciation. Each style of glaciation creates a characteristic suite of deposits and landforms. Many of these can be identified on air photographs, so this study relied heavily on air photograph interpretation for initial identification of landforms and deposits. This was accomplished by terrain mapping using the British Columbia system (Howes and Kenk 1988). Field work allowed more detailed examination of each area, including the collection of stratigraphic information and the confirmation of data obtained from air photographs. Once the landforms and materials were mapped it was possible to determine the style of glaciation which best fit the existing evidence. This methodology is discussed further in Chapter 3.

The results from this study demonstrate that the British Columbia Cascades have undergone mixed mountain and ice sheet glaciation. Thus the patterns of glaciation in the study areas were compared to those of other areas that have evidence of both local glaciers and an overriding ice sheet, such as the Insular Mountains on Vancouver Island and the Green Mountains and Presidential Range in New England, to determine which factors control the pattern of glaciation in areas where mixed mountain and ice sheet glaciation occurs. Finally the terrain maps constructed for this study are reexamined in light of the Cascades model to determine if an understanding of the glacial history of the Cascade Mountains can improve the accuracy of air photo interpretation.
CHAPTER 2 ESTABLISHED STYLES OF GLACIATION

The three styles of glaciation, mountain, ice sheet and mixed, are described in this chapter, with examples from British Columbia, Washington, northeastern United States and Scandinavia.

2.1 Mountain Glaciation

Mountain glaciation occurs in areas which have only a local source of ice. These areas are generally the highest and wettest areas present, such as the Coast Mountains of British Columbia, the Scottish Highlands (Sutherland and Gordon 1993), or the Scandinavian mountains (Flint 1971).

Davis and Mathews (1944) identified four phases of glaciation, three of which occur during mountain glaciation. Phase one is the alpine phase: relief of the land surface greatly exceeds ice thickness, and ice flow is controlled by topography (Figure 2.1a). In phase two, the intense alpine stage, valley glaciers coalesce to form branching systems and ice thickens until relief only slightly exceeds ice thickness (Figure 2.1b). Ice flow is still controlled by topography. During the third phase, the mountain ice sheet stage, ice thickness slightly exceeds relief, a continuous ice sheet is formed, but flow is still largely influenced by underlying topography (Figure 2.1c). The fourth phase is reached when a continental glacier is formed (Figure 2.1d) and ice flow is no longer controlled by relief. The details of glaciation in the
Figure 2.1: Sketch depicting the four phases of glaciation in mountainous areas (after Davis and Mathews 1944). a. Alpine phase: alpine glaciers form around higher summits; relief greatly exceeds ice thickness. b. Intense alpine phase: valley glaciers coalesce to form branching systems; relief slightly exceeds ice thickness. c. Mountain ice sheet phase: ice thickness slightly exceeds relief and a continuous ice sheet is formed; flow is controlled by underlying topography. d. Continental ice sheet phase: ice thickness greatly exceeds relief and flow is controlled by ice surface gradient.
Coast Mountains are described in the following section as an example of mountain glaciation.

2.1.1 Coast Mountains

The Coast Mountains along with the Columbia and Rocky Mountains were the source areas for much of the southern part of the Cordilleran Ice Sheet during Fraser Glaciation. Glaciation started with the growth of alpine glaciers roughly 29,000 years ago. (Ryder et al. 1991; Clague 1981; Ryder 1981). These grew into large valley glaciers and ice caps, which were largely confined to the Coast Mountains until 20,000 to 25,000 years ago (Clague 1989). With continued growth the valley glaciers coalesced into piedmont glaciers along the flanks of the mountains, and expanded over the Thompson Plateau.

Continued expansion created the Cordilleran Ice Sheet over the Interior, and it eventually covered all but the highest peaks in the Coast, Columbia and Rocky Mountains. The ice surface was generally above 2300 m and locally over 2500 m (Ryder et al. 1991; Clague 1989). The maximum extent occurred between 14,000 and 14,500 years BP (Clague 1989).

Retreat began immediately after the Fraser Maximum at 14,500 years BP. The ice cap covering the mountains thinned and separated from the Cordilleran Ice Sheet, and further thinning lead to separation into individual valley glaciers. (Ryder et al. 1991; Tipper 1976, 1971). Deglaciation throughout the Coast Mountains was primarily by frontal
retreat. The absence of dead ice topography such as hummocky ablation moraine, indicates that ice was active throughout retreat. The lack of recessional moraines is evidence of continuous recession.

Several studies document the pattern of deglaciation on the coast (Clague 1984, 1985; Armstrong 1981), where sea levels strongly influenced deglaciation. There are fewer studies of Fraser deglaciation in more inland areas. Desloges and Gilbert (1991) and Gilbert and Desloges (1992) used acoustic methods to study the sediments in Harrison and Stave Lakes. The lower layers of sediment in Harrison Lake are consistent with a high energy ice-proximal environment, but upward the pattern of sedimentation quickly becomes uniform, consistent with a more distal ice source. The sediment record, combined with a lack of end moraines, is taken as evidence of continuous rapid retreat of the Lillooet Valley Glacier. Similar results were obtained from Stave Lake.

Frontal retreat continued until about 11,000 years ago, when lakes in the lower and mid portions of Kwoiek Creek valley, on the eastern side of the Coast Mountains, were ice free (Souch 1989).

There is abundant evidence throughout the Coast Mountains (eg. Ryder and Thompson 1986) that more recent glacial advances have generally been followed by frontal retreat. Large lateral and terminal moraines are common and active glaciers can be seen to be retreating up valleys. Some large valley glaciers, such as the Tiedemann, have dead ice at the
snout and are therefore receding by frontal retreat with stagnation of the snout. It is likely that similar patterns of retreat were followed at the end of Fraser Glaciation.

2.2 Ice Sheet Glaciation

Ice sheet glaciation occurs in lower areas adjacent to mountains, such as the Interior Plateau of British Columbia or the lowlands to the east of the Scandinavian Mountains (Flint 1971). Ice sheet glaciation corresponds to phase four of Davis and Mathews phases of glaciation, and is most common in areas of low to moderate relief (Davis and Mathews 1944). During this phase ice is much thicker than relief and flow is unaffected by topography. Glaciation on the Thompson Plateau during Fraser Glaciation and the Scandinavian Ice Sheet during Weichsel Glaciation are described in the following sections as examples of ice sheet glaciation.

2.2.1 Thompson Plateau

Virtually all the ice covering the Thompson Plateau originated in the Coast and Columbia Mountains (Ryder et al. 1991). The area was overridden by the Cordilleran Ice Sheet after about 21,000 years BP. Expansion continued until 14,000 to 14,500 year BP (Ryder et al. 1991).

At its southern margin in north-central and northeastern Washington the Cordilleran ice sheet began to retreat its margins, the ice sheet receded by frontal retreat (Clague 1981), but elsewhere downwasting and stagnation dominated. By
11,500 to 10,000 years ago the plateaus and valleys of the interior system were completely deglaciated (Clague 1981).

Fulton (1967) identified four phases of deglaciation in the Kamloops area which summarize the style of ice sheet deglaciation for areas of moderate relief. The first phase was ice sheet glaciation. Ice was active, thicker than relief, and the regional ice surface gradient controlled flow (Figure 2.2a). During this stage a blanket of basal till was deposited. Once Interior and Coast Mountain ice separated and upland areas began to be exposed, phase 2 or the Transitional Upland Phase, was reached (Figure 2.2b). There was still minor regional flow, which was now controlled by local topography. Phase 3 or the Stagnant Ice Phase began when continued thinning allowed the exposed uplands to isolate ice in valleys (Figure 2.2c). Ice was still thick enough to behave plastically, local surface drainage was still present and ice marginal drainage was developed. The final, Dead Ice Phase occurred when ice was too thin to flow any longer and the surface gradient was lost (Figure 2.2d). Englacial and subglacial drainage were common resulting in the deposition of hummocky gravel and eskers. As ice downwasted major valleys were dammed by ice, drainage was disrupted and numerous glacial lakes formed (Ryder and Clague 1989; Fulton 1969, 1967; Armstrong and Tipper 1948; Mathews 1944).

Tipper (1971) found similar patterns of ice retreat in central British Columbia.
Figure 2.2: Sketch depicting the four phases of ice sheet deglaciation in an area of moderate relief (after Fulton 1967). a). Ice sheet glaciation: ice is much thicker than relief and flow is controlled by ice surface gradient. b). Transitional upland phase: uplands become ice free; flow is controlled by local topography. c). Stagnant ice phase: valley ice is cut off from its source, but still behaves plastically. d). Dead ice phase: ice is too thin to flow.
2.2.2 The Scandinavian Ice Sheet

The Scandinavian Ice Sheet originated in the Scandinavian Mountains at the start of Weichsel Glaciation with the growth of alpine glaciers. Glaciers flowing eastward from the mountains expanded and coalesced into piedmont glaciers on the Bothnian lowlands (Flint 1971). Further expansion and thickening lead to the formation of an ice sheet which buried the mountains and extended nearly 1300 km to the southeast, where flow was unimpeded. The maximum extent corresponds to continental glaciation or phase 4 of Davis and Mathews.

During deglaciation the ice sheet thinned rapidly and retreated back toward the mountains, leaving a series of end moraines (Eronen and Vesajoki 1988; Flint 1971). The thinning ice caused upland areas in northeastern Finland and several northeast draining valleys to become ice free while thick ice remained in valley bottoms (Johansson 1988; Flint 1971). As a result numerous large lakes were dammed in valleys by ice in their lower reaches, as occurred in the Kamloops area of British Columbia. Highlands in southern Sweden are very flat and ice here downwasted and stagnated, leaving dead ice topography of hummocky ablation till, kames, and kettles (Hillefors 1979).

2.2.3 Summary of Ice Sheet Deglaciation

Ice sheet deglaciation is largely influenced by local topography and proximity to the edge of the ice sheet. Near the margins, frontal retreat is common. Elsewhere, ice
downwastes. Areas with moderate relief, such as the Kamloops area and northeastern Finland, allow the ice sheet to become separated into isolated masses in valleys. This results in downwasting and stagnation with damming of valleys by ice in their lower reaches. Low relief areas in the interior of the ice sheet are likely to experience widespread stagnation and hummocky ablation till is likely to be common.

2.3 Combined Mountain and Ice Sheet Glaciation

Areas with mountains high enough to support alpine glaciers during glacial cycles, and which are located in a position which allows them to be overtopped by an advancing ice sheet, may experience both alpine and ice sheet glaciation. These areas do not develop major accumulation zones with radially outward flow because they do not receive enough snow or do not have sufficient area above snowline. Existing studies in the British Columbia Cascades (Saunders et al. 1987; Mathews 1968) suggest that this style may have occurred here. Thus it is useful to examine several examples of this style of glaciation. The last glaciation in the Washington Cascade Mountains, the Insular Mountains of Vancouver Island, and several areas in the northeastern United States will be described as examples of this style of glaciation.

2.3.1 Washington Cascade Mountains

During Fraser Glaciation the glaciation threshold dropped below the level of most summits in Washington (Porter 1977).
It was lowest where topographic lows, such as the Skagit River valley permitted moist maritime air to penetrate well into the mountains. Thus larger and lower glaciers could form near these valleys.

Studies in the Skagit Valley and Washington Pass area in the central Cascades by Waitt (1975, 1977), the Snowqulamie area by Mackin (1941) and Lake Tapps area by Crandell (1963) (Figure 2.3) show that alpine glaciers advanced down valleys during early Fraser Glaciation, then receded before the Cordilleran Ice Sheet extended into the region, (Waitt

Figure 2.3: Location map of the southern British Columbia and northern Washington Cascades showing the location of studies mentioned in the text.
1977, 1975; Crandell 1963; Mackin 1941) likely prior to 18,000 years BP (Armstrong et al. 1965).

The Cordilleran Ice Sheet flowed southward into Washington through divides not covered by Fraser Glaciation alpine ice (Waitt 1977). At the glacial maximum the ice surface reached an altitude of greater than 2600 m at the International Boundary, higher than most summits in the area (Waitt and Thorson 1983). Very little erosion was accomplished by the ice sheet so Waitt (1977) assumed it overtopped summits only briefly. At its southern limit the Cordilleran Ice Sheet was largely restricted to the Puget Lowland. It flowed into valleys building end moraines and impounding lakes behind ice dams (Crandell 1963; Mackin 1941), but did not overtop summits.

The timing of ice retreat in the Cascade Mountains is not known in detail. However, Mathews (1968) found evidence that the Lightning Creek valley, in southern British Columbia, carried meltwater from the interior toward the coast, implying that the Cascade valleys were ice free before the Interior. This implies deglaciation before 11,500 to 10,000 years BP (Clague 1981). The Chilliwack Glacier had receded by 11,900 years BP (Saunders et al. 1987). Kwoiek Creek in the Coast Mountains was ice free before 11,115 years BP (Souch 1989). Because mountains in the headwaters of Kwoiek Creek are higher and closer to ice accumulation zones than nearby parts of the Cascade Mountains it is probable that the Cascades became ice free first. Because the Washington Cascades are closer to the
periphery of the ice sheet it is probable that they were ice free before areas further to the north.

The Puget Lobe of the Cordilleran Ice Sheet retreated rapidly but irregularly by downwasting and frontal retreat, several stillstands are recorded by terminal moraines (Waitt and Thorson 1983). Marginal lakes formed in front of the ice sheet during retreat, leaving silt deposits and numerous drainage channels.

Alpine glaciers do not appear to have advanced after the retreat of the Cordilleran Ice Sheet as unweathered erratics are widely distributed from valley floors to ridgecrests, including on cirque floors, and no Cascades derived drift overlies Puget lobe drift (Waitt and Thorenson 1983; Waitt 1975).

2.3.2 Insular Mountains of Vancouver Island

At the start of Fraser Glaciation, glaciers formed around the high mountains of north central Vancouver Island and flowed down existing valleys (Howes 1981). Alpine glaciation was well established by 25,000 years BP. Lower peaks to the south remained ice free until overridden by the Cordilleran Ice Sheet (Howes 1983; Alley and Chatwin 1979). During the Fraser maximum Coast Mountain ice coalesced with and overrode Vancouver Island ice, reaching an altitude of at least 1500 m (Howes 1981).

During deglaciation ice downwasted, uplands emerged and the ice sheet separated into discrete valley glaciers. Down
valley flow was maintained while there was sufficient thickness of ice (Howes 1983, 1981). Eventually ice thinned and stagnated depositing thick sequences of ice contact and recessional gravels. Howes (1981) found no evidence that cirque glaciers were active at the end of deglaciation on northern or central Vancouver Island, however it is likely that small ice masses remained in cirques for some time. Alley and Chatwin (1979) found evidence, in the form of radial flow away from the highest summits, for a resurgence of alpine glaciation in some upland areas on southern Vancouver Island.

2.3.3 The Presidential Range, New Hampshire

The Presidential Range in New Hampshire (Figure 2.4) is the highest part of the White Mountains, with peaks to 1900 m and cirque floors between 1200 and 1350 m. Goldthwait (1970) has found evidence of local alpine glaciers which formed at the start of Late Wisconsinan Glaciation. At the peak of Late Wisconsinan Glaciation all summits were overtopped by the Laurentide Ice Sheet which advanced into the region from the north. This has resulted in considerable smoothing and rounding of alpine forms.

During deglaciation the ice sheet downwasted, and ice in valleys was cut of from its source. This resulted in stagnation in many of the larger valleys. Hummocky dead ice deposits are common (Goldthwait 1970).

There is continuing debate about whether or not alpine glaciers were active in the Presidential Range at the end of
Figure 2.4: Location map of northeastern United States showing the location of studies mentioned in the text.

deglaciation (Gerath and Fowler 1982; Bradley 1981; Goldthwait 1970). No terminal moraines of alpine glaciers have been found but fresh striations exist in cirques. Goldthwait (1970) asserted that the striations were formed by post glacial processes such as avalanching and that the alpine glaciers were not present at the end of the last glaciation. Bradley (1981) postulated that alpine glaciers were active during deglaciation to form the striations and that they coalesced with stagnant valley ice, thus eliminating all evidence of termini. Gerath and Fowler (1982) concluded that ice in the Presidential Range downwasted rapidly when the snowline rose.
above the elevation of the summits, and thus no alpine glaciers were present during deglaciation.

2.3.4 Green Mountains, Vermont

The Green Mountains of Vermont rise to an elevation of 1350 m. They were overtopped by ice flowing from the north during the Wisconsinan maximum (Flint 1951). During deglaciation, ice in lowlands downwasted and retreated northward, toward the centre of the Laurentide Ice Sheet, damming a lake behind the ice. Large deltas occur at the mouths of valleys draining from the Green Mountains implying that sediment was supplied by active ice in the mountains (Connally 1982). In addition, end and lateral moraines and outwash plains that exist in these valleys are thought to have been formed by active cirque glaciers during Laurentide Ice Sheet recession (Wagner 1970).

2.3.5 West Central Maine

The highest peak in west Central Maine is Mt. Katahdin at 1600 m. Several cirques are present in this area between 900 and 1050 m. The cirques are thought to have developed before the last ice sheet (Borns and Calkin 1977). All cirques have been modified by overriding ice. During deglaciation the ice sheet thinned and stagnated throughout the mountains without any reorganization of flow which would indicate the presence of active cirque glaciers.
2.3.6 Summary of Combined Mountain and Ice Sheet Glaciation

In all the above areas cirque glaciers formed at the start of the glacial cycle when climate deterioration caused local snowline to drop below summit elevations. No ranges investigated were high enough to generate extensive ice caps but were overtopped by regional ice at the glacial maximum. The regional ice sheet either coalesced with local glaciers as on Vancouver Island (Howes 1983) or local glaciers retreated before the advance of the regional ice sheet as in Washington (Waitt 1975, 1977). Whether or not local glaciers retreated before the advance of an ice sheet likely depended on the nearness of the area to the ice margin. In areas such as Washington, which are near the ice margin, climate amelioration could have begun to effect local glaciers before it affected the ice sheet due to the differing response times of the smaller and larger ice masses. Thus alpine glaciers could recede before the ice sheet attained its maximum extent.

Deglaciation was controlled by local topography, elevation of snowline and distance from the edge of the ice sheet. As climate warming began snowline rose, the ice sheet thinned and upland areas became ice free. In all areas the ice sheet retreated down-valley out of the mountain ranges. Stagnation of the ice sheet in large valleys was common but not universal. If cirques were below the local snowline at the time of deglaciation, no active glaciers were present, for example the Green Mountains (Wagner 1970; Connally 1982). In this case, ice in local valleys may have been cut off from
source areas, leading to downwasting and possibly stagnation. If cirques were above local snowline, glaciers occupied cirques during deglaciation, as occurred in west-central Maine (Borns and Calkin 1977), and some valleys were deglaciated by frontal retreat of active valley glaciers.
CHAPTER 3 METHODS OF DETERMINING STYLES AND PATTERNS OF GLACIATION

In order to determine which of the glaciation styles described in Chapter 2 occurred in a glaciated area, it is necessary to identify diagnostic criteria. Thus the landforms and sediments that are formed by glaciation and their implications for the state of the ice and ice marginal positions are described in this chapter. Techniques of recognizing glaciation style from suites of landforms and deposits are also described. Then the details of the techniques used in this research project, particularly the field program, are discussed.

3.1 Landforms and Sediments Indicative of Glaciation Styles

Landforms and surficial materials provide evidence of the processes that have created the present landscape. In many parts of British Columbia there has been only slight modification of the landscape since the last glaciation. Thus most landforms and surficial deposits relate directly to the processes that occurred during Fraser Glaciation and can be used to infer styles of glaciation.

3.1.1 Erosional Landforms

Erosional landforms include all features carved by ice or meltwater, such as cirques, aretes, troughs and meltwater channels.

Erosional forms may have been created during one or many glacial periods. As erosional forms can not generally be
dated, the easiest way of estimating the length of time required for erosional features to form is by comparing the volume eroded with established rates of erosion. Andrews (1975 p 113) estimated rates of glacial erosion by two methods. The first yields data over time scales on the order of 1 million years by dividing the amount of lowering required to create cirque basins by the total age of the basin. This yielded a rate of roughly 400 mm/1000 years in Scotland, and 50 mm/1000 years for a polar glacier. The second technique is valid for time scales on the order of 10 years and calculates the volume of material transported by present streams discharging from the snout of glaciers. This yielded a higher rate of erosion of between 1000 and 5000 mm/1000 years for rivers in Baffin Island, Norway, Iceland and the Karakoram. These rates are very approximate and involve several assumptions. There is no way of determining for how long out of the total of all glacial periods a particular feature was being eroded. For example a cirque may be carved mainly during the alpine phase of a glacial cycle. The short time scales associated with the measurement of stream load means that these rates may not be representative of longer time scales. The measurements over long time scales also include non glacial periods, so the two rates are consistent. In addition stream measurements made in front of valley glaciers that could be eroding soft sediments, may give erosion rates that are not representative of the erosion of bedrock. These rates should therefore be used only as a rough guide to how much time was required to accomplish
the amount of erosion observed. They may however, be useful for indicating the probability that observed erosion was accomplished only by the last glaciation.

3.1.1.1 Cirques, Aretes and Horns

Cirques, aretes and horns are features of alpine glaciation which formed during the alpine, intense alpine and mountain ice sheet phases of Davis and Mathews. Cirques indicate the general accumulation zones for valley glaciers and mountain ice sheets. Past regional snowline can be approximated from the elevation of cirque floors (Flint 1971). Horns and aretes are usually sharp features which remain above alpine glaciers. Cirques will commonly be reoccupied by alpine glaciers at the start of each glacial cycle, thus alpine features are generally the product of several glaciations.

Where ice later flows overtop of alpine features, they become rounded, smoothed and striated. Thus it can be assumed that rounded alpine features were formed by mountain glaciation prior to overriding by an ice sheet. If alpine glacial landforms are rounded but still prominent, then the duration of the ice sheet was likely short (Waitt 1975; Flint 1971). The point of transition from rounded to sharp ridge crests and summits indicates the approximate elevation of the surface of the ice sheet.
3.1.1.2 Troughs

Glacial troughs were carved by active ice flowing in channels any time from early advance until retreat or stagnation. Three types of troughs have been recognized; each is indicative of a different style of glaciation (Sugden and John 1976).

Alpine troughs were carved by valley glaciers during the alpine phase of Davis and Mathews. They head in cirques and their entire length is overlooked by higher ground (Sugden and John 1976). Thus when these troughs are carved, adjacent peaks stand above the general ice level. In some areas troughs change downstream to V shaped valleys characteristic of fluvial erosion. This transition marks either the terminal position of a valley glacier or the location at which the valley glacier merged with an ice sheet (Ryder et al. 1991). The length of alpine glaciers should be proportional to the area of their cirques (Flint 1971; Goldthwait 1970). Thus if the U - V transition represents the terminal position of the glaciers this relationship should be apparent from regional study of glacial morphometry.

Icelandic style troughs head in broad cols rather than cirques. There is no well defined area of accumulation within the trough. They are thought to have been cut by ice spilling over the trough head, so formed when ice was thick enough to flow across low points on divides (Sugden and John 1976; Embleton and King 1975). These form during the intense alpine and mountain ice sheet phases of Davis and Mathews.
The final type of trough is the through trough, which is open at both ends, often forming low passes through mountain ranges. Like the Icelandic troughs there is no local accumulation area. These troughs are thought to have been eroded under ice sheets, when ice streams into and exploits pre-existing valleys which are then eroded headward and widened and deepened. (Clapperton and Sugden 1977; Sugden and John 1976). They likely form during Davis and Mathews' intense alpine and mountain ice sheet phases.

Troughs will generally be reoccupied by valley glaciers during each glaciation. It is thus difficult to determine how many glacial cycles these features represent. Repeated glaciations of different magnitude can likely result in a blurring of the U-V transition of alpine troughs. Later glaciations which are smaller than earlier ones will leave few erosional remnants as the existing troughs will not be filled.

### 3.1.1.3 Meltwater Channels

Meltwater channels are steep-sided, flat floored channels carved by water from melting ice. Several types have been described by Derbyshire (1962). Three of these have been recognized in the study areas: subglacial channels which formed under ice, lateral meltwater channels which formed at the ice margin, and direct overflow channels, which flowed away from the ice.

Subglacial channels formed under warm based glaciers. Thus they may have formed at any time throughout the glacial
period, although most likely during deglaciation when abundant meltwater was generated. They have possibly been carved by water under high hydrostatic pressure. As a result they tend to be deep rock canyons aligned directly downslope. They yield information about ice marginal positions only if the stream course changes abruptly from subaerial to subglacial.

Lateral meltwater channels formed along the margin of ice sheets and glaciers, between ice and the adjacent hillside (Derbyshire 1962). They tend to be parallel or slightly oblique to contours rather than running directly down slope. They may not contain present-day streams. These are the most useful type of channel for reconstructing glacial history because they mark the position of the ice margin at the time they formed. Where several contemporaneous channels are present, the shape of the retreating ice sheet margin can be reconstructed. When successive channels occur the pattern of retreat can be determined. Lateral meltwater channels also indicate the ice surface gradient and thus indicate the direction of ice flow and whether or not local flow was present during deglaciation (Fulton 1967). They formed during phase 3 or the stagnant ice phase of Fulton’s (1967) four phases of deglaciation.

Direct overflow channels drained away from the ice margin, either directly down valley or over divides if water was ponded on the upstream side of ice. Where flow over divides occurred, channels were cut into cols or ridge tops. This was most common with downwasting and stagnant ice and as
the ice mass thinned during the early stages of frontal retreat. Where meltwater flowed directly down valley, channels are often not apparent although outwash plains, fans, kettled outwash terraces or other glaciofluvial deposits may be present, or the present stream may be underfit to the valley (Fulton 1967). Flow directly down valley occurred during frontal retreat, both normal and with stagnation of the snout.

3.1.1.4 Streamlined Forms

Streamlined forms such as grooves, striations, and roches moutonnees were carved by actively flowing ice under either alpine glaciers or ice sheets. They are useful indicators of the extent of active ice and they indicate ice flow direction. Rounded ridge tops or summits with streamlined forms indicate overtopping by actively flowing ice during the mountain ice sheet stage of Davis and Mathews.

3.1.2 Landforms and Materials of Glacial Deposition

Glacial materials were laid down by active or stagnant ice or by meltwater, during either advance or recession. They may have been draped over existing topography or sculpted into characteristic landforms.

3.1.2.1 Till and Erratics

Basal till is debris released by melting at the base of a glacier. Two types are recognized: lodgement till was deposited beneath actively flowing ice and basal meltout till
was deposited by melting at the base of stationary ice (Dreimanis 1976). Both types of basal till are massive, unsorted and consolidated. Lodgement till is more highly consolidated than meltout till. Clasts are subrounded to subangular and are often faceted and striated (Dreimanis 1976; Flint 1971). Clast lithology is variable and reflects the source areas. Thus if clasts of distinctive lithologies are present, ice flow direction can be determined. Clast lithology can be useful for distinguishing between tills of local alpine glaciers and ice sheets (Heller 1980).

Till fabric is the orientation of the long axes of elongate particles within basal till. Orientation developed from stresses due to glacial transport and deposition (Dreimanis 1976). Till fabric is used to indicate ice flow direction, and has been used to distinguish meltout and lodgement till (Boulton 1971). In meltout till a/b planes are parallel to the plane of deposition. In lodgement till the long axes has a tendancy to dip upflow, but this is commonly masked by the effect of bedrock topography (Boulton 1971).

Ablation till accumulated on top of downwasting ice. It differs from basal till by having a coarser texture and more angular clasts and by lack of consolidation. Large amounts of hummocky ablation till indicate widespread stagnation has occurred.

The presence of erratics on ridge tops and summits confirms that ice overtopped an area. If the source of the erratics is known ice flow direction can be determined.
3.1.2.2 Moraines and other landforms composed of till

End moraines are ridges composed of till that formed at the side or terminus of glaciers when the snout of actively flowing ice remained stationary, or ice readvanced during deglaciation. These can be associated with ice sheets or valley glaciers.

Till plains were deposited under ice sheets. When flutes, drumlins, crag and tail or other streamlined formations are present ice flow direction can be determined. The lower slopes of troughs generally are mantled with till deposited by ice flowing down the valley.

3.1.2.3 Glaciofluvial material

Glaciofluvial material was deposited by meltwater at any time during glacier advance, maximum, or retreat. It consists of sands and gravels, and may include boulder gravels. Extreme ranges and abrupt changes in grain size and sorting may occur. Glaciofluvial deposits are aggradational and as a result are generally thicker than more recent fluvial deposits. Often only a thin cap of fluvial material was deposited on top of glaciofluvial gravels before the present stream began downcutting.

Outwash was deposited by meltwater flowing away from the glacier terminus. Material deposited near the glacier snout tends to be coarse grained and have crude horizontal stratification, while more distal deposits are finer grained and may display planar cross bedding (Miall 1983). Outwash
plains that have been dissected by postglacial streams leave terraces well above present day rivers. Advance outwash was deposited in front of advancing glaciers. As a result it has a coarsening upward sequence and was often overridden by the glacier, resulting in a more consolidated deposit than typical recessional outwash.

Kame terraces and kames were deposited in contact with ice. Generally bedding is more disrupted than that of outwash deposits due to melting of ice blocks after deposition, and the terrace may be pitted with kettle holes (Flint 1971). Eskers and moulin kames are subglacial fluvial material that were preserved only under stagnant ice. Extensive kames, kame and kettle topography and hummocky glaciofluvial deposits indicate deposition in contact with stagnant ice. Kame terraces mark ice marginal positions and are most extensive in association with stagnant ice.

3.1.2.4 Glaciolacustrine material

Glaciolacustrine material is composed of fine sand, silt and minor clay. Normally it is well-laminated or well-bedded. Where thick sequences of this material have been dissected by postglacial streams, terraces are formed. The presence of glaciolacustrine deposits indicates damming of meltwater, by active or stagnant ice, during glacier advance or retreat. The extent of the deposits is indicative of the extent of the glacial lake. As these deposits commonly terminated against an ice dam, their downstream limit may mark a former ice margin.
3.1.3 Suites Of Landforms And Materials Expected With Each Type Of Glaciation

Each style of glaciation is characterized by a suite of landforms and deposits. Most individual features can form under more than one style of glaciation, so several are needed to determine style.

Mountain glaciation is characterized by erosional forms. Cirques, aretes, horns and deep, steep sided troughs dominate; these will be sharp and modified only slightly by post-glacial weathering. Till will commonly be present on lower valley slopes and valley bottoms. Deglaciation by normal frontal retreat commonly results in the deposition of outwash over till on valley floors. Occasional pockets of glaciolacustrine silts may be present on valley sides, likely due to deposition in small ice marginal ponds. Lateral and terminal moraines may be present, however these are easily destroyed by later erosion. Because glaciers were typically active throughout deglaciation, ice stagnation deposits are rare. Deglaciation by frontal retreat with stagnation of the snout, also results in glaciofluvial deposits overlying till on valley floors. At the stagnating snout, kames, and ice contact glaciolacustrine material may have been deposited. Kettle holes may be present where ice blocks became isolated in front of the retreating margin.

Ice sheet glaciation is characterized by subdued topography and rounded, streamlined forms. Sharp summits and ridge crests are absent. Till may be deposited at any
elevation, including on summits and ridge crests. Drumlins, and flutes are common, particularly on upland areas. These were formed while the ice was still active. Deglaciation was commonly by downwasting and stagnation, resulting in features such as eskers, kames, kame terraces and hummocky ablation till in valleys and lowlands. Drainage was often disrupted and large ice contact lakes formed, leaving extensive deposits of glaciolacustrine sediments.

In areas which have experienced both mountain and ice sheet glaciation, alpine forms may be rounded but still well defined. Streamlined forms are common on summits and ridges. Valleys may contain deposits of either frontal retreat, downvalley retreat of active ice, or stagnation and damming depending on whether or not local glaciers were active near the end of glaciation. Thus landforms and deposits of both mountain and ice sheet glaciation may be present and their distribution will depend on local conditions.

3.2 Field Techniques and Analysis Used To Determine Patterns and Styles of Glaciation in the Cascade Mountains

Reconstruction of a complete history of glaciation requires the identification of features formed at several stages of the glacial cycle, from the start of glacier formation until the area is once again ice free. This study relied heavily on the morphology of erosional forms such as cirques and troughs, depositional sequences on valley floors, and ice marginal features such as moraines and meltwater channels. Data
collection began with terrain mapping on air photographs and was followed by more detailed study in the field.

3.2.1 Air Photograph Interpretation

The first step in this study involved terrain mapping on 1:15,000 air photographs of each study area, using the British Columbia system of terrain classification (Howes and Kenk 1988). Under this system, the land surface is divided into polygons on the basis of surficial material type and texture, landforms, and active geological processes. On-site symbols are used to delimit the location and size of features such as cirques, meltwater channels and moraines. Many landforms, such as moraines, meltwater channels and drumlins, are more apparent on air photos than on the ground.

Emphasis was placed on identifying diagnostic landforms and materials that could be used for the interpretation of styles of glaciation and, in particular, ice marginal features such as moraines and meltwater channels.

3.2.2 Field Program

Air photo interpretation was followed by six weeks of field checking in July and August, 1993. The two goals of the field program were to verify the accuracy of terrain mapping and to make additional observations and measurements. Rock weathering and valley fill stratigraphy were recorded, and sediment cores were obtained from a moraine-dammed lake.
3.2.2.1 Surficial Materials

Surficial materials were examined largely in road cuts and stream banks where exposures several metres high were found. This allowed two to three stratigraphic units to be observed and, in several locations, complete sections to bedrock were obtained. Where natural exposures were not present, soil pits were dug, however it was not possible to obtain a good exposure by this method due to the thickness of overlying colluvium (up to 2m) and the bouldery nature of many of the deposits.

Particular attention was paid to the characteristics of till, glaciolacustrine and glaciofluvial materials. At each exposure texture, consolidation, colour, clast lithology, sorting, stratification and stratigraphic position were described for each material present.

Several till fabric measurements were taken in both the Stoyoma and Outram study areas. However, it was difficult to find exposures of sufficient size to obtain the orientation of 50 elongate stones. Those located tended to be clustered along several logging roads, so it was not possible to obtain samples throughout the study area. The samples collected showed fabrics dipping both up and down valleys, but not enough samples were obtained for a pattern to emerge. For these reasons it was decided that fabric measurement was not an efficient use of field time and it was abandoned.
3.2.2.2 Rock Weathering

Rock weathering has been widely used as a method of relative dating, and also for absolute dating where calibration curves have been established (Chinn 1981).

In British Columbia, most rock surfaces were eroded by ice during the last glaciation so they tend to be relatively smooth and unweathered. The exception is peaks that remained above the ice throughout the last glaciation. Thus the upper limit of glaciation can be identified as the highest level of glacial abrasion of bedrock. Peaks above this limit will have much older surfaces than other areas and thus rock should be distinctly more weathered, except where frost shattering and rockfall have exposed fresh rock. On this basis rock weathering was examined to determine if all summits were overtopped by ice sheets or if nunataks existed.

Depth of weathering depends on rock mineralogy and grain size, as well as time (Chinn 1981) so it is important to compare only similar rock types. For this reason measurements were confined to granitic rocks. Granitic rocks do not always form clearly defined weathering rinds. They are subject to granular disintegration: with time surfaces become rougher, more resistant grains are exposed, weathering pits become deeper and grus accumulates (Benedict 1993).

The degree of weathering was measured using a Schmidt hammer. This is a device designed for measuring the hardness of concrete (Matthews and Shakesby 1984; Day 1980). It is a spring loaded hammer which strikes the rock surface and
measures the percent rebound. The harder the surface the greater the rebound. The Schmidt hammer provides a numerical value for hardness which can easily be compared between sites. As hardness is correlated with weathering depth, this is a simple, objective method of determining the relative amount of rock weathering. The most weathered surfaces will be softest and should produce the lowest Schmidt hammer readings.

Variables such as moisture and duration of snowpack can affect weathering rates significantly (Benedict 1993; Hall 1993). These are hard to control when selecting sample sites because amount of precipitation and evaporation both vary locally and with altitude. However, these variables acting over the last 10,000 years since deglaciation should produce smaller differences than those between glaciated and non-glaciated surfaces. If nunataks were present there should be abrupt differences in Schmidt hammer readings. On granitic rocks in the Okanagan Range of the Cascade Mountains Ryder (1989, personal communication 1995) found significant differences in hardness above and below the upper limit of Fraser Glaciation ice.

Measurements were taken at as many levels as possible from low in valleys to summits and ridge crests. Where continuous outcrop was present measurements were taken roughly every 200 m elevation. At each site 50 measurements were taken, either 5 per boulder on 10 boulders in moraines or erratics, or along a grid pattern on rock outcrops. Readings on transported boulders should give values representative of
glaciated surfaces and unglaciated surfaces should have lower readings.

3.2.3 Reconstruction of Glacial History

After completion of field work the mapping on air photos was studied and corrected. The information was then transferred to 1:20,000 topographic maps and direct evidence of glaciation was highlighted.

Early Fraser Glaciation glaciers were reconstructed by fitting glaciers into cirques and troughs. Glaciers were fit into these features by assuming the glacier started at the upper limit of the steep cirque headwall and that glaciers extended down valleys to the U-V transition. Total cirque area of each valley glacier was calculated by measuring the area between the top of the headwall and the end of the floor (taken as the point where the valley narrows and steepens), and summing the areas of each cirque feeding a valley glacier. Cirque areas were measured from 1:50,000 topographic maps and so are approximate values. Trough length was taken as the longest possible length, that is from the furthest up-stream cirque rim to the U-V transition. The total cirque area was plotted against trough length to determine if a relationship existed. Goldthwait (1970) and Flint (1971) predict a relationship if this transition reflects the limit of valley glaciation.

All ice marginal features in the study areas were assumed to have been formed during deglaciation. In no study area is
there sufficient marginal features to determine ice surface slope. It was also assumed that marginal features located at similar elevations within each study area were formed at roughly the same time. Marginal features at similar elevations were plotted on separate small scale maps and connected. This created ice form lines similar to those of Fulton (1967) which indicate the shape of the receding ice margin, but not its age. Due to the lack of marginal features all form lines are very approximate. Ice form lines may mark a position in which the ice front was stationary for only a very brief period of time. These maps were then compared with the distribution of surficial deposits to determine if the indicated pattern of ice retreat was consistent with other evidence, such as stratigraphy of valley fill deposits.

In this manner several maps were produced for each area showing the inferred pattern of mountain glaciation during build up, and a series of inferred ice levels during retreat. These were compared in order to develop a model of glaciation applicable to all three areas. The maps and details of their construction will be discussed in chapters 4, 5, and 6.
CHAPTER 4 MT. STOYOMA AREA

4.1 Introduction

The Mount Stoyoma study area is near the northern limit of the Cascade Mountains (Figure 1.1). At this point the Cascades begin to grade into the Thompson Plateau. The area was selected to encompass topographic variability around moraines which were known to be located in two cirques (Dr. J.M. Ryder personal communication 1992). A plateau area to the east of the cirques and troughs to the west were included in an area bounded on the north by Miner's Creek and on the south by Spius Creek (Figure 4.2).

The entire study area is underlain by the Upper Jurassic and Lower Cretaceous Eagle Granodiorite. This is a coarse grained grey rock of variable mineralogy and foliation (Monger 1989; Monger and McMillan 1989). Outcrops may be competent and rounded or strongly fractured. Nonfoliated white granodiorite also occurs in some locations. This rock is often slightly pitted and is generally more competent than the foliated granodiorite.

4.2 Topography and Erosional Landforms

The eastern part of the study area is a gently undulating plateau, much like the Thompson Plateau, with rounded rocky hills rising to 1700 m (Figure 4.3). Relief is less than 500 m. Lateral meltwater channels trend across slopes at elevations between 1200 m and 1800 m (Figure 4.2). They are discontinuous depressions, 2 to 3 m deep, roughly 10 m wide and 100 to 500 m long. All
Legend for Stoyoma Area Maps

- Cirques
- Breached col
- Moraine
- Meltwater Channel
- Glacier
- Drumlín
- Location of Stratigraphic Sections
- Areas With Clay Rich-Till
- Glaciofluvial Terraces

Figure 4.1: Legend for Stoyoma area maps
Figure 4.2: Location map of the Mt. Stoyoma study area.
observed channels were cut into till. They are often clustered, but isolated channels also occur.

The western half of the study area is higher and steeper with rounded summits, which are remnants of the Miocene upland surface, rising to 2300 m (Figure 4.4). Relief is roughly 900 m. Cirques occur around most summits, horns and aretes are lacking.

Some cirques are simple, single basins, others are complex, consisting of up to six cirque basins. Large complex cirques are present at the heads of Miner’s and Spius Creeks; these face north and east, and their dimensions indicate that they held the largest alpine glaciers. In addition eleven simple cirques are present at the heads of smaller troughs. Of these five face north, two face east, two face west, and two face south (Figure
4.2). All cirques have well developed cliffy headwalls, gently sloping floors and smooth rounded rims. Tarns are common.

Upper valleys are broad U-shaped troughs which commonly are prominently stepped. Passes have all been widened, smoothed and deepened by ice flowing through cols between valleys, but not to the extent that cirque headwalls have been lost. The exception is the south fork of Spius Creek which heads in a through trough, 7 kilometres south of the study area. Most valleys are V-shaped in their lower reaches. The transition from U to V shaped occurs over a distance of roughly 1 km.

Miners Creek flows in a 5 to 10 m deep bedrock canyon for most of its length. The morphology suggests a subglacial meltwater channel. Benches on both sides of the creek appear to be cut into bedrock, with pockets of oxidized boulder gravel on
the surface, which could be a veneer of till or glaciofluvial material.

4.3 Depositional Landforms and Surficial Materials

In the steeper western half of the study area surficial materials are generally restricted to lower slopes and valley floors. Bedrock is exposed on all summits and ridge tops. Colluvial cones and talus slopes are dominant on steep slopes below cliffs, especially in upper valleys (Map 1). Rock outcrops are visible on the floors of most cirques and upper valleys, indicating that deposits in these areas are patchy and thin.

The plateau area to the east is largely covered by a blanket of basal till, with rock outcrops at higher levels. Drumlins indicate ice flow to the southwest, transverse to local valleys. Glaciofluvial terraces occur above Prospect Creek and in the valley of Spius Creek.

4.3.1 Till

Till blankets much of the gently undulating plateau in the eastern half of the study area and is commonly exposed in road cuts. Thickness is unknown, however exposures of 2 to 4 metres are common. Till also likely blankets lower slopes in the western part of the area.

 Throughout the area till is highly consolidated and massive, with subrounded to subangular clasts of a mixture of lithologies. On the plateau, matrix texture is silty sand. In the lower part of Miner's Creek valley, in Spius Creek valley and several of its tributary valleys cohesive clay rich-till occurs (Figure 4.5).
Figure 4.5: Map showing the extent of clay rich till within the study area.
This till may overlie or underlie glaciolacustrine or glaciofluvial deposits (see Figure 4.6 sections 2, 3, 4 and 7). Along Miner's Creek no glaciofluvial or glaciolacustrine materials were observed near the clay-rich till. In sandy tills the dominant clast lithology is foliated granodiorite, while in some clay rich tills mafic volcanics are more abundant. The nearest source area for the mafic volcanics is the Lower Cretaceous Spences Bridge Group which outcrops north of the study area (Monger and McMillan 1989), thus the different till textures cannot be due solely to local variations in lithology.

The location of clay rich tills in valley bottoms, above glaciolacustrine material as in section 7, suggests that this till was probably derived from glaciolacustrine material. Glaciolacustrine material could have formed as a result of damming of these valleys by ice advancing from the east. Fine grained sediment deposited in lakes in front of the advancing ice would then be incorporated into till as the lake sediments were overridden.

4.3.2 Moraines

Moraines dam cirque lakes at 1700 m on Mt. Hewitt Bostock at the head of Miner's Creek, and at 1860 m at Cabin Lake (Figure 4.2). The Miner's Creek moraine is strongly arcuate into the cirque, it is roughly 400 m long at the lake and extends a further 400 m along the north side of the valley in several discontinuous, en echelon strands. These are roughly level or climb slightly downvalley, suggesting that the glacier was actively flowing up valley at the time of deposition. The Cabin
Figure 4.6: Selected stratigraphic sections. See figure 4.2 for the locations of sections.
Lake moraine is linear and slightly longer than the cirque width. It is roughly 500 m long, 15 m high and has a broad flat top (Figure 4.7).

Both moraines have a plan form that is more consistent with deposition by Cordilleran Ice flowing up valley, than by cirque glaciers flowing downvalley. The Cabin Lake moraine has a fine sand matrix, and 40 to 60% clasts dominantly of foliated granodiorite and lesser mafic volcanics. The cirque headwall is composed entirely of foliated granodiorite. Thus the orientation, composition and length of the Cabin Lake moraine are inconsistent with deposition by a cirque glacier.

Figure 4.7: Cabin Lake with moraine ridge to the left of the lake.
4.3.3 Terraces, Glaciofluvial and Glaciolacustrine Deposits

Glaciofluvial deposits comprise well defined terraces which are present in Spius Creek valley and on the south side of Prospect Creek (Map 1). In the study area the Spius Creek terraces begin near the junction of the north and south forks. Along the south fork a terrace is present 30 m above the creek. It is 800 m wide and extends for at least 3 km upstream of the study area (Figure 4.2). A terrace generally less than 100m wide extends 5 km downstream from the junction on both sides of the creek, but is broken by colluvial fans in several places. At its downstream limit the terrace is 180 m above Spius Creek. Along Prospect Creek terraces are smaller. At the junction of Miner’s and Prospect creeks, a 200 m wide terrace extends 400 m downstream and a second terrace of similar size is present 500 m downstream, both are at 925 m elevation or 90 m above Prospect Creek. A third terrace is present straddling a small tributary at 1150 m elevation or 480 m above Prospect Creek (Fig 4.2).

Near the Junction of Miner’s and Prospect Creeks a blanket of glaciofluvial material overlies till on a moderate slope. This material could have been deposited adjacent to an ice margin continuously retreating downslope towards Prospect Creek.

Glaciofluvial deposits consist of loose, stratified sand and gravel; well-sorted, laminated sand is interbedded with more massive gravel. Locally deposits are massive and coarse-textured with clast sizes to large cobbles. Glaciofluvial deposits are several metres thick and commonly overlie till (Figure 4.6 sections 2 and 7). In section 3, clay-rich till appears to overlie glaciofluvial gravels, although the contact was not well
exposed. If this relationship is true this could be advance outwash, or fluvial gravels deposited by the creek before ice advanced up the valley from the east.

Glaciolacustrine material is rare in this area. No extensive deposits were observed although pockets up to 2 m thick are present in road cuts and stream banks along several creeks (Map 1). This material is cohesive, thinly laminated, sand, silt, and clay with occasional stones. Glaciolacustrine material overlies glaciofluvial gravels in a terrace along Prospect Creek (Figure 4.6 section 1). Here the glaciolacustrine material must have been deposited during deglaciation. Section 7 is a road cut through a terrace in Spius Creek. In this section laminated silts and clays occur below till and glaciofluvial gravels, indicating that the glaciolacustrine material was deposited before ice advanced into the valley. In an unnamed tributary of Spius Creek the timing of deposition of the glaciolacustrine material is not clear. Clay rich till and glaciolacustrine silts are exposed in road cuts and stream banks (sections 4, 5, and 6), however their stratigraphic relation is not clear, as no contacts are exposed. Section 4 is a composite, including a steam cut which exposed laminated silts and clays with occasional dropstones, and a road cut above this section which exposed clay rich till. Sections 5 and 6 are located across the creek from section 4 and at a similar elevation. No terraces are present in this area. The glaciolacustrine material in not consolidated, while the clay rich till is highly consolidated. Likely these glaciolacustrine sediments were deposited in lakes dammed by retreating ice, but there is no firm evidence of this.
4.4 Rock Weathering

Rock weathering was difficult to assess due to the variable lithology and fractured nature of the rock. It was difficult to be certain that surfaces dated from the last glaciation and were not the result of more recent fracturing. An effort was made to sample only the more rounded surfaces on top of outcrops rather than fracture planes. Bustin and Mathews (1979) found that relatively small differences in biotite content had large effects on weathering rates in granodiorite in till. For this reason it is important that lithology not vary between sites. However lithology and degree of foliation are highly variable in this study area so it was difficult to distinguish the effects of length of exposure and rock type on the variability of Schmidt hammer readings.

Rock weathering was measured at 11 sites between 1840 and 2270 m. The lowest site comprised boulders in the Cabin Lake moraine. These boulders began to weather as soon as the ice left the moraine, thus any unglaciated areas should be considerably more weathered.

At all sites there is a wide spread in the data (Appendix 1). There is a decrease in mean Schmidt hammer readings above 2100 m (Figure 4.8), but no sharp change as would be expected if summits had remained above the ice. The decrease in readings with elevation could be due to such factors as increased freeze-thaw activity or local snowpatch effects. There is also no grus formation or extensive pitting present at any elevation.
Figure 4.8: Relation between mean Schmidt hammer readings and elevation. Bars represent +/- 2 standard deviations of the mean.

4.5 Glacial History

Ice form lines have been reconstructed to show the extent of glaciers at various stages of Fraser Glaciation. No dates have been determined for any of these maps, thus they indicate only the sequence of events and not absolute timing. The elapsed time between successive maps is likely highly variable.

4.5.1 Valley Glacier Phase

The presence of well developed erosional landforms of alpine glaciation around higher summits indicates that alpine glaciers formed at the start of Fraser Glaciation. The rounded nature of
all alpine features suggests that they formed before an ice sheet covered the area. This phase is likely equivalent to the prolonged alpine glaciation that Clague (1981) described for the Coast Mountains. Glaciers of this phase were reconstructed by assuming that glaciers extended from the top of cirques headwalls downvalley to the point where valleys change from U to V shape cross profile (Figure 4.9).

The longest glaciers occurred in troughs with the largest total cirque area at their heads (Figure 4.10), as would be expected if the U-V transition represents the termini of alpine glaciers (Goldthwait 1970) (see Chapter 3 for method used to calculate cirque area). Snowline, as determined from the lowest north facing cirques, was roughly 1500 m during this period.

4.5.2 Ice Sheet Stage

At the Fraser Glaciation maximum all summits were overtopped. Evidence of this consists of: well rounded ridge tops, relatively unweathered bedrock with little change in weathering from moraines to summits, and the erosion of cols.

Ice flow direction appears to have changed from westward to southward as the Cordilleran Ice Sheet thickened and advanced onto the plateau area. If the clay-rich tills in east draining valley bottoms are due to overriding of earlier glaciolacustrine material, they must have been deposited by ice advancing from the east, from the Thompson Plateau rather than from the Coast Mountains. This westward ice flow would have occurred early in the ice sheet stage. Drumlins on the Spius Plateau indicate ice flow to the southwest. As the drumlins are not at high elevations
Figure 4.9: Map of early Fraser Glaciation alpine glaciers in the Mt. Stoyoma area.
Figure 4.10: Plot of cirque area versus trough length showing a weak non-linear relation.

they may have been formed before or after the glacial maximum. This is transverse to most local valleys, which allowed the preservation of the fine textured till in valley bottoms. This ice flow direction also suggests that local glaciers were not a significant influence on ice flow direction, and may not have contributed to the ice sheet. North of the study area ice flow, at the glacial maximum, was south to southeast (Ryder 1981), more consistent with a source in the Coast Mountains.
4.5.3 Deglaciation

Evidence of deglaciation on the Spius Plateau is relatively abundant. Lateral meltwater channels give good control on the ice margins. The moraines at Cabin Lake and the head of Miner's Creek provide an ice position at the western margin of the plateau and into the mountainous area. These features are consistent with retreat of an active ice margin away from the mountains.

Ice form lines were drawn by connecting ice marginal features of roughly equal elevation. The earliest ice marginal features are the two moraines at 1700 and 1860 m (Figure 4.11). These mark the ice front position during a small readvance of active ice. A second form line (Figure 4.12) is based on two clusters of meltwater channels at 1580 m. Below this level meltwater channels become more sparse. Successively lower ice margins are approximately parallel. There are no deposits in the study area which indicate stagnation, so it is likely that ice remained active and retreated continuously.

The moraine in Miner's Creek valley is evidence that no cirque glacier was present here during deglaciation, and hence frontal retreat of local glaciers could not have occurred. This is a relatively high east facing cirque located below 2160 m Mt. Hewitt Bostock. This is one of the higher peaks in the northern Cascades, with a relatively extensive area above 1700 m. This is the type of location which seems most likely to have functioned as an accumulation zone. The fact that it was ice free before Cordilleran ice disappeared from the area is evidence that all cirques were below snowline by the time the ice sheet had receded locally. Thus snowline was above 1700 m throughout deglaciation.
Figure 4.11: Ice form line during deglaciation, based on moraines at Cabin Lake and in Miner's Creek.
Figure 4.12: Ice form line during deglaciation based on meltwater channels at 1580 m.
Both Spius and Miner's creeks should have been dammed or partly dammed throughout deglaciation by ice receding downvalley. However no deposits indicative of damming were found in any valleys, with the possible exception of the unnamed creek at section 4 (Figure 4.2). This suggests that meltwater escaped subglacially or that glaciolacustrine deposits were either too thin to be found or mostly destroyed by later erosion. The bedrock canyon in Miner's Creek is possible support for subglacial drainage, but poses a problem that will be discussed further in chapter 7.

Spius Creek contains extensive glaciofluvial outwash terraces with small pockets of glaciolacustrine silt. The former deposits are more consistent with up valley frontal retreat than with downvalley recession. However it is unlikely that the cirques at the head of Spius Creek held glaciers because they are lower than those at the head of Miner's Creek and have roughly the same orientation. In addition the head of South Spius Creek is a low pass, or through trough. A possible source of the glaciofluvial material is from a lobe of Cordilleran ice from the Coast Mountains pushing through the col at the head of Spius Creek.

No landforms of deglaciation were found in the troughs in the western portion of the study area. A meltwater channel cuts across the divide into Spius Creek, indicating that these troughs may have held ice longer that those to the east.
4.5.3.1 Sediment Cores

Two complete sediment cores were obtained from Cabin Lake. The first is 1.85 m long and the second 0.90 m. The bulk of both cores is composed of soft, brown gyttja, with a layer of Mazama tephra roughly halfway down. Both cores terminate in a layer of uniform grey clay. A wood fragment was found in the second core in silty gyttja at a depth of 70 cm, 6 cm above the basal clay. This fragment was dated at 9319 + 120 years BP (sample number TO-4325).

This is a minimum age for the deposition of the Cabin Lake moraine. If uniform deposition is assumed, roughly 2500 years would have been required to deposit the gyttja and clay below the wood sample. This gives a very tentative estimate of 11,800 years BP for the age of the lake, slightly older than Souch’s (1989) 11,100 years BP for deglaciation of Kwoiek Creek valley. This is consistent because the Cascade Mountains are lower and drier than the summits at the head of Kwoiek Creek so would likely have become ice free first.

4.6 Neoglacialization

No deposits or erosional features related to a Neoglacial advance were seen within this area. Nivation hollows in talus slopes below the headwalls of north and east facing cirques near Mt. Hewitt Bostock are evidence of late lying snowpatches. The undisturbed lateral moraines high in Miner’s Creek and at Cabin lake are evidence that cirque glaciers have not existed since Fraser deglaciation.
CHAPTER 5 ANDERSON RIVER AREA

5.1 Introduction

The Anderson River study area is centred around a cluster of prominent horns between the North and South Forks of Anderson River (Figure 5.2). Horns are uncommon in other parts of the Cascade Mountains so it was thought that this area may have undergone a different style of glaciation from other areas. The Anderson River flows to the northwest, into the Fraser River, and thus towards the Coast Mountains.

Granodiorite of the Eocene Needle Peak Pluton forms all higher summits. This is a competent unit with widely spaced joints. The Dewdney Creek Formation of the Lower Jurassic Ladner Group underlies lower areas. This is a mixed unit containing sandstone, argillite and local mafic to intermediate volcanics. Generally rock of this unit is severely fractured and weak (Monger 1989). At the junction of the North and South Forks of the Anderson River bedrock is highly fractured, fine grained, black argillite.

5.2 Topography and Erosional Landforms

Horns between the North and South Forks are between 1770 and 1980 m, and surrounding ridges reach elevations of between 1520 and 1700 m. Total relief is 600 to 900 m. Horns are prominent but summits and ridges are well rounded and smooth (Figure 5.3). Large subangular granodiorite boulders, which
Figure 5.1: Legend for Anderson River area maps.
Figure 5.2: Location Map of the Anderson River study area.
do not appear to have weathered in place, are common on many summits. Fluting on top of Chamois Peak and on ridges 3 km to the west of the study area indicate ice flow toward the south to south-southwest, approximately parallel to the Fraser River.

The North and South Forks of Anderson River flow in large Icelandic style troughs which are U-shaped throughout the study area. These head in broad glacially carved cols which often partially breach pre-existing cirques (Figure 5.4). Small alpine troughs occur only on the north side of lower elevation ridges, particularly the ridge between the North Fork and East Anderson River.

Cirques occur most commonly on the north side of ridges, several face east or west and only one south facing cirque was observed. Those at the head of the North and South Forks are complex shallow features composed of several cirque basins. They are 1 km to 3 km wide with rounded rims, low cliffy headwalls and commonly have no well defined cirque floors. Cirques at the head of small alpine troughs are much smaller, simple features, generally less than 1 km wide and also with rounded rims. All cirques are strongly degraded and many headwalls and sidewalls are breached. Where present, cirque floors are gently sloping, and commonly rocky. The cirque floor below Gemse Peak contains well defined grooves, which are parallel to the valley. In general north facing slopes are steep cliffs, while south facing slopes are moderately angled slabs.
Figure 5.3: Horns between the North and South forks of the Anderson River.

Several small meltwater channels cross the ridge on the west side of the South Fork (Figure 5.2). Their slope indicates flow to the west. These channels must have drained lakes ponded by ice in the South Fork. A large, steep sided bedrock canyon, that is probably a subglacial meltwater channel, flows into Boston Bar Creek from a col south of the head of South Fork.

5.3 Depositional Landforms and Surficial Materials

Surficial materials are present on lower slopes and valley floors throughout the study area. Summits and steep upper
slopes are rock. Colluvial cones and talus slopes are common at the base of steep rock slopes (Map 2).

Figure 5.4 Head of the North Fork of the Anderson River, cirque headwall is rounded and clearly overridden.

5.3.1 Till

Well consolidated basal till occurs on lower slopes and in valley fill throughout the study area. Generally till is covered with 1 to 2 m of colluvium on lower slopes or with glaciofluvial and glaciolacustrine deposits in valley fill. Till blanketed slopes tend to display a well developed pattern of subparallel gullies. Bedrock commonly outcrops in stream beds of both Anderson River and tributary creeks, indicating
that till on the lower slopes is generally less than 5 m and commonly only 1 to 2 m thick.

In the upper parts of the valleys the till has a coarse sand matrix. The clasts in this till are dominantly granitic, locally with minor amounts of volcanic and sedimentary rocks. At the junction of the North and South forks the till matrix is clay-rich and highly cohesive. This matrix texture is very similar to that of glaciolacustrine deposits located further upstream in both forks. The clay-rich till contains dominantly black argillite and volcanic clasts. It thus seems probable that the matrix texture of the till reflects both local bedrock geology, and reworking of fine textured valley fill. The location of the fine-textured till is consistent with its derivation from lake sediments, the result of damming of Anderson River by a lobe of ice from the Coast Mountains during early Fraser Glaciation.

5.3.2 Terraces, Glaciofluvial and Glaciolacustrine deposits

The typical valley fill sequence in the upper parts of both forks of Anderson River is a glaciofluvial terrace cap 2 to 5 m thick, overlying 0.5 to 3 m of glaciolacustrine silts and clays, that in turn overlie several metres of basal till (Figure 5.5 sections 1 and 9). These deposits are dissected by meltwater channels giving the valley floor a hummocky appearance.

Below sections 1 and 8 the North and South Forks are incised into till, with occasional bedrock outcrops.
Figure 5.5: Selected stratigraphic sections; see figure 5.2 for locations
Glaciolacustrine sediments are generally well-laminated to massive silts and clays, with 10 cm to 30 cm interbeds of fine to coarse sand. They are very cohesive and poorly drained. Deposits are most often found as part of the valley fill, usually underlying glaciofluvial material, and vary from 0.5 m to greater than 5 m in thickness. Occasionally pockets of laminated fine sands, or silts and clays are present on valley sides above the level of terraces (Map 2, Figure 5.5 sections 2 and 4). These are interpreted as ice-marginal lake deposits.

Glaciofluvial terraces are present along the North Fork upstream of section 1, along the South Fork upstream of section 7, and along the Anderson River at and below the junction of the two forks (Map 2, Figure 5.2). Road cuts and stream channels expose a complete valley fill sequence from terrace surface to bedrock in several locations. The highest terrace is located downstream of the junction of the North and South Forks, this terrace is at 850 m, 180 m above river level and extends for 3 km downstream. Roughly 30 m of glaciofluvial gravels form a cap and below this is clay-rich till. A short distance upstream at the junction the terrace level is 790 m or 120 m above the river (Map 2). This terrace is composed dominantly of clay-rich till with a thin cap (< 2 m) of glaciofluvial gravels.

Glaciofluvial deposits are loose, well sorted sand and sandy gravel, ranging in thickness from 2 m to greater than 10 m. Sand beds are massive to laminated, while gravel beds are
generally massive to weakly stratified. Bedding is horizontal. Well developed fluvial structures are lacking. These deposits are likely outwash deposited in front of a valley glacier receding by normal frontal retreat.

5.4 Rock Weathering

The Anderson River study area contains several Neoglacial moraines which enclose highly polished slabs of granodiorite. Slabs are also present on the floors of older Fraser Glaciation cirques outside of the Neoglacial moraines, and on horns. This allowed more direct observation of granodiorite weathering than the other areas.

Neoglacial slabs have very smooth slippery surfaces. The Schmidt hammer readings for these surfaces were very tightly clustered, with a mean of 66 and standard deviation of 1.8 (Appendix 2).

All surfaces outside of the Neoglacial moraines, including cirque floors, summits, and boulders on summits displayed a similar degree of weathering. These surfaces are rougher and more variable than Neoglacial slabs, individual grains have weathered out and shallow pits, generally less than 1 cm deep and 15 cm across are common. Small flakes 1 to 2 grains thick and several centimetres long can be easily pulled off. Scattered glacially polished surfaces, similar to the Neoglacial slabs, are also present. Means of 11 Schmidt hammer samples are between 40 and 50 with standard deviations of 8 to 12 (Appendix 2). Variability within individual samples
depends on local conditions, with lowest values occurring on or near flakes and higher values occurring on remnant polished surfaces.

There is a weak trend of decreasing mean Schmidt hammer readings with elevation (Figure 5.6). However, if summits were unglaciated they should have significantly lower mean readings and possibly higher standard deviations than slabs in cirques. The difference between glaciated and unglaciated should be of a similar order to the difference between Neoglacial and Fraser Glaciation (see Figure 5.6). Ryder (Personal

Figure 5.6 Plot of mean Schmidt hammer values versus elevation. Bars represent +/- 2 standard deviations of the mean.
communication 1995) found that in the Okanagan Range of the Cascade Mountains surfaces which were above the elevation reached by the Cordilleran Ice Sheet had much lower Schmidt hammer readings. The observed variability must therefore be due to factors other than glaciation, such as local moisture conditions, duration of snow patches, or slight differences in mineralogy of the granodiorite.

5.5 Glacial History

5.5.1 Valley Glacier Phase

Prominent horns, cirque remnants, and alpine troughs indicate that, as in the Mt. Stoyoma Area, glaciers formed around higher peaks and on north sides of ridges during the early stages of glaciation. The Anderson River valley narrows downstream. However no clear transition from trough to V shaped valley occurs above its confluence with the Fraser River. It is thus probable that valley glaciers flowed towards the Fraser River and were confluent with a lobe of Coast Mountain ice advancing up Anderson River relatively early in the glacial cycle (likely during the intense alpine phase of Davis and Mathews). A lake may have been temporarily ponded in the Anderson River before local and Coast Mountain ice coalesced. The elevation of the lowest north facing cirques indicates snowline was roughly 1500 m during this stage.

Two small troughs on the north side of Anderson River Mountain become V-shaped a short distance above the North
Figure 5.7: Early Fraser Glaciation alpine glaciers.
Fork, suggesting that these glaciers were not confluent with
the main valley glacier (Figure 5.7). A small trough on the
east side of the South Fork, between Chamois and Ibex also
becomes V-shaped before joining the South Fork.

Early Fraser Glaciation alpine glaciers were
reconstructed on the basis of cirque remnants, and the U-V
transition, where present. Valley glaciers in the main river
valleys are drawn with dotted lines as their terminal
positions are unknown (Figure 5.7).

5.5.2 Ice sheet Stage

Evidence that an ice sheet covered all peaks in the Anderson
River Area is extensive. Cirque rims and horns are rounded and
smooth, the summit of Chamois Peak is fluted, and striations
occur near the summit of 2040 m Yak Peak, 3 km east of the
study area, and 60 m higher than any peaks in the study area.
Large Icelandic style troughs are common. Schmidt hammer data
indicates no significant increase in weathering from low on
ridges to summits. Subrounded granodiorite erratics which are
too fresh to have weathered in place are common on summits.

Ice flow at the Fraser maximum as indicated by fluting
was to the south.

5.5.3 Deglaciation

Little evidence exists to indicate the pattern of
deglaciation, as no unequivocal ice marginal features were
found in this study area. Small pockets of glaciolacustrine
deposits on hillslopes are thought to have been deposited in small lakes ponded by thinning ice. The degraded nature of the cirques and the lack of Fraser recessional moraines suggests that cirque glaciers were not very active at the end of glaciation. However, there were likely small remnant alpine glaciers, which supplied debris and meltwater to form the glaciofluvial terraces.

Glaciolacustrine deposits occur throughout both forks of the Anderson river, up to an elevation of 1370 m. These are thin deposits which may have been laid down in small lakes ponded between ice and valley walls, or more likely, on the upstream side of ice receding down valleys toward the Coast Mountains. These lakes possibly followed the receding ice down valley, so never became very large. Drainage of these lakes is problematical as no lateral or subglacial meltwater channels were found. Upstream of the lakes, creeks deposited fluvial and glaciofluvial material on top of the glaciolacustrine silts.

At the start of deglaciation the ice sheet thinned and highest horns were exposed as nunataks. Ice was still thick enough to flow and erosion of the cols continued. It is probable that the large subglacial meltwater channel which drains into Boston Bar Creek (Figure 5.2) was largely formed at this time, before ice receded below the col at the head of the channel. The Fraser River valley likely still held a large volume of Coast Mountain ice, blocking drainage in this direction. Thus large amounts of meltwater from rapidly
receding glaciers could have been funnelled through this channel. The col at the head of the meltwater channel is one of the lowest passes between the Fraser River/Anderson River system, and the Boston Bar Creek/Coquihalla River system, above their junction. Boston Bar Creek and Coquihalla River may have been an alternate route to the Fraser River for meltwater from a large part of the northern Cascades.

The ice form line in figure 5.8 is based on the upper limit of glaciolacustrine deposits at 1370 m. Glaciolacustrine material occurs up to this level in both the North and South Forks and also a tributary of the North Fork (Figure 5.2 sections 2, 5, and 11). The form line was drawn to show the highest lake levels. This fits with the valley fill sequence, but is not based on any ice marginal features, so it is only a very approximate and likely transient ice position.

Siwash Pass is a broad pass north of Siwash Creek and 2 kilometres west of the study area. It is filled with hummocky deposits, which are probably ice contact glaciofluvial deposits, or ablation till. The north fork of Siwash Creek, makes a sharp bend at this point. Its most direct route to the Fraser is to the northwest through the pass, instead the creek turns to the southwest and flows into Siwash Creek. This pass is at a similar elevation to the high terrace below the junction of the forks of the Anderson River. It is thus likely that they formed at roughly the same time, likely adjacent to a lobe of ice from the Coast Mountains which was generally active, with some stagnation at the snout, and retreating down
Figure 5.8: An ice level during deglaciation, based on the uppermost lacustrine sediments at roughly 1370 m.
Figure 5.9: An ice level near the end of deglaciation, based on the terrace below the junction of the North and South Forks and hummocky deposits in a pass to the south.
valley towards the Coast Mountains (Figure 5.9). The ice would have blocked drainage to the northwest forcing Siwash Creek into its present channel.

It is probable that both deglaciation maps show very short lived ice margins. The lack of ice marginal and stagnation features suggests that deglaciation was rapid and continuous.

5.6 Neoglacialation

Neoglacial activity in the Anderson River area was restricted to north facing bowls at the base of cliffs, where avalanches fed the accumulation zones. Fresh moraines, found within a few hundred metres of headwalls, tend to be small features, less than 10 m high. Neoglacial cirques are very steep with little floor area. Head walls are fresh and granodiorite floors are highly polished.
6.1 Introduction

The Outram area (Figure 1.1) is located in the Hozameen Range of the Cascade Mountains. This range, along with the Skagit Range to the south, contains the highest and most rugged mountains in the British Columbia Cascade Mountains. Lateral moraines of the Cordilleran Ice Sheet were known to be present in two cirques on the south side of Mt. Outram (J.M. Ryder personal communication 1992). The study area was selected to encompass topographic variation in an area around these cirques. The area includes Nicolum and Sumallo valleys which form a large through trough, craggy summits and alpine troughs around Mt. Outram, and the Podunk Plateau area between Montigny and Sowaqua Creeks (Figure 6.2).

The area is underlain by mixed sedimentary and mafic volcanic rocks of the Permian to Jurassic Hozameen Complex. This is intruded by granodiorite of the Eocene Mt. Outram Pluton (Monger 1989). Granodiorite underlies much of the area south of Sowaqua Creek with the exception of the highest summits, which are volcanic, and a bench along the south side of Sowaqua Creek, which is underlain by fine grained sediments and volcanics. Ridges to the north are largely underlain by sandstones and conglomerates.
Legend for Mt. Outram Area Maps

- Cirques
- Breached Col
- Moraine
- Meltwater Channel
- Glacier, position roughly known
- Glacier, position unknown
- Location of Stratigraphic Sections
- Kame Terraces
- Hope Slide
- Glaciofluvial Terrace
- Ice flow indicator, direction unknown

Figure 6.1: Legend for Mt. Outram area maps.
Figure 6.2: Location map of the Mt. Outram Study Area.
6.2 Topography and Erosional Landforms

The area between Nicolum-Sumallo trough and Sowaqua Creek contains the best developed alpine glacial features and the highest peaks in the study area, including 2440 m Mt. Outram (Figure 6.3). Total relief is 1370 m. The highest summits are relatively sharp. There are well defined cirques, many with sharp rims, around most summits (Figure 6.2).

Figure 6.3: Mt. Outram and surrounding south facing cirques, from Johnson Peak.

Podunk Plateau in the northeastern part of the study area consists of remnants of a Miocene erosion surface (see Mathews, 1968). This area is characterized by low rounded summits, the highest being 1980 m Mt. Davis, and broad gently
sloping upland areas. The total relief is 300 m. The few cirques present are all incised below the plateau surface and are strongly degraded, with rounded rims and low headwalls. The plateau surface is irregular and boggy. Hummocky rock outcrops are common.

A total of 19 cirques is present in the study area. Of these eight face north, four west, four south and 3 east. Most cirques have prominent, steep, cliffy headwalls and well defined flat to gently sloping floors. Tarns and lakes dammed by talus are common. Headwalls of north facing cirques are cliffs some of which are 200 m high. South facing headwalls are less steep, and broken by benches and short cliff bands. In general the largest cirques and the longest troughs occur on the north side of ridges. Aretes and most summits are rounded (Figure 6.4).

Alpine troughs are the most common trough type in the study area. Trough lengths vary from 0.5 km for troughs heading in single south facing cirques on the Podunk Plateau, to 19 km for Sowaqua trough which is fed by several higher north facing cirques. Troughs which run into Sowaqua Trough from Mt Outram are truncated by the Sowaqua Trough, creating hanging valleys. This suggests these glaciers were confluent with Sowaqua glacier. All alpine troughs change to V-shaped, fluvial forms within several kilometres of their head.

The cols between the heads of Eleven Mile Creek and Eight Mile Creek and between Eleven Mile Creek and the unnamed valley to the east have been breached, either under an ice
Figure 6.4: Johnson Peak showing rounded summit and steep north side. Note protalus ramparts to right of Johnson Peak

sheet or by diffluence. One cirque at the head of Sowaqua Creek has been breached and replaced by a pass which drops abruptly into Snass Creek (Figure 6.2). It is likely that much of the original cirque was eroded under overriding ice. Ghost Pass is a broad (1 km) trough which could not have been carved by alpine glaciers because no peaks above the pass supported glaciers on the sides facing the pass.

Shallow, linear troughs cross the slope below Johnson Peak at an elevation of between 1770 and 1850 m (Figure 6.1). These features are 2-3 m wide with flat floors and small ridges (2-4 m high) on their downslope side. These have been interpreted as sackung by Dr. Wayne Savigny (personal
communication 1993). Pits dug in the centre of two of these reveal a 60 cm thick organic layer overlying well sorted sand, whereas soil pits in slopes above the troughs contain coarser textured material with abundant cobbles. This suggests that these troughs may have carried glacial meltwater. It is likely that meltwater exploited pre-existing troughs created by bedrock sagging near the margin of the glacier.

Snass Creek flows in a striking V-shaped valley, which begins abruptly at the edge of Podunk Plateau. This has been interpreted as a large subglacial meltwater channel (Figure 6.2).

In the middle reaches of Sowaqua Creek many tributaries have a distinct down valley deflection (Figure 6.2), rather than flowing directly downslope into the creek. The deflection could be a result of redirection around a retreating ice tongue.

Montigny Creek originates in a cirque incised below Podunk Plateau. The cirque has been breached and two steep walled gullies cut through the headwall. The creek is incised into a bedrock canyon, which resembles a subglacial meltwater channel for part of its length.

6.3 Depositional Landforms and Surficial Materials

All upper valley slopes are steep and dominated by bedrock, locally with a thin colluvial cover. Lower slopes are less steep and are mantled with till and colluvium. Valley floors and lower slopes of smaller creeks are filled with colluvial
cone and fan deposits which obscure any earlier glacial deposits. Sowaqua Creek valley contains extensive terrace deposits of glaciofluvial gravels overlying till (Map 3, Figure 6.2).

6.3.1 Till

Basal till is exposed in road cuts on moderately steep lower slopes throughout the study area. It fills hollows in bedrock and thickness varies from less than 1 m to over 10 m. Commonly a colluvial veneer or blanket derived from rock outcrops upslope overlies the till.

Till is invariably highly consolidated. Generally it is poorly drained and seepage is common over the surface. The matrix texture is sandy with occasional finer textured pockets. The texture may reflect the bedrock geology, with granitic areas having more sandy till than areas underlain by finer textured rocks. Clasts are subrounded to subangular, and largely pebbles with occasional cobbles. Throughout the study area granitic rocks are the most common clast type; mafic and felsic volcanics and sediments are minor components.

A mound of unsorted, moderately consolidated sandy rubble in Sowaqua Creek valley could be either ablation till or colluvium. The deposit resembles colluvium, but occurs on a terrace. Alternatively, it could be landslide debris transported by ice, then deposited as the ice melted. After deposition, this material was then partially covered by glaciofluvial outwash.
6.3.2 Moraines

Two well defined moraine ridges loop into adjacent southwest facing cirques at an elevation of 1600 m between Mt. Outram and Johnson Peak (Figure 6.2). The northernmost of these moraines was visited. This is the largest moraine at 500 m long and 5 to 10 m high (Figure 6.5). The second moraine is roughly 200 m long. Rounded to subrounded granodiorite boulders are common on the surface of the northern moraine. However a soil pit revealed that pebbles and cobbles are dominantly mafic volcancics with granodiorite and felsic volcanic rocks as lesser constituents. The cirque headwall is

Figure 6.5: Moraine damming lake on the south side of Mt. Outram. Note breached col behind the lake.
composed of granodiorite, so a cirque glacier should leave deposits with only this rock type. It is thus concluded from both their plan-form and composition that these are moraines which were deposited by active ice in the Nicolum-Sumallo trough rather than by cirque glaciers.

A broad low ridge in a tributary valley on the south side of Sowaqua Creek at 1200 m is interpreted as a lateral moraine deposited by the Sowaqua Glacier. The ridge has an irregular hummocky surface with scattered granodiorite boulders. A soil pit reveals it is composed of silty sand with subrounded volcanic and granodiorite clasts.

6.3.3 Terraces, Glaciofluvial and Glaciolacustrine Deposits

Extensive valley fill deposits are present in Sowaqua valley, where terraces of glaciofluvial outwash gravels overlie till. The thickest of these deposits occurs at the mouth of Montigny Creek where 30 m of glaciofluvial gravels overlying over 5 m of till are exposed along the path of a slide into the creek (Figure 6.6 section 1). The surface of this terrace is at an elevation of 850 m, 60 m higher than adjacent terraces. Upstream of Montigny Creek the terraces are continuous for 5.5 km on both sides of Sowaqua Creek, and have a surface gradient of 6% (Map 3).

Terrace deposits along Sowaqua Creek are crudely stratified sandy gravels, typical of ice proximal outwash deposits (Brodzikowski and van Loon 1991; Miall 1983). Near the upstream limit of the terraces deposits become more
Figure 6.6: Selected stratigraphic sections, see figure 6.2 for locations.
irregular (Figure 6.7); bedding is disrupted and lenses of till and glaciolacustrine sediments are present (Figure 6.6 section 2). Glaciolacustrine material is up to 5 m thick and composed of massive to laminated silt. These deposits are interpreted as ice contact. Further upstream the valley of Sowaqua creek contains only basal till.

Figure 6.7: Road cut through upper end of glaciofluvial terrace in Sowaqua Creek valley, near section 2, exposing irregular bedding. Till and glaciolacustrine silts are exposed 50 m downstream.

Debris from the Hope Slide, a large rock avalanche which occurred in 1965, fills much of the Nicolum-Sumallo trough (Map 3, Figure 6.2). Between Eight Mile Creek and the Hope Slide is a large kettled terrace (Figure 6.2). A road cut
through this terrace reveals a 40 m thick section of glaciofluvial gravels and sands (Figure 6.6 section 3 and 4, Figure 6.8). The deposit is planar bedded, and clast size varies from fine sand to boulders 30 cm in diameter. Well sorted beds are from 10 cm to 2 m thick. No structures indicative of flow direction were found. A pocket of laminated silts, 1 m thick also occurs in this section. This terrace is likely a kame terrace deposited adjacent to ice in the Nicolum Valley.

Figure 6.8: Road cut through kame terrace above Nicolum Creek at section 3, exposing over 30 m of planar bedded sands and gravels.

Two to three kilometres upstream from its junction with Skagit River, a wider section of Snass Creek valley contains a
hummocky glaciofluvial terrace, similar in appearance to the kame terrace near Eight Mile Creek. This was likely deposited adjacent to ice in the lower Snass valley.

6.4 Rock Weathering

It is impossible to compare weathering of rock in valleys with that on the highest summits because they exhibit different lithologies and styles of weathering. The highest summits in the study area, 2440 m Mt Outram and 2160 m Macleod Peak, are composed of fractured volcanic rocks. On Mt. Outram volcanic rocks outcrop above 2000 m. Here rock is severely fractured and nivation hollows in talus are pronounced. Outcrops of volcanic rocks are rare in valley bottoms and, where present are also severely fractured.

Granodiorite weathering in the Mt. Outram study area showed more variability than the other areas. Most rock outcrops are relatively smooth with minor differential weathering of individual grains. Round to subrounded boulders, displaying a similar degree of weathering are common on ridge tops. Six sites on these smooth outcrops and boulders, ranging in elevation from 1400 m to 1900 m, were sampled using the Schmidt hammer. A range in mean Schmidt hammer readings of 40 to 45 with a standard deviation of 6 to 10 was obtained (Figure 6.9, Appendix 3).

Above 1800 m on Johnson Peak granodiorite outcrops are rounded with crumbly flaky surfaces coated with several millimetres of grit sized fragments of weathered granodiorite
(grus). Grus forms thick aprons at the base of outcrops. Schmidt hammer readings were lower here than elsewhere with means of five samples ranging from 23 to 30 and standard deviation of 5 to 6 (Figure 6.9). No large boulders are present in this area.

Between 1750 m and 2085 m on ridges between Mt. Outram and Macleod Peak granodiorite is knobby and pitted. Weathering pits are circular, generally less than 10 cm wide and may be either shallow (1-2 cm) with sloping sides or up to 10 cm deep with vertical walls. Knobs are up to 10 cm high. One small 30 cm diameter boulder is attached to the underlying slab.

![Figure 6.9: Plot of mean Schmidt Hammer readings with elevation. Bars represent +/- 2 standard deviations of the mean.](image-url)
at its base, so it appears to have weathered in place. No
flaking or grus formation is present in this area. Schmidt
hammer readings were similar to unpitted granodiorite in other
localities with the means of six samples ranging from 40 to 45
and standard deviation of 7 to 11 (Figure 6.9 Appendix 3).

The type of knobby, pitted weathering in the Outram area
is very similar in appearance to that observed by the writer
in the Lizzie Lake area of the Coast Mountains. (Lizzie Lake
is located in a large alpine area to the east of Lillooet
Lake, near the head of the Stein River.) Several summits
around Lizzie Lake have slabs with knobs and pits which are
the same size as those near Mt. Outram. Many of the Lizzie
Lake slabs also have fresh striations on the same outcrop, so
the knobs and pits must have formed after Fraser Glaciation.

Twidale and Corbin (1963) have found weathering pits of
10 cm in diameter on surfaces in Australia known to be at most
a few thousand years old. Sorensen (1988) has observed thin
veneers of grus, similar to that on Mt. Johnson, on surfaces
in southeastern Norway which emerged from the sea only one
thousand years ago. Sugden and Watts (1977) argue that large
weathering pits, greater than 75 cm wide and 20 cm deep, on
Baffin Island likely formed prior to the last glaciation and
were preserved under cold based ice. Ryder (personal
communication 1995) observed that weathering pits in the
Okanagan Range of the Cascade Mountains were more than 1 m in
diameter on surfaces above elevations reached by the
Cordilleran Ice Sheet during Fraser Glaciation. Although these
studies are from a variety of environments, they do suggest that the depth of grus and the size of pits observed in the Mt. Outram study area could have formed since the last glaciation.

The plot of mean Schmidt hammer readings versus elevation (Figure 6.9) reveals there is no trend between mean values and elevation. Means of samples on normal and pitted weathered surfaces are somewhat clustered. An exception is a sample at 1600 m with normal weathering which has an anomalously low value. This surface was somewhat rougher than other sites, with more individual grains protruding. The lack of a sharp drop in mean values with elevation suggests that factors other than level of glaciation are responsible for the variability in the readings.

Some of the variability in weathering could be a function of variations in snow pack duration and moisture. Benedict (1993) found that weathering of granitic rocks is fastest where snow cover is thin to moderate and meltout occurs early, so there are more freeze thaw cycles. Hall (1993) found the opposite, that the maximum chemical weathering occurred under late lying snow packs. A second possibility could be minor lithologic variations within the granodiorite, particularly the percentage of biotite (see Bustin and Mathews 1978).

Soil development varies with rock weathering. On Johnson Peak in the vicinity of the flaky granodiorite, soil is composed of grus, with very little fine grained matrix; LFH and A horizons are up to 20 cm thick. This is very different
from soil developed on the moraines and in areas with either smooth or pitted granodiorite. In these areas soil has a silty sand matrix and granodiorite clasts to cobble size, typical of soil formed on till. LFH and A horizons are generally 5 to 10 cm thick. This and a general lack of boulders suggests that there is little or no till deposition in the area of flaky slabs near Johnson Peak, or thick grus was deposited over till after deglaciation.

The flaky surfaces of Johnson Peak have lower Schmidt hammer readings than any other sites in the study areas. These values are similar to those in areas not affected by Fraser Glaciation in the Okanagan Range (J.M. Ryder 1989, personal communication 1995). However, striations occur at a similar elevation roughly 3 km to the east. Ridges 2 km northeast have erratics at the same elevation as the flaky slabs. The location of Johnson Peak and its relatively low elevation (1950 m) makes it unlikely that it was not overtopped by Cordilleran Ice, thus the anomalously low values are difficult to account for. They may indicate that little erosion was accomplished by the Cordilleran Ice Sheet due to short duration, thin ice, or little basal movement.

6.5 Glacial History

6.5.1 Valley Glacier Phase

The presence of well defined erosional landforms of alpine glaciation again indicates that alpine and valley glaciers
formed around higher summits during the early stages of glaciation (Figure 6.10). The extent of these glaciers is reconstructed entirely from erosional topography.

Figure 6.10: Early Fraser Glaciation alpine glaciers.
It is assumed that glaciers extended down valleys to U-V transitions, which are well defined in this area. There is a weak relation between cirque area and trough length (Figure 6.11) (See Chapter 3 for a description of the method used to measure cirque area and glacier length) and north facing cirques are larger and their troughs longer than those facing south. These facts lend support for the U-V transition marking the terminus of valley glaciers (Flint 1971; Goldthwait 1970).

This phase is responsible for dominant topographic forms such as cirques and alpine troughs.

Figure 6.11: Plot of cirque area versus trough length.
6.5.2 Ice Sheet Stage

At the Fraser Glaciation maximum all summits were likely overtopped by ice. Ridge crests are generally broad and rounded and erratics are common on ridges below Mt. Outram and MacLeod Peak. The presence of breached cols, such as at the head of Eight, Eleven Mile, and Sowaqua Creeks, and broad passes such as Ghost Pass is further evidence for overriding Cordilleran ice. Nicolum-Sumallo pass is a deep steep-walled through-trough fed almost entirely by valleys that become V-shaped before entering it. It must therefore have been carved largely by Cordilleran Ice.

Rock weathering is variable in the Mt. Outram Study area, but no granodiorite outcrops are weathered strongly enough to indicate they could not have been glaciated. The enhanced weathering on Johnson Peak and the ridge between Mt. Outram and Macleod Peak may indicate that the ice sheet covered these ridges for only a short period of time.

Both Mt. Outram and MacLeod Peak are craggy summits composed of highly fractured rock. There is no strong evidence that they were overtopped by Cordilleran Ice. However at the International Border 32 km south of the study area the ice surface was higher than 2600 m (Waitt and Thorson 1983), 200 m higher than Mt. Outram. As Mt. Outram is further from the periphery of the ice sheet it is likely that it was overtopped, at least briefly.

Few ice flow indicators are present in this area. Grooves in meadows below Mt. Outram have an orientation of between 78°
and 83°. A single poorly preserved groove west of Mt. Davis has an orientation of 140°. In both cases the grooves are roughly parallel to adjacent valleys suggesting that when they formed flow was controlled by local topography. It is likely that at the Fraser Maximum ice from the Fraser River flowed eastward through Nicolum-Sumallo Trough (Evans 1990).

6.5.3 Deglaciation

Evidence of deglaciation is patchy. In the Nicolum-Sumallo trough two moraines and one kame terrace provide two ice levels. In Sowaqua Creek one lateral moraine and one kame terrace are the only ice marginal features observed. As a result, maps of deglaciation are not precise, each map is based on only one or two features. They have been drawn to demonstrate a possible sequence of deglaciation which is consistent with the available evidence.

The highest ice level recorded is 1680 m at the elevation of the lateral moraines west of Mt. Outram (Figure 6.12). The moraines were likely deposited during a brief resurgence of ice in this valley. At this time ice in the pass was over 1000 m thick and likely was still active and flowing east. Alpine glaciers were present in higher north facing cirques.

The deposits in Sowaqua Creek are most consistent with up-valley retreat, possibly with minor stagnation of the glacier snout; glaciofluvial terraces grade upstream into ice contact deposits and finally till. This glacier was fed almost entirely from high north-facing cirques, many of which appear
Figure 6.12: An ice level near the start of deglaciation, based on moraines below Mt. Outram.
relatively fresh. It is therefore possible that glaciers remained active in this valley throughout deglaciation and retreated by a combination of thinning of active ice and frontal retreat, with local stagnation of the glacier snout. The ice level shown in figure 6.13 is based on a single lateral moraine found at 1220 m on the south side of Sowaqua Creek. The map was drawn assuming similar rates of thinning throughout the study area. This may not be a valid assumption; ice in Nicolum-Summallo trough may have either been largely cut off from its source areas or it may have still been fed by a lobe of active ice from the Fraser Valley. In either case it is unlikely that it retreated at the same rate as Sowaqua Glacier, which was fed by local cirque glaciers.

As the Sowaqua Glacier continued to recede it would have thinned and separated into discrete valley glaciers, which continued to retreat up valley. Figure 6.14 is drawn at the time of kame terrace formation in Sowaqua and Nicolum Creeks. No evidence of terminal positions of other valley glaciers was found, so the glaciers were drawn assuming a similar relation of cirque size to glacier length as was found for the alpine glacier phase (Figure 6.11).

During the final stages of retreat, ice in the Nicolum-Sumallo trough may have stagnated, while meltwater deposited kame terraces at the mouth of Eleven Mile Creek and in Snass Creek (Figure 6.14). The kame terrace at Eleven Mile Creek is approximately the same elevation as the pass between Nicolum and Sumallo creeks. It is therefore possible that meltwater
Figure 6.13: A later ice level, based on a lateral moraine at 1220 m in Sowaqua Creek.
Figure 6.14: An ice level near the end of deglaciation, based on the kame terraces in Nicolum and Sowaqua Creeks.
drained eastward into the Skagit basin rather than to the Fraser River at this time. Eight Mile Creek heads in a north facing cirque which is lower and less fresh looking than north facing cirques to the east. It does not appear to have held a large active glacier throughout deglaciation, but small glaciers may have existed on surrounding peaks and contributed material to the kame terrace in Nicolum valley via a meltwater channel which ran along the margin of the ice in Nicolum valley (Figure 6.14). This channel may also have carried material to the terrace from active ice behind the stagnant snout in Nicolum Creek.

Much of Podunk Plateau, which is relatively high, likely remained ice covered until relatively late during deglaciation. Ice here downwasted and stagnated, depositing hummocky till. The heads of both Montigny and Snass Creeks are at the plateau edge so it is likely they carried meltwater from the plateau. The large glaciofluvial terrace at the mouth of Montigny Creek, which is 60 m higher than adjacent terraces in Sowaqua Creek may have been formed by meltwater from the plateau. Similarly plateau meltwater may have contributed material to the glaciofluvial terrace in Snass Creek.

Extensive downwasting or downvalley retreat should produce considerable damming of upper valleys and tributaries, as these would become ice free first. However lacustrine deposits are rare throughout the study area. This suggests that subglacial drainage was common, or glaciolacustrine
deposits have been eroded or buried by colluvium since deglaciation.

6.6 Neoglaciation

Neoglacial (Little Ice Age) activity was restricted to north facing cirques. The dominant activity appears to have been the formation of protalus ramparts. In addition small glaciers may have existed at the base of steep slopes where avalanches fed the accumulation zones. A remnant of one such glacier still exists on the north side of Mt. Outram.
CHAPTER 7 DISCUSSION

7.1 Glaciation of the Northern Cascade Mountains

A consistent pattern of glaciation has been observed in all three study areas. At the start of Fraser Glaciation alpine glaciers formed around higher summits, dominantly on their north sides (Figure 7.1). They likely advanced to a roughly stable position marked by a transition from trough to V shaped valleys. A plot of cirque area versus trough length for both the Outram and Stoyoma areas reveals a consistent relation in both areas (Figure 7.2).

The U-shaped portions of valleys are between 100 and 300 m wider than the V-shaped portions. (As measured across valley flats, roughly 1 km above and below the transition zone). Cirque floors are 100 to 500 m lower than adjacent ridge tops, and cols have been lowered by up to 400 m assuming that a continuous, roughly flat ridge previously existed. Using Andrews (1975) rates of erosion (see chapter 3) as a rough guide the length of time required to accomplish this much erosion can be calculated. At the slowest quoted rate of 400mm/1000 years, up to 1 million years would be required to accomplish the observed lowering of cols and the erosion of the largest cirques. Even the fastest quoted rate of 5,000 mm/1000 years would require 80,000 years. Between 40,000 and 125,000 years would be required for the observed widening of the troughs. As the total duration of Fraser Glaciation was roughly 20,000 years, it is clear, even from very rough
Figure 7.1: Glaciation model for the northern Cascade Mountains. a). Early Fraser Glaciation alpine glaciers. b). Early advance of the Cordilleran Ice Sheet up north trending valleys. c). At the Fraser Maximum all summits are overtopped by an external ice sheet. d). During deglaciation ice retreats down valleys without a local ice source or those of lower and south facing cirques which are below the local snowline. Higher north facing cirques may remain above snowline glaciers will remain active and retreat up valley.
Figure 7.2: Plot of Outram and Stoyoma cirque areas versus trough length.

calculations, that much of the erosion in all study areas, including the formation of the large troughs and cirques, predates Fraser Glaciation. The shortest time of 80,000 years for the erosion of the large cirques is similar to the 100,000 year cycle for the last major glaciation (Sugden and John 1976). However it is unlikely that this erosion rate would have been sustained for the entire glaciation because for part of this time cirque glaciers would likely have been very small. Thus most alpine forms must be a result of erosion
during the alpine stage of several major glacial cycles. Therefore the alpine and intense alpine phases must have been repeated during each glaciation.

In Washington near the periphery of the ice sheet, alpine glaciers retreated before the ice sheet moved into the area. In the study areas there is no evidence to suggest that alpine glaciers retreated before the advance of the Cordilleran Ice Sheet. No more than one till was found at any location. It is likely that alpine glaciers coalesced with or were overridden by the Cordilleran Ice Sheet.

Toward the end of the intense alpine phase (Davis and Mathews 1944), glaciers from the Coast Mountains began to advance into surrounding lowlands. The Cascade Mountains would have posed a significant barrier to this ice. South of Lytton, Coast Mountain ice likely would have been diverted south, down the Fraser River. Lobes would then have advanced up major Cascade valleys which drained into the Fraser, particularly those which drained to the northwest such as the Anderson River, Nicolum Creek and Silverhope Creek (Figure 7.3). North of Lytton, Coast Mountain ice would have moved eastward and southeastward up the Nicola valley. A major depression on the east side of the Cascade Mountains, extends from the Thompson River to the Similkameen River (Figure 7.3), and likely served as a conduit for Coast Mountain ice. While advancing down this depression, lobes of ice likely would have blocked and flowed up Cascade valleys that drained to the northeast, such as
Figure 7.3: Map of the Cascade Mountains and surrounding areas depicting probable routes for the early advance of ice from the Coast Mountains into the Cascade Mountains.
Spius Creek, Coldwater River, Coquihalla River and the uppermost Similkameen River.

Thus northward draining valleys near both the east and west margins of the Cascade Mountains were likely dammed by lobes of Cordilleran ice relatively early in Fraser Glaciation, while Cascade alpine glaciers were still confined to upper valleys. Continued advance of both alpine glaciers and the ice sheet would have caused them to coalesce and override the fine textured glaciolacustrine sediments deposited in front of the advancing ice sheet. These sediments would have been incorporated into till, resulting in the fine textured till of these valleys.

In the Stoyoma area drumlins indicate flow to the southwest on the plateau. As these are not at high levels, they may represent ice flow during early Fraser Glaciation. Flow to the southwest on the Spius Plateau is consistent with a lobe of Cordilleran Ice advancing up Spius Creek from the northeast, and expanding over the plateau area.

Following mountain glaciation, an ice sheet covered all areas including the highest summits (figure 7.1c). Ice flow north of the Stoyoma area at the Fraser maximum was generally to the south (Ryder 1981). In the Anderson River area fluting on summits and ridge tops indicates flow at the Fraser maximum was also to the south, oblique to local valleys. This suggests that in this area ice thickness was considerably more than relief, that is much greater than 2000 m. Grooves near ridge tops in the Outram Area are roughly parallel to local valleys.
suggesting that in this area ice flow was controlled by local
topography, and ice may not have been much thicker than
relief. Thus the ice surface may not have been much higher
than 2400 m Mt. Outram.

Where the ice sheet was flowing roughly parallel to local
valleys, these would have continued to be widened and
deepened. In the northern two areas where the ice sheet was
flowing oblique to valleys at the Fraser maximum, erosion of
valleys would have ceased while the ice was at its maximum
thickness.

At the start of deglaciation the ice sheet remained
active and thinned until higher peaks became ice free.
Snowline had risen above lower elevation and south facing
cirques before they emerged from under the ice sheet, thus
they were no longer able to support alpine glaciers. However
some higher north facing cirques in the Mt. Outram area, and
likely also in the Anderson River area, remained above the
snowline and were able to support glaciers throughout
deglaciation. As a result, ice generally remained active and
thinned while retreating back up these valleys. Some
stagnation of glacier snouts likely occurred during frontal
retreat, allowing the deposition of kame terraces in valleys
such as Sowaqua Creek.

Down-valley retreat of active lobes of Cordilleran ice
appears to have been the dominant mode of retreat throughout
the study areas. The main evidence for this is the moraines
located in the Stoyoma and Outram areas and the general lack
of fresh features in most cirques. Once the higher summits became ice free the Cordilleran Ice Sheet retreated away from the mountains, in roughly the reverse direction of advance. Early during deglaciation, there was a brief resurgence when the ice sheet deposited the moraines in several south facing and low elevation cirques. There are few deposits indicative of stagnation so it is likely that subsequent retreat was continuous. Lateral meltwater channels most often form when ice is active (Fulton 1967), thus the channels in the Stoyoma area are further evidence of active ice.

There is little evidence in any study area to suggest how meltwater escaped from valleys in which active ice retreated down-valley. If ice is active, the ice surface slope should control subglacial flow of meltwater (Shreve 1972). This would prevent down-valley flow, and result in lakes being ponded up-valley from the retreating ice margin. Only the Anderson River study area contains significant glaciolacustrine deposits in valleys that should have been dammed by downwasting ice (see Deglaciation maps). There is no indication of how these lakes drained.

There are several possible explanations for the problem of drainage in valleys which underwent down-valley retreat. 1. The ice surface slope could have been very close to 0° allowing bedrock topography to dominate subglacial flow. This may have permitted these valleys to drain subglacially. Bedrock canyons in Miner's and Montigny Creeks could have originated as subglacial tunnel valleys, and are possible
evidence of subglacial drainage. 2. There could have been damming in many valleys but no evidence of lake sediments was found due to erosion, or mixing of thin glaciolacustrine deposits with till by periglacial processes, or insufficient field checking. 3. The lobes of Cordilleran ice were stagnant rather than active, but carried little debris so few deposits remain. In this case meltwater could escape down-valley either subglacially or supraglacially. 4. Lateral drainage was present and no channels were observed. This would require lakes to be ponded to a sufficient depth to allow flow around the up-valley sloping ice tongues. More detailed field work is required in valleys which should have been deglaciated by down valley retreat in order to determine the best explanation.

Neoglacial activity throughout the Cascades was restricted to north facing slopes that were fed by avalanches (and to large volcanoes in Washington). No extensive glaciers, as were present in the Coast Mountains existed in the Cascades. This may be due to a drier climate without enough precipitation, and a lack of sufficiently high large alpine areas for ice accumulation.

7.2 Pattern of Glaciation in areas with Mountain and Ice Sheet glaciation

The broad details of the Cascades glaciation appear to be in common with other areas that had both local and external ice sources, such as the Washington Cascades (Waitt 1977, 1975; Waitt and Thorson 1983), the Presidential Range in New
Hampshire (Goldthwait 1970; Bradley 1981), the Green Mountains in Vermont (Wagner 1970; Connally 1982), mountains of west-central Maine (Borns and Calkin 1977) and the Insular Mountains on northern Vancouver Island (Howes 1981, 1983). The early glacial stages of local alpine glaciation followed by overtopping by a regional ice sheet occurred in all areas. In all areas, deglaciation was controlled by local topography, elevation of snowline and distance from the edge of the ice sheet. Thus deglaciation of areas that experienced mixed mountain and ice sheet glaciation followed a complex pattern that depended strongly on local conditions. It is thus more difficult to predict the pattern of deglaciation in these areas than in those where only one style of glaciation, either mountain or ice sheet, occurred.

7.3 Application of Glaciation Models to Terrain Mapping

The manner in which glaciers move into and out of an area largely determines the resulting types of landforms and surficial materials. Knowledge of glacial history can allow the prediction of deposits, particularly those on lower slopes and valley bottom. Commonly, valleys contain several layers of material, which have been deposited throughout the glacial and post glacial period. The uppermost unit is likely to determine the surface expression and lower units may be difficult to recognize.

If valleys have been dammed at any time in the glacial cycle fine textured sediments may be present. Typically these
will be overlain by coarser material and will have no surface expression. Because fine textured sediments have a strong influence on stability, it is important to be able to predict their occurrence.

Each of the three styles of glaciation, that is mountain, ice sheet or mixed, produces a different suite of landforms and deposits, and so knowledge of the glaciation style should be helpful in predicting the expected materials in valley fill.

7.3.1 Examples from the northern Cascade Mountains

In the Cascade Mountains the style of glaciation was mixed mountain and ice sheet. Under this model, at the start of glaciation, alpine glaciers began to expand down valleys, particularly those which face north. Later during the advance phase, lobes of the Cordilleran Ice Sheet began to advance up many valleys that drained out of the Cascades. These lobes eventually coalesced with the alpine glaciers, and ice eventually overtopped all summits. Lower and south facing valleys were deglaciated by down-valley retreat of active lobes of Cordilleran ice while higher, north facing valleys underwent frontal retreat of local valley glaciers.

As a consequence of this sequence of events, valleys that drain toward The direction from which the Cordilleran Ice Sheet was advancing, may have had a period during advance when ice was moving up valleys at the same time as alpine glaciers were advancing down valleys. While the middle portion of the
valley was ice free it would have been dammed by the ice sheet and the resulting glaciolacustrine deposits overridden as ice continued to advance.

The Mt. Stoyoma area contains clay-rich till in the lower parts of several east and northeast facing valleys (see figure 4.4), consistent with damming by overriding Cordilleran ice. The presence of this till would not have been predicted on the basis of lithology as the local rock is granitic. If the area had been glaciated primarily by local alpine glaciers, valleys would not have been dammed and till would likely be much coarser textured. Clay-rich till near the junction of the North and South Forks of the Anderson River is also likely related to reworking of fine-textured valley fill by ice advancing up-valley. The local bedrock is fine grained, but the till matrix texture is very similar to that of glaciolacustrine sediments up valley.

In the Cascades, valleys which did not support significant late stage alpine glaciers generally head in broad cols, highly eroded breached cirques, or have no local source area, and a few contain ice sheet moraines. These valleys were deglaciated by down-valley retreat of active ice. Meltwater flowing down these valleys should have been blocked by the active ice, unless the ice surface slope was very low. The Anderson River study area is the only one which contains relatively extensive glaciolacustrine material in valleys which should have been dammed by down-valley retreat. These are not thick deposits, but occur throughout all valley
bottoms. Damming of the Anderson River likely was caused by a lobe of active ice from the Coast Mountains in the lower part of the Anderson River valley. In the other two areas no fine textured sediments were observed in valleys which should have been dammed during deglaciation. Nicolum valley contains a large kame terrace that was deposited next to stagnant ice.

It is difficult to distinguish on airphotos valleys that contain glaciolacustrine sediments from those that do not. Thin but extensive glaciolacustrine deposits in Anderson River are not visible on air photos. In some cases, such as Miner's Creek, an incised meltwater channel is present in valleys without glaciolacustrine sediments suggesting that subglacial drainage may have occurred (either due to stagnant ice or a low ice surface slope). Other valleys, such as Eleven Mile Creek in the Outram area, have neither an incised stream nor glaciolacustrine deposits. Thus in the study areas, the model does not allow the accurate prediction of the location of glaciolacustrine deposits. However it does suggest that valleys which head in truncated cirques, contain ice sheet moraines and lack obvious subglacial meltwater channels, should be suspected of containing glaciolacustrine sediment and targeted for more detailed field work, particularly if identification of this material is a priority of terrain mapping.

Valleys in which frontal retreat of local glaciers was dominant generally have glaciofluvial outwash overlying till in the valley fill, and few fine-textured sediments are
present. These valleys can be recognized because they head in
greater looking cirques than other valleys and by the presence
of glaciofluvial terraces.

Through troughs which originate on the west side of the
Cascade Mountains, such as South Spius Creek may have been
conduits for meltwater from the Coast Mountains. These valleys
contain thick deposits of glaciofluvial material which cannot
be explained by local features. Some valleys, where ice
retreated down-valley to the east, may also have been dammed.
In this case glaciolacustrine silts may underlie the
glaciofluvial material in the lower reaches of the valley.

Till thickness is not well predicted by the model. There
is some indication that thick deposits of till may be common
in northeast and northwest trending valleys, in which till was
deposited by ice advancing up valley early during glaciation.
These valleys are generally transverse to ice flow at the
maximum, allowing the preservation of this till.

7.3.2 Conclusions

All areas which have been glaciated by mixed mountain and ice
sheet glaciation have rounded and subdued alpine features such
as cirques and horns. Valley fill stratigraphy varies, and
depends largely on the style and pattern of deglaciation. The
style of deglaciation may be inferred from air photo analysis,
combined with an understanding of regional ice flow
directions. This allows some broad generalizations to be made
about probable valley fill stratigraphy.
The model of glaciation developed for the Cascade Mountains cannot accurately predict valley bottom deposits in all study areas. The model does however predict which valleys are most likely to have been dammed during advance and retreat, and thus where fine-textured till and glaciolacustrine sediments are most likely to be located. Not all of these valleys will contain fine-textured sediment, but the model does identify where more detailed field checking may be worthwhile to determine the extent of fine-textured sediment, particularly if stability issues are important.

The model also indicates which valleys are likely to have undergone frontal retreat and will thus contain glaciofluvial material overlying till. In addition, valleys which head in low passes between the Fraser River and inland valleys are likely to contain outwash deposits from Cordilleran Ice.

The model of mixed mountain and ice sheet glaciation for the Cascade Mountains is useful to terrain mapping, but is not completely successful in predicting valley fill deposits. More work in nearby areas is desirable to further refine and test its applicability.
REFERENCES


Johansson, Peter (1988) Deglaciation pattern and ice-dammed lakes along the Saariselka mountain range in northeastern Finland. Boreas, 17: 541-552.


APPENDIX 1

This appendix contains histograms of Schmidt hammer data for the Mt. Stoyoma study area. See Map 1 for the location of sites.

Figure A1.1: Schmidt hammer results Site 1, at 1920m.
Figure A1.2: Schmidt hammer results Site 2, at 2050m.

Figure A1.3: Schmidt hammer results Site 3, at 2140m.
Figure A1.4: Schmidt hammer results Site 4, at 2270m.

Figure A1.5: Schmidt hammer results Site 6, at 1850m.
Figure A1.6: Schmidt hammer results Site 7, at 1930m.

Figure A1.7: Schmidt hammer results Site 9, at 1980m.
Figure A1.8: Schmidt hammer results Site 10, at 2040m.

Figure A1.9: Schmidt hammer results Site 11, at 2100m.
Figure A1.10: Schmidt hammer results Site 12, at 2230m.

Figure A1.11: Schmidt hammer results Site 13, at 1840m.
APPENDIX 2

This appendix contains histograms of Schmidt hammer data for the Anderson River study area. See Map 2 for the location of sites.

Figure A2.1: Schmidt hammer results Site 1, at 1210m.
Figure A2.2: Schmidt hammer results Site 2, at 1500m.

Figure A2.3: Schmidt hammer results Site 3, at 1640m.
Figure A2.4: Schmidt hammer results Site 4, at 1620m.

Figure A2.5: Schmidt hammer results Site 5, at 1930m.
Figure A2.6: Schmidt hammer results Site 6, at 1980m.

Figure A2.7: Schmidt hammer results Site 7, at 1770m.
Figure A2.8: Schmidt hammer results Site 8, at 1710m.

Figure A2.9: Schmidt hammer results Site 9, at 1850m.
Figure A2.10: Schmidt hammer results Site 10, at 1930m.

Figure A2.11: Schmidt hammer results Site 11, at 2060m
APPENDIX 3

This appendix contains histograms of Schmidt hammer data for the Mt. Outram study area. See Map 3 for the location of sites.

![Histogram of Schmidt hammer data](image)

**Figure A3.1:** Schmidt hammer results Site 1, at 1750m, on normally weathered slabs.
Figure A3.2: Schmidt hammer results Site 2, at 1940m, on normally weathered slabs.

Figure A3.3: Schmidt hammer results Site 4, at 1940m, on grussified slabs.
Figure A3.4: Schmidt hammer results Site 5, at 1820m, on normally weathered slabs.

Figure A3.5: Schmidt hammer results Site 7, at 1770m, on normally weathered slabs.
Figure A3.6: Schmidt hammer results Site 8, at 1600m, on normally weathered boulders in moraine at Outram Lake.

Figure A3.7: Schmidt hammer results Site 40, at 1425m, on normally weathered slabs.
Figure A3.8: Schmidt hammer results Site 45, at 1843m, on normally weathered slabs.

Figure A3.9: Schmidt hammer results Site 46, at 1863m, on grussified slabs.
Figure A3.10: Schmidt hammer results Site 47, at 1848m, on grussified slabs.

Figure A3.11: Schmidt hammer results Site 48, at 1813m, on grussified slabs
Figure A3.12: Schmidt hammer results Site 49, at 1933m, on grussified slabs

Figure A3.13: Schmidt hammer results Site 50, at 3513m, on pitted slabs
Figure A3.14: Schmidt hammer results Site 51, at 2085m, on pitted slabs

Figure A3.15: Schmidt hammer results Site 52, at 1935m, on pitted slabs
Figure A3.16: Schmidt hammer results Site 53, at 1935m, on pitted slabs

Figure A3.17: Schmidt hammer results Site 54, at 1855m, on pitted slabs
Figure A3.18: Schmidt hammer results Site 58, at 1813m, on normally weathered slabs.

mean 34.5
std 6.3

Figure A3.19: Schmidt hammer results Site 59, at 1495m, on normally weathered slabs.

mean 41.0
std 7.7