RECONSTRUCTION OF THE GLACIAL HISTORY OF THE COLUMBIA ICEFIELD, ALBERTA

by

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ABSTRACT

Classically, glacier fluctuations have been used as primary indicators of climate change. The histories of six glaciers at the Columbia Icefield (Castleguard, Columbia, Kitchener, Manitoba, Saskatchewan and Stutfield) were reconstructed using dendroglaciological techniques. Overridden trees document early Little Ice Age (LIA) advances after 1271 at Stutfield and 1474 at Manitoba Glaciers. Athabasca, Dome, Saskatchewan and Castleguard Glaciers reached their maxima in the mid 19^{th} century when all glaciers advanced to close to their LIA maximum positions. Columbia/Manitoba, Kitchener and Stutfield Glaciers record maxima ca. 1739, 1713 and 1758, respectively. Two sites have evidence for late 18^{th} and early 19^{th} century advances. Glacial recession has been dominant in the 20^{th} century with maximum rates in the 1940's-60's and 1980's-90's. Frontal recession rates decreased in the1960's and 70's and Kitchener and Columbia Glaciers advanced. Periods of LIA glacier advance were broadly synchronous between glaciers but the maximum positions were not achieved simultaneously.

Keywords: Glacier, Dendroglaciology, Little Ice Age, Columbia Icefield, Canadian Rocky Mountains

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CHAPTER I

INTRODUCTION

1.1 Introduction

In many alpine areas, the study of recent glacial fluctuations has been used to define basic ideas about recent climate history (Matthes 1939, Porter & Denton 1967, Grove 1988). In the Canadian Rocky Mountains, existing evidence indicates a pattern of progressively more extensive glacier advances during the Neoglacial (the last 5,000 years) culminating in the Little Ice Age[1] (LIA) maximum of the 18th and 19th centuries (Heusser 1956, Osborn & Luckman 1988, Luckman 1995, 1996a). As the most recent glacier advance was the most extensive, morphological and stratigraphic evidence documenting earlier glacier margins has been largely eliminated. Consequently, the bulk of the information concerning glacier responses to climate has been focused on reconstructing LIA events at valley and cirque glaciers. Dating moraine formation during the LIA in the Canadian Rockies indicates at least two general periods of moraine building; one in the early 1700's and the other during the mid 1800's (Luckman 1996a). Within this general chronology, there are differences in the timing of events at different glaciers, but a broad distribution of dates for a particular event is expected due to variations in the responses of individual glaciers to changes in mass balance (e.g. Oerlemans 1989, Smith *et al.* 1995). However, no studies have specifically targeted the study of outlet glaciers receiving nourishment from a common source. Therefore, it is uncertain whether outlet glaciers from icefields[2] react similarly to valley glaciers in their response to climate change. A common ice source would be expected to have a moderating response on the outlet glaciers, possibly resulting in a longer lag between a change in the regional climate and the

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[1]The term the Little Ice Age was initially proposed by Matthes (1939) who defined it as "an epoch of renewed but moderate glaciation which followed the warmest part of the Holocene" that spanned a period of over 4,000 years. It has been subsequently used to describe a synchronous, world wide period of sustained cooler conditions lasting approximately 400-500 years, ending in the late 19th-early 20th century which frequently was the most extensive episode of alpine glacier activity (Denton & Karlén 1973, Grove 1988, Bradley & Jones 1993).

[2]An icefield can be defined as an approximately level area of ice lacking the characteristic domelike shape of an ice cap and whose flow is strongly influenced by the underlying topography. Icefields form on gentle topography at altitudes sufficient for ice accumulation (Sugden & John 1976, Benn & Evans 1998).

associated glacier response than that observed in adjacent valley glaciers. In such a small area, the climate forcing/mass balance changes should be synchronous, although differences in morphology, topography and aspect may result in slightly different histories for the outlet glaciers.

Results from studies at Athabasca and Dome Glaciers, the two most accessible outlets of the Columbia Icefield (Luckman 1988) suggest that precise chronologies might be constructed for various other outlets of the Icefield and establish whether these glacial events were synchronous. Precise evidence at Athabasca and Dome Glaciers indicates synchronous advances and it is important to determine whether this is a local or Icefield-wide event. In addition, several of the outlet glaciers have not been studied in detail. Therefore, this study will endeavour to complete this task by reconstructing the glacial histories of six additional glaciers from the largest contemporary icefield in the Canadian Rockies. The main objective of this study is to determine whether the major outlet glaciers from the Columbia Icefield experienced temporally synchronous responses during the Little Ice Age.

Glacial histories are often developed using a variety of techniques, the most effective and frequently applied being dendrochronology, lichenometry and radiocarbon dating. However, discrepancies between the dating of events at various glaciers may be real differences in the timing of glacier advances or may simply reflect differences in the dating techniques used (Luckman 1996a). Assessment of the synchroneity of individual advances would be much easier if the same dating technique can be used to reconstruct the glacial histories for several glaciers. Therefore, a fundamental goal of this thesis is to attempt to produce precise chronologies for the glaciers at these sites.

1.2 Site Selection

The Columbia Icefield is the largest and most accessible icefield in the Canadian Rocky Mountains (Ford 1983, Figure 1.1). As a result of its size, a variety of different glacier types and environments are represented. However, all major outlet glaciers extended below treeline at their Holocene maxima, creating well developed trimlines that may be precisely dated using dendrochronology. Dating based on a single technique

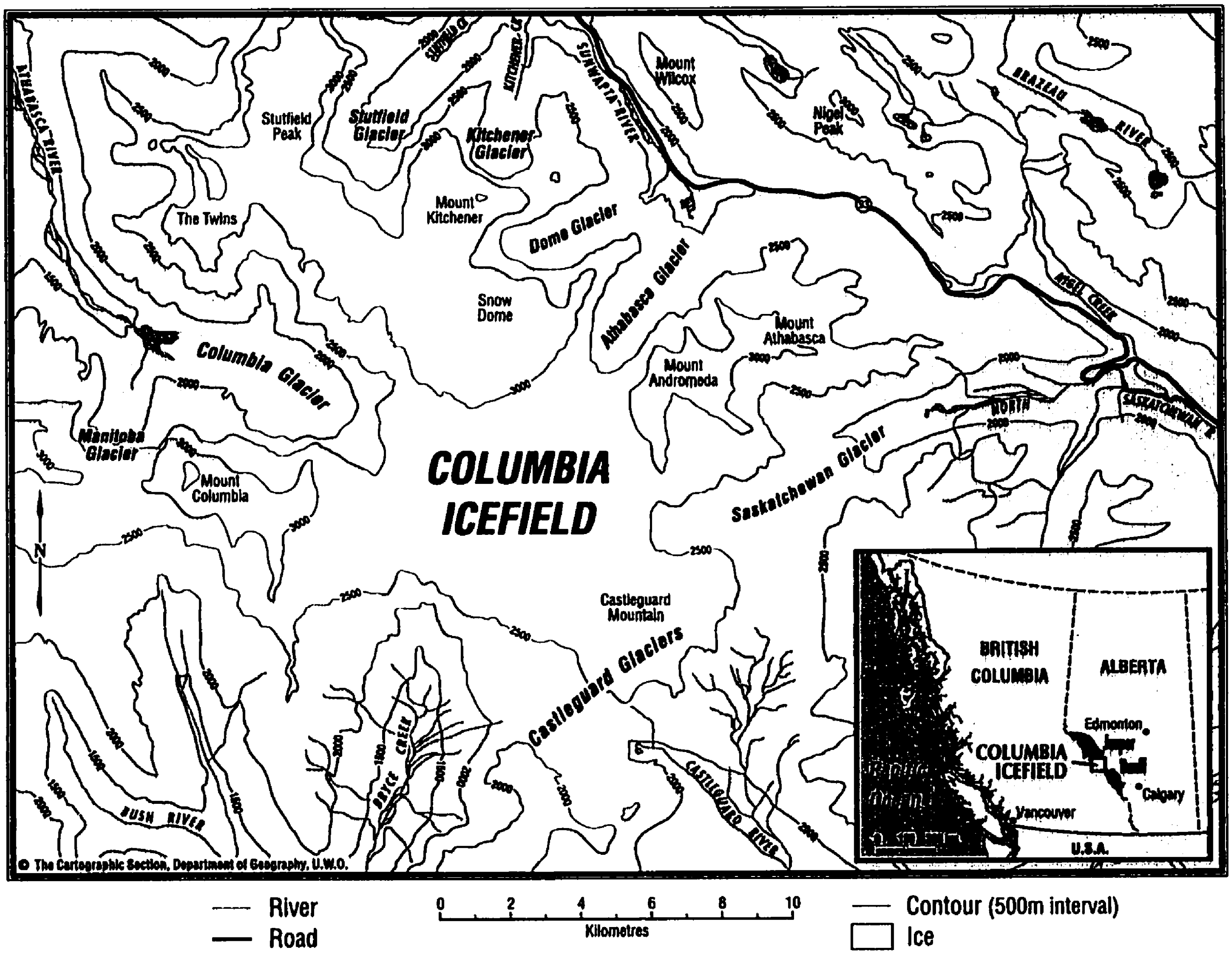

Figure 1.1 The Columbia Icefield and environs.

should reduce possible dating discrepancies due to variations in the technique used.

As a result of its accessibility, several outlet glaciers of the Columbia Icefield have been previously studied and glacial histories for four outlets have been previously reconstructed (Heusser 1956, Luckman 1988). At Athabasca and Dome Glaciers, Luckman (1988) used dendrogeomorphic evidence to date the LIA maxima precisely. It therefore appeared that similar, precisely-dated evidence should be obtainable from other sites in the area. It has also been argued that the local history of the Athabasca Glacier is reasonably representative of the regional glacial history of the Canadian Rockies (Luckman *et al.* 1997). The presently available glacial histories of Columbia and Saskatchewan Glaciers are based on the classic work of Heusser (1956). Since that time, regional reference tree-ring chronologies have been developed that permit the cross-dating of dead material which can be used to determine precise dates for older events, e.g. by dating *in-situ* portions of overridden forest which are presently being exposed as the ice retreats (Luckman 1993, 1995). In order to construct a more representative glacier history of the Columbia Icefield, it was necessary to enlarge the sample of glaciers studied. The six glaciers selected for this study complement the existing data base and provide the coverage necessary to establish regional characteristics of the Icefield. While all were located in reasonably close proximity to each other, each glacier has distinctive topographical and morphological characteristics that may have influenced its activity.

Saskatchewan and Columbia Glaciers were selected because they are the two largest outlets of the Columbia Icefield. Although Heusser (1956) developed the histories for these glaciers, new reference chronologies are now available to enable precise dating as well as the fact that the chronologies produced by Heusser (1956) at other sites have been considerably revised by more recent studies (e.g. Luckman 1988, 1995, 1996c). In addition, the Columbia Glacier is the only actively calving glacier, the major outlet on the western periphery of the Icefield and was thought to have advanced in the 1970's (Baranowski & Henoch 1978). Since the Columbia Glacier coalesced with the Manitoba Glacier at the LIA maximum, the Manitoba Glacier was also investigated. Kitchener and Stutfield Glaciers are heavily debris covered lobes whose LIA histories have not been studied. Both glaciers are located in adjacent valleys near the north end of the Icefield but

Kitchener Glacier terminates almost 400 m higher. The final site selected was Castleguard Glacier. This glacier was selected because it: has southerly position, is laterally unrestricted and appeared to have a well developed, distinct trimline in the aerial photographs.

1.3 Outline of this Study

The presentation of this research will proceed as follows. Chapter II will discuss several methods whereby contemporary glaciers can be used to infer recent (Holocene) climate change. Chapter III will provide a general description and history of the study area. The methodology employed in this study will be outlined in Chapter IV. This will include the mapping procedures, dating techniques used and their limitations, sample collection and data analysis. Chapters V to IX will present a detailed description, data and chronology for glacial events at each of the six glaciers investigated. Finally Chapter X will discuss the overall chronology of events at the Columbia Icefield. This chapter will also evaluate the synchroneity of these events and comment on the precision of the dating technique used.

CHAPTER II

CLIMATE CHANGE DOCUMENTED IN CONTEMPORARY GLACIERS

2.1 Introduction

Glaciers are thermal phenomena that depend on the addition, removal and deformation of ice for their existence (Paterson 1981). Therefore, they require specific combinations of climatic conditions for their persistence and have historically been used as indicators of former climate (Meier 1965, Porter 1981). Temperate mountain glaciers are especially sensitive to climate change for four reasons: 1) as a result of their thermal regime and proximity to melting conditions, only a small climatic change is required to cause a change in the glacier characteristics, 2) because of their small size, a change in volume is more readily noticeable, 3) their small size and shorter length results in a more rapid response time between climate change and the associated glacier response and 4) many of these glaciers are located in easily accessible areas and have at least a limited record of human occupation (Haeberli *et al.* 1989a, Haeberli 1995a, Luckman 1996, pers. comm.). This chapter will review different ways that such glaciers have been used to document climate change in alpine areas.

2.2 Climate Change and Glacial Fluctuations

There is a close connection between regional climate, the mass balance of a glacier and its snout fluctuations. Figure 2.1 is typical of diagrams used to demonstrate how the position of glacier margins are related to climatic conditions (e.g. Meier 1965, Sugden & John 1976, Porter 1981, Whalley *et al.* 1989). Variation in the energy and mass fluxes at the glacier surface result in changes in the amount of accumulation and ablation. The difference between these two amounts defines the net mass balance and the gain or loss of mass for a particular glacier. The change in mass balance alters the flow dynamics of the glacier, affecting the velocity and possibly resulting in a change in area. An increase in mass causes the glacier to increase in area, advance and create landforms such as moraines that remain to mark the former ice margin. This simplified account of the interaction between climate and glacier position does not account for any feedback effects, outside

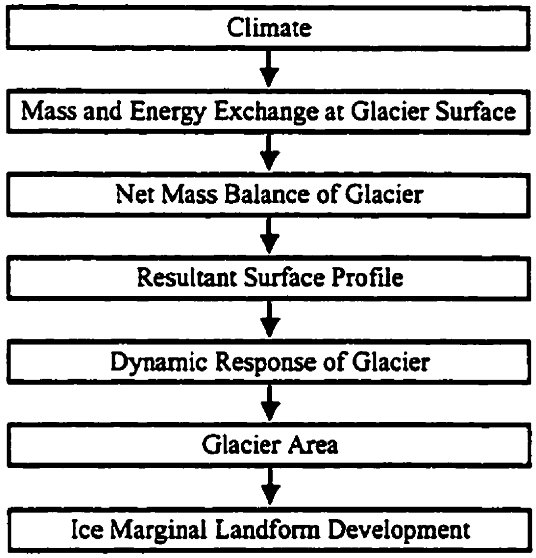

Figure 2.1 Relationship Linking the Position of the Glacier Snout with Climate (Adapted from Meier 1965).

influences or phenomena such as surging or calving that may also influence these relationships.

Climatic change can be detected using glaciers in several ways. Glacier mass balance studies identify the glacier's direct response to climate whereas changes in glacier area reflect an indirect, delayed climate signal (Haeberli 1995a, b). The equilibrium-line altitude (ELA) is a critical reflection of the mass balance of a glacier. Therefore, studies involving ELA fluctuations provide an intermediate signal on the continuum between mass balance and glacier area. Glacier inventories document changes in the global distribution of glaciers that are also an important source of information concerning climate change. Each of these methods will be subsequently described and discussed.

2.2.1 Mass Balance Studies

Studies of glacier mass balance document "changes in mass of a glacier and the distribution of these changes in space and time" (Paterson 1981, p. 42). Mass balance studies monitor the accumulation and ablation of a glacier directly, enabling identification of the direct, undelayed response to climate (Haeberli 1995a). For this reason, mass balance studies are the most useful type of glacial study to document climate change since they provide an important link connecting mass variations to climatic fluctuations.

The mass balance is the change in mass due to differences in accumulation and ablation measured over a given period time (Sugden & John 1976, Østrem & Brugman 1991, Benn & Evans 1998). Accumulation refers to all processes that increase a glacier's mass, including but not exclusive to firnification, avalanching, rime formation and freezing rain. Ablation includes all processes through which mass is lost; such as melting run-off, evaporation, sublimation, snow removal by wind and calving (Paterson 1981, Benn & Evans 1998). The thickness of the snow and ice is measured at specific points on the glacier's surface at various times through the year to determine how much mass was gained or lost and then integrated over the glacier's area to determine the total mass balance (Østrem & Brugman 1991). The total mass balance is usually calculated on a yearly basis, reflecting the change in mass over a hydrological year (Paterson 1981, Østrem & Brugman 1991). The final balance may be positive or negative, depending on whether there was a net increase or decrease in mass.

Since mass balance studies are concerned with measuring parameters directly affected by the climate, they reflect an unfiltered climate signal. In addition, the mass balance is calculated on a yearly basis, enabling scientists to monitor changes in a particular glacier at that temporal scale and provides a means to document change over short time intervals (Haeberli 1995b). Furthermore, it is possible to express changes in ice thickness as a result of melting or freezing in terms of the amount of energy required (Haeberli 1995a). If gains and losses in mass can be expressed as energy, it is possible to compare these to differences in radiation. For these reasons, it would appear that mass balance studies are the most appropriate method to document climate change from glaciers.

Unfortunately, mass balance studies require intense, time consuming and expensive sampling. Furthermore, since mass balance cannot be measured retroactively, information obtained from these studies is restricted to the period of record, commonly less than 45 yrs. (Oerlemans 1994, Haeberli 1995a). While long term, large scale climate fluctuations cannot generally be identified from a 45 yrs. record, proxy indicators such as meteorological and tree-ring data can be used to estimate mass balance for past environments and extend the length of the potential record.

2.2.2 Changes in Glacier Dimensions

A change in the size of a glacier may be manifested as a change in length, area and/or volume. Changes in glacier length represent a simple measure for changes in size whereas area variation provides a more accurate representation of changes in size as a result of the inclusion of the dimension of width. Although volume changes are ideal and provide the most information concerning glacier change, such measurements are expensive, time consuming and technologically intense. Therefore, changes in the glacier area are frequently used as a surrogate for volume.

Areal changes of a glacier are most clearly observed as changes in the position of the snout. Although several studies report minor oscillations of glacier snouts directly related to the interannual variability of climate, most significant changes in area result from longer term (decade to century long) climatic events (Sugden & John 1976, Haeberli 1995a). Haeberli (1995a, b) has characterised these changes as filtered, delayed glacial responses to changes in climate since the area does not significantly change until there has been a shift in the mass balance. Negative mass balances generally result from increased ablation which tends to affect the snout directly and therefore negative mass balances may be reflected almost immediately in the retreat of the snout. However, response to positive mass balances may be delayed at the glacier snout because it takes a period of time for the excess mass to be transferred from the accumulation area (Sugden & John 1976).

Most snout variations are adjustments to changes in mass balance. These variations indicate that the glacier is attempting to achieve a new equilibrium with the environment (Sugden & John 1976). Many authors argue that a glacier is constantly in a state of flux between positive and negative mass balance and never achieves true equilibrium (e.g. Sugden & John 1976). Therefore, there is a lag between the change in mass balance and achievement of a new equilibrium snout position that is called the response time (Paterson 1981). Response times differ between glaciers as a result of variations in size, bed topography, elevation, mass distribution, aspect, etc. (Oerlemans 1994, Haeberli 1995a). As a result of these variations in response times, changes in different sized glaciers may reflect climate change of different scale. Small cirque glaciers and snowfields may respond almost immediately to a change in climate, reflecting yearly climatic variation (Sugden & John 1976, Haeberli *et al.* 1989a). However, larger valley glaciers typically have response times ranging from several years to several decades. Therefore, changes in the frontal positions of these glaciers reflect longer term climatic shifts; annual oscillations are filtered out and what remains is the smoothed, generalised climate signal (Haeberli 1995a).

Although climate changes affect glacier mass, in most cases glacier area is substituted for volume as an index of these changes. There are several advantages in substituting area variation for volume as a means to document climate change. Primarily, glacier volume is difficult to determine because the bed topography and therefore depth of most glaciers is unknown. Area variation studies do not require information about the depth of ice since it uses change in area as a proxy for any change in volume. This simplifies the calculations that must be done and reduces the cost since expensive equipment is not required. Area change can be determined from maps or air photos without additional data requirements.

The greatest advantage of using area variation studies is that they do not require continuous monitoring of ice front positions. In Europe, reliable historical records exist for glacier positions dating back to the beginning of the 18th century (Sugden & John 1976). In addition, former ice front positions can be reconstructed from moraines and other associated glacial landforms without direct historical records. Moraines delimit the absolute maximum position for a glacial advance, reflecting the transition from a period of net positive mass balance to a negative mass balance (Porter 1981). Therefore, if the age of these deposits can be determined, this data marks the transition from a period of cooler and/or wetter conditions to a warmer and/or drier climate.

There are limitations to the use of glacier area to infer past climate conditions. The first concerns the nature of the information obtained from these types of studies. Since the processes linking climate to glacier behaviour are complex, specific parameters such as temperature or precipitation cannot be calculated. Therefore, changes in glacier area can only be used as indicators for changes in general climatic conditions (Meier 1965, Sugden & John 1976, Whalley *et al.* 1989, Haeberli 1995a). In addition, changes in snout position may be triggered by factors unrelated to climate, e.g. surging or calving.

Although there have been numeric simulations linking climate to mass balance to length variations for several glaciers, they are only possible for glaciers with extensive data on ice volume and based on large samples (Haeberli 1995a). For most glaciers, the period of record is insufficient to link specific climate conditions directly to area variation. Therefore, inferences about specific mass balance parameters, e.g. accumulation and ablation cannot be made directly from studies of area variation (Meier 1965, Whalley *et al.* 1989).

Another limitation is that the record of former ice front positions is discontinuous and therefore incomplete. Morphologic and stratigraphic evidence of early glacier margins is frequently destroyed by subsequent advances. The evidence found in most forefields represents ice front positions that have not been overridden by later, more extensive advances. Therefore, a complete record only exists for the most extensive advance and subsequent smaller advances (e.g. Luckman 1988, 1995a). Although these limitations restrict the applicability of area variation studies, these studies continue to be widely accepted as a method to document climate history using glaciers (e.g. Smith *et al.* 1995, Luckman 1995a, Clague & Mathewes 1996).

2.2.3 Equilibrium-Line Altitude (ELA) Fluctuations

Estimates of changes in the ELA may be used to make inferences about climate change (Benn & Evans 1998). The equilibrium-line is the theoretical boundary between the accumulation and ablation zone at the end of the melt season; where the annual accumulation is equal to the annual ablation. Since the equilibrium line is closely linked to the mass balance of a glacier, variation in the ELA is considered to be a climatically sensitive parameter that can be used to assess changes in climate through time (Porter 1981, Nesje 1992, Nesje & Dahl 1992, Benn & Evans 1998). The position of the ELA associated with zero mass balance is referred to as the steady-state ELA (Benn & Evans 1998). When the annual mass balance of a glacier is negative, either from decreased precipitation or increased ablation, the ELA rises. Conversely, with increased precipitation or decreased ablation, the ELA falls, reflecting positive mass balance conditions.

The most accurate method of determining the ELA of a glacier is through mass balance studies. However, direct measurement of mass balance is expensive, time consuming and restricted to relatively short lengths of record (Benn & Evans 1998). Therefore, to avoid these costs, many researchers use the position of the snowline at the end of the ablation season as a proxy indicator for the location of the ELA. Any changes in the snowline are indicators of a climate change (Porter 1981, Nesje & Dahl 1992, Torsnes *et al.* 1993, Aa 1996).

Studies using ELA can also be used to determine quantitative estimates for climate change for periods beyond the historical record. Once the former margin of the glacier has been identified, the corresponding steady state ELA is estimated. Several techniques have been created to provide accurate estimates of the ELA: most are based on the relationship between accumulation area, ablation area and/or total area. Some of the most common techniques used are: the maximum elevation of lateral moraines (MELM), median elevation of glaciers (MEG), toe-to-headwall altitude ratio (THAR), accumulation area ratio (AAR) and the balance ratio method. These have been discussed extensively in the literature (e.g. Hawkins 1985, Nesje 1992, Nesje & Dahl 1992, Torsnes *et al.* 1993, Aa 1996, Benn & Evans 1998).

One of the greatest advantages in using snowline as a proxy for ELA for present day glaciers is that intense field measurement is not required. Unfortunately, without detailed measurements for comparative purposes, there is no method to assess the validity of that assumption for a particular glacier. Since the ELA is a climate sensitive parameter dependent mainly on summer temperature and winter precipitation, a shift in the ELA reflects a change in either or both of these factors (e.g. Nesje & Dahl 1992, Caseldine & Stötter 1993, Torsnes *et al.* 1993, Aa 1996). A change in the ELA can be converted into a change in temperature by multiplying the elevational difference by the environmental lapse rate. Since most reconstructed ELA studies emphasise summer temperatures, a change in ELA is commonly assumed to be thermal change assuming relatively constant precipitation. However, as the temperatures get colder, the amount of precipitation frequently decreases (Nesje & Dahl 1992, Caseldine & Stötter 1993). Therefore, resultant estimates of temperature depressions may be in error. Another limitation to this type of study is that reconstructed ELA's are based on known ice front positions and are therefore subject to many of the same limitations as areal studies such as the incomplete and discontinuous nature of the deposits. Finally, the numerous assumptions involved in the techniques may lead to significant differences in the reconstruction of the ELA by different methods (see discussion in Nesje 1992, Torsnes *et al.* 1993).

2.2.4 Glacier Inventories

The main purpose of glacier inventories is to document the distribution of glacierised areas and provide baseline data for perennial ice masses in a region (Haeberli *et al.* 1989b). They serve to classify and catalogue glaciers in a standardised fashion and provide documentation of a variety of descriptive parameters of permanent ice bodies (Haeberli 1995a). This enables researchers to make generalisations about ice bodies that would not be possible from smaller samples or detailed mass balance or area variation studies. Repetition of these inventories permits the identification of large scale changes in glacier distribution that may be related to regional or global climate changes (Haeberli *et al.* 1989a, b).

One of the greatest advantages of glacial inventories is that the bulk of the information can be collected from published or remotely sensed data sources such as satellite images, aerial photographs or topographic maps by relatively unskilled personnel. Detailed fieldwork is not required, and therefore glacier inventories are not restricted by accessibility limitations, allowing documentation of all glaciers, even in remote areas. A large sample population of glaciers minimises local effects and allows researchers to determine the general attributes of glaciers in an area, observe regional signals and identify anomalies (Sharp 1960, Haeberli *et al.* 1989b). This provides a more accurate representation of the regional conditions than could be obtained from glacier specific studies (Haeberli *et al.* 1989a). However, when glacier inventories are used in conjunction with mass balance studies, they can be used to evaluate the representativeness of glaciers selected for continuous measurement (Haeberli 1995a).

Glacier inventories can provide a valuable source of information concerning climate fluctuations, but the results need to be verified. Without detailed fieldwork or skilled personnel assembling the data, ambiguous features such as heavily debris covered glaciers may be misrepresented on maps and misinterpreted or missed on aerial photographs. Glacier inventories are limited to the period of time with direct observations, maps or aerial photographs. In North America, this is commonly the period for which aerial photographs are available. Although former ice margins can be inferred from terminal and lateral moraines in air photos, without detailed fieldwork, the age of these deposits is unknown, limiting the amount of information that can be obtained.

2.3 Conclusion

There are four major methods in which contemporary, temperate glacier data can be used to document climate change, namely: mass balance studies, area/snout variations, ELA fluctuations and glacier inventories. Mass balance studies document direct, generally short term glacier changes, whereas glacier inventories provide invaluable information concerning the spatial distribution and variation of ice bodies. However, both are limited in their applicability to the length of record and period for which data sources are available. Studies that reconstruct former ice margins based on glacial landforms provide evidence for the maximum and some glacier recessional data although the overall record of glacier fluctuations is incomplete. Although ELA studies have been used to provide changes in summer temperature (e.g. Nesje & Dahl 1992, Caseldine & Stötter 1993, Torsnes *et al.* 1993, Aa 1996), the large number of assumptions required also increase the possible sources of error. Due to the complex relationship between climate change and snout position, only general inferences can be made about past conditions. Area variation studies are simple, easily accomplished and provide estimates of changes in glacier size. Therefore, they are the most frequently applied method for reconstructing past glacial environments and the associated climate. The abundance of studies has enabled the identification and examination of regional or larger scale climatic events such as the Little Ice Age.

CHAPTER III
THE COLUMBIA ICEFIELD AND GLACIAL FLUCTUATIONS OF THE SOUTHERN CANADIAN ROCKIES

3.1 Introduction

This chapter reviews the general characteristics of the Columbia Icefield region. It also contains a detailed glacial history of the Canadian Rocky Mountains from the Wisconsin deglaciation until the present. This will include a discussion of previous work on outlet glaciers of the Columbia Icefield that are not examined in this study (i.e. Athabasca and Dome Glaciers). Detailed descriptions of previous work and site characteristics of the six glaciers that are the focus of this thesis will be presented in subsequent chapters.

3.2 Location and Topography

The Columbia Icefield lies within the Main Ranges of the Canadian Rocky Mountains (Figure 1.1) and is the most accessible icefield in the Canadian Rockies. It straddles the continental divide, feeding the Saskatchewan, Athabasca and Columbia rivers flowing to the Atlantic, Arctic and Pacific Oceans, respectively, and is one of only two triple-continental divides in North America (Thorington 1925, Baranowski & Henoch 1978). To the west, the Columbia Icefield merges into Clemenceau Icefield. Robinson (1996) carried out a detailed glacier inventory of the Columbia Icefield area and estimated ice extent to be 228 km^2 in 1992. It is the largest contiguous icefield in the Canadian Rockies (Ford 1983).

The Columbia Icefield occupies an upland plateau at about 2700 m.a.s.l and has a northwest-southeast trending axis. It is bounded by Castleguard Mountain (3083 m), Mt. Andromeda (3442 m), Stutfield Peak (3471 m), Mt. Kitchener (3511 m), The Twins (3689 and 3564 m) and the second highest peak in the Canadian Rockies, Mt. Columbia (3747 m). The ice mass is almost equally distributed on both sides of this axis, with 45% east of the main axis and 55% to the west (Robinson 1996). The icefield has a gently rolling surface topography with 65% of the area above the average 1992 snowline of

2500 m.a.s.l.. It extends from 3460 m.a.s.l. at Snow Dome to below 1400 m.a.s.l. at the toe of an unnamed outlet glacier at the south end of the icefield. The Columbia Icefield has six major outlet glaciers; Athabasca, Castleguard, Columbia, Dome, Stutfield and Saskatchewan. Numerous, smaller unnamed glaciers with an average area of 4.8 km^2 (Robinson 1996) are also present. Although two thirds of the 27 glaciers are located west of the central axis, the Columbia Glacier is the only major glacier that flows to the west. Most of the outlet glaciers are characterised by spectacular icefalls between 2000 to 2800 m.a.s.l. and flow into deeply incised valleys.

3.3 Geology

The Columbia Icefield occupies an upland plateau developed on resistant Lower Paleozoic formations which have been overthrust from the west along the Simpson Pass Thrust onto Upper Paleozoic rock (Beaudoin 1984, Gadd 1995). The bedrock underlying the Icefield area is predominantly Mid-Cambrian limestones, including the Cathedral, Stephen, Eldon, Pika and Arctomys Formations (GSC 1977, Ford 1983, Beaudoin 1984). However, the Athabasca Valley to the west is underlain primarily by the Early Cambrian Gog Group quartzite (GSC 1977). At the south margin of the Icefield, Upper Cambrian limestone and shale outcrops can be observed (Ford 1983).

3.4 Vegetation and Soils

The vegetation and soils of this area are discussed in detail in the ecological survey carried out by Holland & Coen (1982). Areas above treeline (ca. 2100 m.a.s.l., Kavanagh 1997, pers. comm.) are dominated by ice, bedrock and alpine tundra vegetation composed of sedges, mosses and krummholz trees (Holland & Coen 1982). Below treeline, vegetated valley sides are mainly closed coniferous forest dominated by Engelmann spruce (*Picea engelmannii*) and subalpine fir (*Abies lasiocarpa*) with minor populations of Whitebark pine (*Pinus albicaulis*) and Buffaloberry (*Shepherdia canadensis*). Glacier forefields are dominated by unmodified glacial and fluvioglacial deposits and are largely vegetation free. However, beyond the LIA limits, the valley floors often contain extensive areas of active outwash or are covered by shrub and bog communities on older floodplain surfaces (Holland & Coen 1982).

Soils in the general vicinity of the Columbia Icefield have been studied by Bowyer (1977), Beaudoin (1984), King (1984) and Beaudoin & King (1994) among others. In this region the soils are predominantly brunisols and regosols (Holland & Coen 1982); however, individual glacier forefields have relatively sparse vegetation cover and are generally classified as unmodified glacial deposits. Regosols commonly occur where geomorphic processes inhibit soil development, whereas brunisols are frequently found in forested environments with an intermediate degree of soil development (Hausenbuiller 1985). However, Holland & Coen (1982) indicate that these designations are based on generalisations and may not reflect the diversity of soils at a specific site.

3.5 Climate

Data concerning the present climatic conditions at the Columbia Icefield are sparse. Fragmentary records were kept at the Icefield Centre for the period between 1961 and 1990 (Gadd 1995) and a detailed summary of regional climate records can be found in Luckman & Seed (1995). Longer records exist for nearby meteorological stations at Lake Louise, Banff and Jasper in Alberta and Valemount and Golden in British Columbia (Luckman 1997, Luckman *et al.* 1997).

The climate of the Canadian Rocky Mountains is continental, with long, cold winters and cool summers (Janz & Storr 1977). The several intervening mountain ranges between the Rockies and the Pacific Ocean provide a partial rain shadow effect. However, the absence of any major ranges to the east allows moisture from Hudson Bay, the Arctic Ocean and the Gulf of Mexico to reach this area (Janz & Storr 1977). This invasion of continental tropical air in the summer and continental arctic air in the winter may result in extreme and variable conditions (Ryder 1989). Precipitation is evenly distributed throughout the year, although the majority of it falls as snow (Janz & Storr 1977). The most important controls on local climate for this region are topography and elevation, with higher elevation sites being wetter and cooler. This results in considerable variability between sites due to differences in aspect and topography (Janz & Storr 1977).

3.6 Glacial History

Radiocarbon dates from a variety of sources indicate that Late Wisconsin glaciers in this region had receded to within or near their present positions prior to 10 ka BP. Levson *et al.* (1989) provide the earliest date for deglaciation, 11,900 ± 120 yrs. BP based on gastropods collected from a glaciolacustrine deposit interpreted to be an ice-marginal lake in the Pocahontas area of the Athabasca Valley. Since this site is located in the Front Ranges of the Rockies, about 40 km east of the Jasper townsite, it may not provide precisely limiting dates for events in the main valleys to the west. In the Bow Valley to the south, Reasoner *et al.* (1994) obtained a radiocarbon date of 11,330 ± 330 yrs. BP from the base of sandy clays directly overlying a Wisconsin diamict. As this site is very close to LIA moraines in the floor of the main Bow Valley, it indicates that it is unlikely that the glacier had receded to a position within LIA limits prior to 11,300 yrs. BP. Basal dates from other sites near the Columbia Icefield are slightly younger; Tonquin Pass (9660 ± 280 yrs. BP, Luckman & Osborn 1979), Wilcox Pass (9660 ± 350 yrs. BP, Beaudoin 1984), Castleguard Meadows (9600 ± 300 yrs. BP, Luckman & Osborn 1979), Watchtower Basin (9445 ± 370 yrs. BP, Luckman & Kearney 1986) and Saskatchewan Crossing (9300 ± 170 yrs. BP, Westgate & Dreimanis 1967), most of which are located in close proximity to the Columbia Icefield.

The Holocene glacial history of the Canadian Rockies can be divided into several distinct periods. Generally, the LIA advance was the most extensive, eliminating most of the evidence of earlier advances. However, evidence for several earlier events exists within the LIA limits (Table 3.1 and Luckman *et al.* 1993a). Immediately following Wisconsin deglaciation, there is evidence for an advance similar in magnitude to the LIA maximum called the Crowfoot Advance (Luckman & Osborn 1979). The next recorded phase of glacial activity occurred in the Neoglacial, a period of generally cooler conditions that began approximately 5000 yrs. BP (Porter & Denton 1967). Within this period, there is evidence for several discrete glacier advances that will be subsequently discussed.

Table 3.1

Dating Control for Pre-Little Ice Age Events in the Canadian Cordillera

Site	Advance Date	Evidence	Source
	Canadian Rockies		
Bow & Crowfoot Lakes	11,330 - 10,100 yrs. BP	inorganic sediments	Reasoner *et al.* 1994
Crowfoot Glacier and Cliff	> 6800 yrs. BP	tephra near surface of moraine & rock glacier	Luckman & Osborn 1979
Athasbasca Glacier	> 6800 yrs. BP	tephra on kame moraine	Luckman & Osborn 1979
East Flank Mt. Cathedral	> 6800 yrs. BP	tephra on rock glacier	Luckman & Osborn 1979
Hector, Crowfoot and Bow Lakes	>7500 yrs. BP 4000-2650 yrs. BP 900-750 yrs. BP	increased sedimentation rates	Leonard 1986, 1997
Boundary Glacier	<4200-3800 yrs. BP	overridden peat & associated tree remains	Gardner & Jones 1985
Robson Glacier	3700-3400 yrs. BP	detrital logs & *in-situ* stumps	Luckman 1995
	3300-3100 yrs. BP	overridden forest	Luckman *et al.* 1993a
	1142-1350 AD *	overridden forest	Luckman 1995
	> 450 yrs. BP		Heusser 1956
Peyto Glacier	3300-2490 yrs. BP	detrital logs, *in-situ* stumps	Luckman *et al.* 1993a
	1900-1500 yrs. BP	*in-situ* stumps	Luckman 1996c
	1246-1375 AD*	*in-situ*, sheared stumps	Luckman 1996c
Saskatchewan Glacier	3300-2800 yrs. BP	detrital logs	Luckman *et al.* 1993a, 1994
	> 2500 yrs. BP	detrital logs	
Yoho Glacier	> 2830 yrs. BP	detrital log	Luckman *et al.* 1993a
Cavell Glacier	1900-1600 yrs. BP	detrital logs	Luckman *et al.* 1994, 1995

(cont'd)

Table 3.1 (cont'd)

Site	Advance Date	Evidence	Source
		British Columbia	
Bugaboo Glacier	> 6800 yrs. BP 3390-3070 yrs. BP 2500-1900 yrs. BP > 900 yrs. BP	tephra, charcoal & humus paleosol wood & paleosol paleosol	Osborn 1986, Osborn & Karlstrom 1989
Mt. Garibaldi, Tiedemann, Klinaklini Franklin, Gilbert and Bridge Glaciers	6000-5000 yrs. BP (Garibaldi phase) 3300-1900 yrs. BP (Tiedemann advance) >900 yrs. BP	overridden stumps peat, meltwater sediments & wood *in-situ* tree remains	Ryder & Thomson 1986
Frank Mackie, Berendon, Salmon (Tide and Summit lakes, Berendon Fen)	2800-2200 yrs. BP >1600 yrs. BP 1000-700 yrs. BP	conifer needles, seeds tree fragments, conifer needles wood and needles	Clague & Mathews 1992, Clague & Mathewes 1996
Jacobsen, Purgatory, and Fyles Glaciers (Ape Lake)	< 2500 yrs. BP >770 yrs. BP	wood & soil forest litter & log in lacustrine sediments	Desloges & Ryder 1990
Kiwa Glacier	1150-1350 AD	overridden stump	Luckman 1986
Stikine-Iskut area (Scud Glacier and Glacier "b")	600-500 yrs. BP	paleosol & *in-situ* stump	Ryder 1987

Note: * previous research at these sites indicated radiocarbon dates of 1150-1250 AD at Robson Glacier (Luckman 1986) and 1140-820 yrs. BP at Peyto Glacier (Luckman *et al.* 1993a).

3.6.1 Crowfoot Advance and the Early Holocene

The earliest post-Wisconsin advance, the Crowfoot Advance, pre-dates the eruption of Mt. Mazama ca. 6800 yrs. BP (Bacon 1983). Mazama tephra is found near the surface of moraines and rock glacier deposits in a 500 km strip extending from northern Montana to the southern portion of Jasper National Park (Luckman & Osborn 1979, Osborn & Luckman 1988). Initially, this provided a minimum limiting age of 6800 yrs. BP for the Crowfoot Advance. This was supported by lacustrine records from Bow, Crowfoot and Hector lakes that suggest there was a period of increased glacial activity prior to 7500 yrs. BP (Leonard 1986). Recently, Reasoner *et al.* (1994) provided closely limiting dates for this advance from inorganic sediments associated with the Crowfoot moraine in Bow and Crowfoot Lakes. Bracketing dates for these sediments indicate they were deposited between 11,330 and 10,100 yrs. BP, making the Crowfoot Advance synchronous with the European Younger Dryas cold event.

At those sites with evidence of the Crowfoot Advance in the Canadian Rockies, the Crowfoot moraines are present just beyond the LIA maximum (Cavell Advance, Luckman & Osborn 1979). In the majority of cirques containing Holocene glacial deposits, only late-Neoglacial (Cavell) deposits are present. This suggests that the late Neoglacial advances were usually more extensive than the Crowfoot Advance. However, the Crowfoot Advance has not yet been documented throughout the Canadian Cordillera. A buried paleosol incorporating Mazama tephra overlies a till thought to be Crowfoot age at Bugaboo Glacier in the Purcell Mountains of southeastern British Columbia (Osborn & Karlstrom 1989). Further west in the Coast Mountains, Ryder & Thomson (1986) found no evidence to support a glacial advance concurrent with the Crowfoot.

Following the Crowfoot Advance, proxy climate indicators suggest generally warmer conditions throughout the Hypsithermal. Warmer than present conditions have been recorded by palynological (Kearney & Luckman 1983a, b, Beaudoin 1984, Luckman & Kearney 1986) and geomorphological studies (Luckman 1988, Osborn & Karlstrom 1989, Luckman *et al.* 1993a). Kearney & Luckman (1983a, b) reported higher than present timberlines in the period extending from 9000 to 4500 yrs. BP, punctuated by a brief period with lower treelines between 7000 and 6700 yrs. BP. Detrital wood recovered from the snouts of Athabasca and Dome Glaciers is consistent with a warmer early Holocene (Luckman 1988, Luckman *et al.* 1993a). This material indicates that ca. 6000-6500 (Dome) and 7500-8300 yrs. BP (Athabasca), forest occupied areas upvalley of the present ice fronts that are currently occupied by ice and glacial deposits. This implies that the glaciers were less extensive and treelines were higher than present. Furthermore, the paleosol documenting the Crowfoot event at Bugaboo Glacier also provides evidence for early and middle Holocene warming and/or drying (Osborn & Karlstrom 1989).

3.6.2 Early Neoglacial Advances

The majority of evidence for Neoglacial events prior to the late-Neoglacial maximum comes from stratigraphic studies in lateral moraines (e.g. Ryder & Thomson 1986, Osborn & Karlstrom 1989) and glacial forefields (e.g. Gardner & Jones 1985, Luckman *et al.* 1993a). However, several palynological (e.g. Kearney & Luckman 1983a, b, Beaudoin 1984, Luckman & Kearney 1986) and lacustrine sedimentation studies (e.g. Leonard 1986, 1997) support the findings from these geomorphological studies.

In the Coastal Ranges of western British Columbia, glacially overridden stumps in growth-position at Garibaldi Park indicate that the earliest recorded Neoglacial advances occurred between 6000 and 5000 yrs. BP (Ryder & Thomson 1986). However, there is no evidence for a corresponding advance in the Rockies (Osborn & Luckman 1988). The earliest evidence for a Neoglacial advance in the Canadian Rockies was obtained by Gardner & Jones (1985) at Boundary Glacier. Dates of overridden peat and an associated tree stump located ca. 500 m upvalley of the LIA maximum position reveal an advance after 3800 yrs. BP. This roughly corresponds to a date from a buried forest site, dating between 3700 and 3400 yrs. BP which is presently appearing from the snout of Robson Glacier (Luckman 1995).

The next documented glacier advance occurred between 3300 and 2800 yrs. BP and is referred to as the Peyto Advance. At Peyto Glacier, *in-situ* stumps, detrital wood and a paleosol found near the snout have yielded wood dating between 3300 to 2490 yrs. BP. However, 9 of the 14 samples indicate that the trees probably died between 3000 and 2800 yrs. BP and only one postdates 2800 yrs. BP (Luckman *et al.* 1993a). This suggests that Peyto Glacier was advancing between 3000-2800 yrs. BP. Similarly dated events have also been documented at Robson, Yoho and Saskatchewan Glaciers (Luckman *et al.* 1993a, 1994, Luckman 1995). Approximately 3.5 km upvalley of the snout of Robson Glacier, Luckman *et al.* (1993a) found an *in-situ* stump and several detrital logs dating to the period between 3300 and 3100 yrs. BP.

The range of dates available for early Neoglacial events (e.g. at Boundary and Robson Glaciers) create some uncertainty as to whether these represent a single "Peyto Advance" or several advances (Luckman 1995). Although the exact dates for the inception and maximum stand of the Peyto Advance are unclear, there is considerable evidence supporting a phase of glacier advance between 3300 and 2900 yrs. BP throughout the Canadian Cordillera (Leonard 1997). In the Coast Mountains of British Columbia, Ryder & Thomson (1986) recorded an early-Neoglacial advance at three glaciers between 3300 and 1900 yrs. BP, referred to as the Tiedemann Advance. This is consistent with an advance dated to approximately 2500 yrs. BP documented through lacustrine records at sites near Bella Coola (Desloges & Ryder 1990). In the Purcell Mountains, Osborn & Karlstrom (1989) demonstrated several advances of the Bugaboo Glacier during the early Neoglacial. The first overrode a pre-existing moraine between 3390 and 3070 yrs. BP, while the second occurred between 2500 and 1900 yrs. BP. These events are all considered to be consistent with the Peyto Advance between 3300 and 2800 yrs. BP in the Rockies.

Evidence supporting the above findings is now available from a number of sources. The timing of an increase in sedimentation rates in Bow and Crowfoot Lakes between 3500 and 1900 yrs. BP corresponds closely to the advance phase of the Peyto Advance (Leonard 1986, 1997). Further west in the northern Coast Mountains of British Columbia, lake and fen cores record an expansion of the local glaciers during the period 2800-2200 yrs. BP (Clague & Mathews 1992, Clague & Mathewes 1996). Palynological evidence from a variety of locations in Jasper National Park show that timberlines dropped to levels similar or below present elevations, during the period 5000 to 4000 yrs. BP (Kearney & Luckman 1983a, b, Luckman & Kearney 1986). The vegetation assemblages found at these sites indicate moist and cool conditions which favour glacier expansion.

Most of the evidence for the Peyto Advance indicates that it did not extend much farther downvalley than contemporary ice fronts. Furthermore, since most of the evidence for these events has been obtained from buried trees and deposits which have only been exposed recently (e.g. Luckman *et al.* 1993a, 1994, Luckman 1995, 1996d) it suggests that the glacier retreat of the last 150 yrs. has removed the net accumulation of ice which built up over the last 3000 yrs. (Luckman 1996d, Luckman pers. comm. 1997). This suggests an unprecedented change in mass balance in the latter portion of the Holocene influenced by the climatic conditions of the last 150 yrs.

Evidence for a mid-Neoglacial event that postdates the Peyto Advance has been reported at two sites within the Canadian Rockies. At Peyto Glacier, a paleosol with *in-situ* stumps dating between 1900 and 1500 yr BP was uncovered approximately 1 km downvalley of the 1990 ice front (Luckman *et al.* 1994). This is consistent with detrital wood retrieved from a stream channel near the snout of Cavell Glacier with dates of 1910 ± 80 and 1660 ± 60 yrs. BP (Luckman *et al.* 1994, 1995 Luckman 1996c). A corresponding advance after 1600 yrs. BP is shown by lacustrine records at Frank Mackie and Berendon Glaciers in the northern Coast Mountains (Clague & Mathews 1992, Clague & Mathewes 1996). In the Canadian Rockies, Kearney & Luckman (1983b) recorded a further lowering of timberline and a decline in regional temperatures beginning ca. 1700 yrs. BP which was superimposed on the overall cooling trend which began at the start of the Neoglacial.

3.6.3 Little Ice Age or Cavell Advance

The term Little Ice Age is used to refer to the most recent period of moraine building which culminated in the mid 18th and 19th centuries, during which many glaciers of the Canadian Cordillera attained their maximum Holocene extent (Luckman 1986, Osborn & Luckman 1988, Desloges & Ryder 1990). This provides abundant dateable deposits, making it the most widely documented glacial event of the Holocene. Luckman & Osborn (1979) assigned the term Cavell Advance to the Little Ice Age event in the Canadian Rockies based on moraine data from 24 glaciers. Although this term has gained relatively widespread application, the Little Ice Age remains the preferred name for this advance.

3.6.3.1 Inception of the Little Ice Age in the Canadian Rockies

The earliest evidence for a Little Ice Age event was documented at Robson Glacier by Heusser (1956 and Luckman 1986, Luckman 1995, 1996a). Heusser (1956) found a buried forest site 400-500 m upvalley of the LIA maximum position that yielded a radiocarbon date of ca. 450 ± 150 yrs. BP, suggesting that the Little Ice Age began between 1300 and 1600 AD (Heusser 1956). Resampling of this site by Luckman (1986) revealed further stumps and detrital snags, three of which yielded radiocarbon dates between 1150 and 1250 AD. Tree ring series in these snags were later crossdated with long living tree chronologies from Bennington Glacier (Luckman 1993) to provide calendar death dates for the samples. Calendar dates for 8 *in-situ* stumps and 32 detrital logs indicate that Robson Glacier began overriding forest between 1142 and 1150 AD and continued to advance until at least 1350 AD (Luckman 1995).

Similar results were obtained from overridden trees approximately 500 m upvalley of the LIA maximum position at Peyto Glacier (Luckman 1996c). *In-situ* and detrital logs associated with a buried paleosol were radiocarbon dated to between 1110 and 840 yrs. BP (Luckman *et al.* 1993a) and subsequent dendrochronological investigations provided outer rings dates ranging from 1246-1375 AD (Luckman 1996c).

Glacial evidence consistent with the results presented above has been found at a variety of sites throughout B.C. A radiocarbon dated stump at Kiwa Glacier in the Premier Range indicates that it was probably killed during an advance between 1150 and 1350 AD (Luckman 1986). At Bugaboo Glacier in the Purcell Mountains, a paleosol incorporated in a lateral moraine also reflects an advance ca. 900 yrs. BP (Osborn 1986, Osborn & Karlstrom 1989). Evidence from Klinaklini, Franklin, Bridge, Scud and Purgatory Glaciers in the Coast Mountains (Ryder & Thomson 1986, Ryder 1987, Desloges & Ryder 1990), provide radiocarbon dates that document a glacier advance beginning between 900 and 600 yrs. BP.

Other proxy evidence from the Canadian Cordillera is also in agreement with an early LIA advance. Lacustrine records from the Bella Coola region (Desloges & Ryder 1990) and the northern Coast Mountains (Clague & Mathews 1992, Clague & Mathewes 1996) suggest an advance beginning ca. 1000-700 yrs. BP. In the upper Bow River drainage basin, a strong peak in sedimentation rate occurred in the mid 1200's with a secondary peak in the mid 1300's, coincident with glacial advances recorded elsewhere in the Canadian Rockies (Leonard 1997). In addition, tree-ring chronologies from a variety of sites in the Canadian Rockies show periods of reduced ring widths, which are thought to reflect cooler conditions, during the 13-14th century (Luckman 1996a). A recent summer temperature reconstruction from chronologies near Athabasca Glacier identified several intervals of cooler summer temperatures during the early 1110-1130's, 1190-1240's, 1280-1300 and in the 1320-1350's which are compatible with a period of advance at that time (Luckman *et al.* 1997).

3.6.3.2 Little Ice Age Maximum

Over the last 40 yrs., numerous studies have reconstructed the LIA histories of many glaciers in the Canadian Cordillera. Unlike the studies that use subsurface material located within the forefield to determine the beginning of the LIA, evidence concerning the maximum position is derived from moraines, trimlines and other related glacial features. Most of the data have been obtained using minimum dendrochronological dates and lichenometry using *Rhizocarpon* sp. (e.g. Luckman 1977) or *Xanthoria elegans* and *Aspicilia* (e.g. Smith *et al.* 1995). Recently, Luckman (1996a) has summarised the glacial record for 60 glaciers in the Canadian Rockies and the Premier Ranges. A compilation of these data are shown in Table 3.2.

Examination of the regional glacial record reveals 2 major periods of glacial advance in the last 300 yrs. (Figure 3.1). The largest temporal concentration of moraines date to the beginning of the 1700's and between 1825-49. At many glaciers, large numbers of moraines were formed between ca. 1840 and 1900-1920 at readvance positions after the LIA maximum. Periods of increased sedimentation rates in lacustrine records (Leonard 1997) and reduced ring width in tree ring chronologies from a number of sites in the Canadian Rockies (Luckman 1996a) provide complementary, indirect evidence of glacier fluctuations. Furthermore, the summer temperature reconstructions from the Columbia Icefield indicate that these two major periods of moraine building were preceded by the coldest decades in the last 300 yrs. (Luckman *et al.* 1997).

Table 3.2

Dated "Little Ice Age" moraines in the Canadian Rockies (Luckman 1996a)

Glacier	Outer	Other	Source	Method
Robson	1782	1804,1864,1891,1907a	Heusser 1956	D
Cavell	1723	1723,1783,1871,1901	Heusser 1956	D
	1705	1720,1858,1888	Luckman 1977	D
Columbia	1724	1842*,1871,1907a	Heusser 1956	D, Dg
Dome	1870	1900,1919a	Heusser 1956	D
	1846*		Luckman 1988	Dg
Athabasca	1714*	1841,1900a	Heusser 1956	D
		1843*	Luckman 1988	Dg
Saskatchewan	1807-1813*	1854,1893a	Huesser 1956	Dg, D
Hilda	1790		Heusser 1956	D
SE Lyell	1840*	1855,1885,1894a	Heusser 1956	Dg, D
Freshfield	1853	1881*,1905	Heusser 1956	Dg, D
Peyto	1711	1861,1880,1888a	Heusser 1956	D
		1837-1845*	Luckman 1996c	Dg
Bow	1669	1847*,1852,1889,1894,1898	Heusser 1956	D, Dg
Yoho	1865	1880,1884	Heusser 1956	D
	1844	1877,1890's	Bray & Struik 1963	D
Wenkchemna	1906-1925		Gardner 1978	P,D
President's Peak	1714	1834	Bray 1964	D
Scott	1780	1825	Shafer 1954	D
Hector	ca. 1660		Brunger 1967	D
Drummond	ca. 1570-1660		Nelson 1966	D

(cont'd)

Table 3.2 (cont'd)

Glacier	Outer	Other	Source	Method
Penstock Creek	1765	1810,1876,1907	Luckman 1977	D
Fraser	1595	1725,1880,1915	Luckman & Osborn 1979	L
Dungeon Peak	1355?	1620,1780,1880,1905a	Luckman & Osborn 1979	L
Redoubt S.	1705	1890,1905	Luckman & Osborn 1979	L
Redoubt N.	1630	1860,1890,1920	Luckman & Osborn 1979	L
Bastion Peak	1705	1885,1915	Luckman & Osborn 1979	L
Turret	1530	1602,1735,1895a	Luckman & Osborn 1979	L
S. Cirque	1700	1880	Bednarski 1979	L
Centre Glacier	1500	1650,1808,1862,1915	Kearney 1981	L
Mary Vaux	1613	1835-1842	Kearney 1981	D
Maligne Icefield	1841		Kearney 1981	D
Crowfoot	1505	1735,1851,1865	Leonard 1981	D
		1839*	Leonard 1981	Dg
Balfour	1847	1870,1889	Leonard 1981	D
Wapta Icefield	1875		Leonard 1981	D
Bennington	1703 (?)	1825,1870,1891,1906a	McCarthy 1985	D,L
Berg	1722		Luckman 1995	D
Hargreaves	1725	mid 19th century	Luckman 1995	D
Small River N.	<1539	1592,1820,1860	Luckman 1995	D,L
Small River main	1720	1838*	Luckman 1995	D, Dg
Beatty	1864		Smith *et al.* 1995	L
Elk	<1595	>1792, 5 moraines 1842-1897	Smith *et al.* 1995	D,L
French	<1410	1543-1609,1841	Smith *et al.* 1995	D,^{14}C

(cont'd)

Table 3.2 (cont'd)

Glacier	Outer	Other	Source	Method
Foch	1848	1919	Smith *et al.* 1995	L
Haig Main	1877(1844)	1877	Smith *et al.* 1995	L
Haig E.	1830		Smith *et al.* 1995	D,L
Haig W.	1833		Smith *et al.* 1995	D
Jellicoe N.	<1734		Smith *et al.* 1995	L
Jellicoe S.	und.	1903(1900),1900,1920	Smith *et al.* 1995	L
Mangin	1755		Smith *et al.* 1995	L
Marlborough	1859		Smith *et al.* 1995	L
Murray	>400 yr	1857	Smith *et al.* 1995	D
NW Northover	und.	1868	Smith *et al.* 1995	L
Petain Upper	1855		Smith *et al.* 1995	L
Petain	1836	1888	Smith *et al.* 1995	D
Putnik	1880		Smith *et al.* 1995	D
Rae	1856		Smith *et al.* 1995	L
Sarrail	<1507	>1616,1820-1840(est.)	Smith *et al.* 1995	D

Notes: a = minor, younger 20th century moraines excluded
* = calendar dated surface
Abbreviations used for techniques:
^{14}C = estimated calendar age from radiocarbon date
D = Dendrochronology (from minimum surface date)
Dg = Dendroglacialogical date (date from tree tilted or killed by the glacier)
L = Lichenometry
P = historical photographs

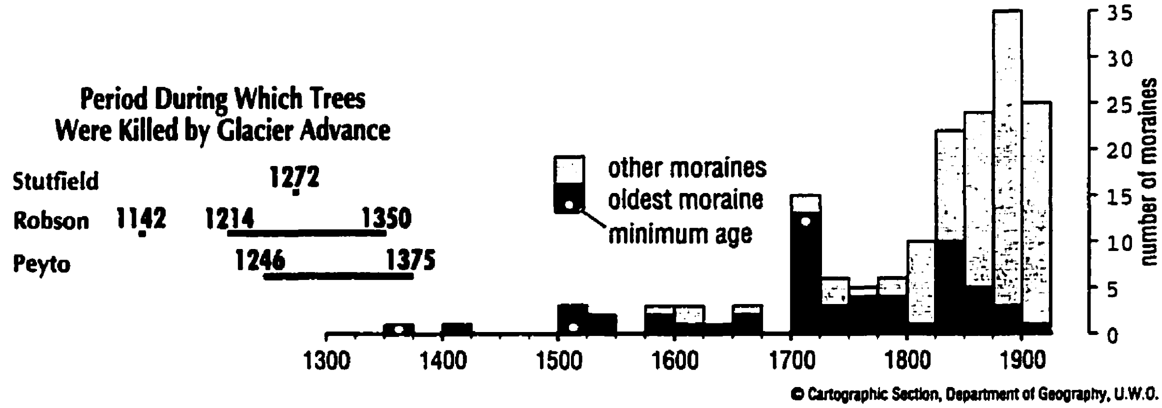

Figure 3.1 Regional record of dated Little Ice Age moraines in the Canadian Rockies. It is a composite of 60 sites (Luckman *et al.* 1997).

The maximum extent of glaciers during the LIA can be defined as the point when the glacier reached its maximum downvalley extent, normally delimited by the outermost moraine. At the majority of sites, this event was in the 19th century. Evidence from a limited sample of tilted and killed trees suggests it can be precisely dated to the 1840's (Table 3.3, Luckman 1996a). Typically, documentation of an early 18th century event occurs slightly beyond younger 19th century moraines or, more frequently is preserved as lateral moraine fragments despite a subsequently more extensive downvalley event. It must be recognised that, at several glaciers, the oldest LIA moraines do not necessarily define the maximum LIA advance. Evidence of earlier events is often preserved as small fragments of moraines in lateral positions when equivalent moraines have been overridden downvalley by subsequent advances. Therefore, when dating lateral moraines, it should be recognised that these locations may have been reoccupied several times during the LIA with the ice front in a similar position. Therefore, the oldest preserved moraine fragment in these locations does not necessarily represent the maximum position (Luckman 1996a). The maximum areal coverage of the glacier is commonly represented by the farthest downvalley terminal moraine.

3.6.3.3 Intermediate Little Ice Age Activity

At present, evidence of glacier fluctuations from the period between 1350 and 1700 AD is sparse and fragmentary. Although a few moraines have been dated to the

Table 3.3

Calendar Dated Little Ice Age Events in the Canadian Rockies after Luckman 1996a.

Glacier	Year	Evidence	Source
Athabasca	1714*	Tilted tree	Heusser 1956
	1843-44*	Tilted tree	Luckman 1988
Saskatchewan	1807,1813	2 tilted trees	Heusser 1956
Small River	1838-1839	Tree killed	Luckman 1995
Crowfoot	1839	Tilted tree	Leonard 1981
SE Lyell	1840	Tilted tree	Heusser 1956
Peyto	1837-1846**	7 *in-situ* trees killed	Luckman 1996c
Columbia	1842	Tilted tree	Heusser 1956
Dome	1846	Tilted tree	Luckman 1988
Bow	1847	Tilted tree	Heusser 1956
Freshfield	1881	Tilted tree	Heusser 1956

Notes: * at Athabasca Glacier, the 1714 date is from the lateral moraine and the 1843/44 is from the terminal moraine

** at Peyto Glacier, six trees were killed along the lateral moraine between 1837 and 1841. One tree was tilted in 1846.

early 15th and 16th centuries (Table 3.2 & Figure 3.1), the sample size is small and the dates are often minimum surface ages that are poorly constrained (Luckman 1995, 1996a). At many of these sites, the dates obtained are nearing the age limits for the tree species used and there is a possibility that the moraines were deposited prior to the LIA and the trees are survivors of a much older forest (Luckman 1995, Smith *et al.* 1995).

Although there have been no reported occurrences of non-glacial deposits between two LIA tills, there is some evidence of multiple advances during the LIA. Lacustrine records at Hector Lake indicate that the lowest sedimentation rates of the last millennium occurred between 1375 and 1550 AD, followed by sharp increase (Leonard 1997). This indicates a shift in the glacial activity in the upper Bow drainage basin during the early portions of the LIA. Furthermore, evidence from the Columbia Icefield suggests a higher treeline than present between 1300-1700 AD (Luckman 1986, Hamilton 1987, Luckman *et al.* 1997, Luckman & Kavanagh 1998). This implies a period of general warming which is incompatible with a continuous expansion of glaciers (Luckman *et al.* 1997). These records indicate that the Little Ice Age was not a single event which culminated in the 18 and 19th century maxima, but a series with distinct periods of glacial advance (Luckman 1995).

As a result of the lack of direct evidence of an intervening glacial event at many sites, several researchers at other locations in the Canadian Cordillera have advocated a single, continuous expansion throughout the LIA. They felt that lacustrine records at Tide Lake in northwestern B.C. (Clague & Mathews 1992) and Ape Lake in the Coast Mountains (Desloges & Ryder 1990) suggested that the related glaciers began advancing between 1000 and 750 yrs. BP and continued until reaching their Neoglacial maximum position during the 16th through 19th centuries. Furthermore, Osborn (1986) and Ryder & Thomson (1986) concluded that the glaciers they examined remained downvalley of their present positions from ca. 900 yrs. BP until the LIA maximum, based on the locations of overridden organics. However, when a glacier advances, it destroys the evidence of earlier, less extensive events. Therefore, the lack of intervening deposits at these sites does not necessarily imply that there were no intervening events, only that they were less extensive than the 18 and 19th century maxima. Unfortunately, until stronger evidence from the period between 1350 and 1700 AD is found, these questions cannot be resolved.

3.7 Previous Glacial Studies at the Columbia Icefield

The Columbia Icefield was first observed from Mt. Athabasca by Norman Collie and Hermann Woolley (Stutfield & Collie 1903) who described it as follows:

> The view that lay before us in the evening light was one that does not often fall to the lot of modern mountaineers. A new world was spread at our feet; to the westward stretched a vast ice-field probably never before seen by human eye and surrounded by entirely unknown, unnamed and unclimbed peaks. From its vast expanse of snows the Saskatchewan glacier takes its rise, and it also supplies the head-waters of the Athabasca; while far away to the west, bending over in those unknown valleys glowing with the evening light, the level snows stretched, to finally melt and flow down more than one channel into the Columbia River, and thence to the Pacific Ocean.
>
> (Stutfield & Collie 1903, p. 107)

Since its discovery in 1897, relatively few studies have examined the glacial history of the Icefield. The first written account of the Columbia Glacier was made by Jean Habel (1902), with the next recorded visit being made by Mary Schäffer in 1907, who travelled to the Athabasca Glacier in the following year (Schäffer 1908, 1980). During the 1910's, many glaciers in the Rockies were photographed by surveyors employed by the Interprovincial Boundary Commission (e.g. Cautley & Wheeler 1924). However, it was not until the 1920's that formal glacier studies were undertaken, primarily by Field (1924, 1949, Field & Heusser 1954), Palmer (1925) and Thorington (1925, Ladd & Thorington 1924). The first aerial photographs of Athabasca and Dome were taken in 1938 but photos of the icefield were not available until the 1940's. Between 1945 until 1954, the Dominion Water and Power Bureau conducted detailed surveys of run-off, glacier movement and variations for numerous glaciers, including Athabasca and Saskatchewan Glaciers (e.g. McFarlane 1946, Meek 1948). These surveys were continued by the Water Survey of Canada until 1980 (WSC 1982, Luckman 1988). The Saskatchewan Glacier was the focus of glaciological studies by a group from the Californian Institute of Technology in the 1950's (Meier *et al.* 1954). However, the most comprehensive evaluation of glacial histories in this area was undertaken by Heusser in the summer of 1953 (Heusser 1956). Using dendrochronology to determine moraine ages, he investigated 12 glaciers, including the Athabasca, Dome, Columbia and Saskatchewan

Glaciers in a study of regional glacier fluctuations. Following this work, only a handful of additional studies have targeted the outlets of the Columbia Icefield; primarily Athabasca, Dome and Saskatchewan Glaciers (e.g. McPherson & Gardner 1969, Baranowski & Henoch 1978, Luckman 1986, 1988).

3.7.1 Athabasca and Dome Glaciers

Athabasca and Dome Glaciers have been the foci of detailed work concerning their chronologies by Heusser (1956) and Luckman (1988). This will be briefly summarised below. Descriptions of the available documentary evidence available for each of the glaciers investigated in this study will be presented in chapters V-IX concerning the respective glacier.

At Athabasca Glacier, Heusser (1956) identified 2 periods of glacier advance: ca. 1714 and in the mid 19th century. The date of the first advance was obtained from a tilted tree located near the confluence of the lateral moraines of Athabasca and Dome Glaciers. Similar dates were obtained for additional moraine fragments found near the terminal position. Later investigations by Luckman (1988) confirmed the 18th century dates of these features (Luckman 1988). However, it was during the 19th century event that the glacier reached its maximum downvalley extent, advancing onto the lower slopes of Mt. Wilcox, directly opposite the glacier. Based on minimum surface age estimates, Heusser (1956) determined that recession from this advance began in ca. 1841. Studies by Luckman (1988) located a tree growing from the base of the distal face of this terminal moraine that began growth in the 1750's, but had been tilted during emplacement of the moraine. When this tree was sectioned in 1984, it was revealed that the trunk (from where the earlier sample had been taken) had developed from a "leader" after the initial main stem had been knocked over by the deposition of the moraine. The original stem had survived, but had lost apical dominance and been encased in callous growth within the main trunk. Examination of the lower section of the tree revealed that the tilting event occurred in 1843/44 based on a sequence of narrow rings following the damage (Luckman 1988).

At Dome Glacier, Heusser (1956) did not find any evidence for an 18^{th} century advance concurrent with the one at Athabasca. Furthermore, he did not locate any ice damaged trees, thus retreat positions have been used to date the maximum position. He determined that Dome Glacier had achieved its maximum downvalley extent in 1870, based on a tree growing on a remnant of the terminal moraine. Later investigation by Luckman (1988) showed that this ice limit could be precisely dated to 1846 from a scarred tree that grew in a small area where the western lateral moraine joins the outwash.

3.8 Summary

Studies in the Southern Canadian Rockies have documented several distinct periods of glacial advance during the Holocene: the Crowfoot Advance (Younger Dryas equivalent), several early Neoglacial advances (e.g. Peyto Advance) and the Little Ice Age. During the LIA, glaciers reached their maximum Holocene extents with major moraine building episodes in the early 18^{th} and mid 19^{th} centuries. The earliest reported evidence for the LIA advance is at Robson and Peyto Glaciers, between ca. 1200-1370 AD. At the Columbia Icefield, glacial histories of Athabasca and Dome Glaciers are consistent with this regional record. At Athabasca Glacier, two glacial advances have been dated precisely to 1714 and 1843/44 AD. At Dome Glacier, the only evidence reported was a mid 19^{th} century event, dated precisely to 1846 AD.

CHAPTER IV
DATING GLACIER FLUCTUATIONS

4.1 Introduction

The gathering of appropriate, accurate and sufficient data is critical to all scientific investigations. In order to assess the sychroneity of glacial advances, detailed histories of the individual glaciers must be constructed. This requires the determination of both the spatial and temporal context of former ice front positions. This chapter describes the field and laboratory methods employed in this type of research, including a description of the procedures employed to identify and map the salient features in each forefield. Following a general discussion, the specific techniques employed in this study will be outlined.

4.2 Determination of Former Glacier Extent

The former extent of glaciers is usually determined by the detailed mapping of landforms, deposits and vegetational features that have been formed or influenced by past glacial activity. These features can be identified from field investigations, maps produced by previous researchers (e.g. Field 1949, Heusser 1956), topographic maps and aerial photographs. Field observations in conjunction with aerial photographs prove to be the most appropriate method since they provide considerable topographic and morphologic detail without the "cartographic interpretation" inherent in topographic maps. Furthermore, aerial photographs can be used to identify recent (<50 yrs.) ice front positions not marked by moraines or trimlines.

4.3 Dating Techniques

Dating the formation of landforms, features or deposits is necessary to provide chronological control for glacial events. Several techniques are available but they vary in temporal resolution and may not be equally applicable in a given situation as a result of the lack of suitable material. The main techniques used in the Canadian Rockies are documentary sources, biological methods and radiocarbon dating.

4.3.1 Documentary Sources

Direct glacier observations from the 19th century are rare in the Canadian Rockies, but where they do exist, they can result in an effective method of dating glacier margins from that period. Tourists and mountaineering parties visiting the area at the turn of the century provide local information concerning frontal positions for individual glaciers during this period (e.g. Habel 1902, Stutfield & Collie, 1903, Schäffer 1908). The most useful regional source is the photo-topographic work completed by the Interprovincial Boundary Commission between 1913 and 1924 which produced both topographical maps and a series of dated photos of individual glaciers (e.g. Cautley & Wheeler 1924). Beginning in the 1920's, limited formal studies were undertaken to record contemporary ice front positions and determine glacial histories. Most of these are published in alpine journals (e.g. Field 1924, Ladd & Thorington 1924, Palmer 1925, McFarlane 1946, Field & Heusser 1954, Heusser 1956). While the first aerial photographs for this region were taken in 1938 (of the Athabasca Glacier), they were not generally available until the 1940's. Since then, they have been taken at regular intervals, providing precisely dated and well-documented ice front positions for that period. Overall, there is considerable variation in the quality of the different documentary sources; however, they tend to provide the best temporal precision for reconstructing recent glacial events.

4.3.2 Biological Techniques

In the absence of documentary records, the most suitable dating tools for Little Ice Age moraines in the Canadian Rockies are biologically based, namely dendrochronology and lichenometry (Luckman & Osborn 1979). Since calcareous substrates in the study area inhibit lichen colonisation, dendrochronology is the primary technique employed in the Columbia Icefield area (Luckman 1986).

4.3.2.1 Dendrochronology

Dendrochronology involves the use of tree rings to determine the age of a specimen. The tree ring is an annual growth layer in the wood or xylem of the tree (Fritts 1965). Each ring consists of earlywood; a set of large, thin walled cells produced in the spring and a layer of small, densely packed cells called latewood which are produced towards the end of the growing season (Fritts 1976, Schweingruber 1988). The transition between the earlywood and latewood is gradual whereas there is an abrupt boundary between the latewood of one year and the earlywood of the subsequent year. Therefore, an annual ring is defined by two consecutive, abrupt latewood-earlywood boundaries and represents a complete growth year.

Dendrochronology can be used to determine the age of temperate tree species that produce annual growth rings. The discipline can be subdivided into several categories, several of which involve the application of dendrochronological techniques to problems concerning the environment and climate. Dendroglaciology refers to those techniques employed to reconstruct recent glacier fluctuations, and has been useful in dating Little Ice Age limits at sites where glaciers extended below present treeline (e.g. Heusser 1956, McCarthy 1985, Ryder & Thomson 1986, Watson 1986, Luckman 1988, Osborn & Luckman 1988, Luckman 1996b, c).

Conventionally, three types of dendroglaciological evidence are used to date glacier fluctuations: namely minimum surface ages of landforms, damage to living trees and death dates for trees (Table 4.1). The first two involve the use of living trees whereas the last involves the use of dead material. When using living trees, the rings can simply be counted to determine the age of the tree or the date of damage to the cross section. Trees growing on particular features within the forefield provide minimum surface ages and thereby limiting dates for their formation (Luckman 1986, 1996b). Records of damage preserved in ring-width records provide the most precise dating of glacier events, resulting in exact calendar dates for damage. Unfortunately, these events are commonly only associated with the outermost moraines in a glacier forefield and are rarely preserved (Luckman 1996b).

The use of dead material is more complex because it requires the use of crossdating to determine the calendar dates of the tree-ring record (Fritts 1976). Crossdating is possible because tree growth and therefore ring properties are affected by environmental conditions during the growing season. Similar interannual patterns of ring-width variation are produced by many trees in an area that are exposed to the regional environmental conditions and are subject to the same limiting factors. Matching ring patterns with dated tree ring series is used to establish the exact year a particular ring or series of rings was formed. In this way, the death date of the tree can be determined and provide a limiting date for a glacial event if the tree was killed by the glacier.

Table 4.1

Evidence Employed in Dendroglaciological Dating

Evidence	Temporal Precision	Information Provided	Limitations
Tilted &/or scarred tree	Exact calendar date of damage	- damage date indicates position of glacier at a specific time	- moderately rare - dead trees may require crossdating
Trees killed by glacier	Exact calendar date of outer ring, dating precision for event depends on preservation of wood and loss of outer rings	*in-situ* - death date indicates position of glacier at a specific time detrital wood - provides limiting date for death (glacial event)	- requires crossdating with existing chronologies to determine age - loss of outermost rings in some cases
Trees growing in the forefield	10-20 yrs.	- tree provides minimum age for surface	- ecesis difficult to estimate (10-50 yrs.) - assumes the oldest tree has been sampled

4.3.2.2 Damaged Trees

When a glacier advances into forest, it creates a trimline at its margin by removing trees (Luckman 1986). During the advance, ice or glacial debris may tilt or damage trees and if those trees survive, they will attempt to restore their upright position by preferentially adding wood to one side. This results in ring-width eccentricity and a distinctive type of growth known as reaction wood. In coniferous trees, reaction wood occurs on the lower side of the lean and is characterised by wider rings with a larger proportion of latewood than normal rings (Fritts 1976). The glacier may also come into contact with the tree and remove portions of its bark. This creates a scar on the outer surface as no ring is produced in the section where the bark was removed. Subsequently, the tree may grow over the portions which were removed, but the scar tissue may be recognised by dark resinous surfaces produces within the ring series or by morphology (Luckman 1995, pers. comm.). The particular event can be dated by the last complete ring before the scar. With both tilted and scarred trees, the date of the growth aberration corresponds with the damage event, providing a dated ice front position (Luckman 1986, Watson 1986).

4.3.2.3 Trees Killed by the Glacier

As a glacier advances into pre-existing forest, trees are killed. Some may be snapped or sheared above the root crown leaving an *in-situ* stump with root stock. Others may be completely uprooted and broken trunks or complete trees may be carried some distance downvalley by the glacier before being deposited. If one can demonstrate that the cause of death is undeniably glacial, via shearing or uprooting, then the death dates of these trees correspond to a period of glacial advance.

The information that can be obtained from trees killed by the glacier depends on the location of the tree relative to its growth site. If the tree remains are in growth position, i.e. the same position as when it was living, it is referred to as *in-situ*. Death dates from this type of evidence provide specific information concerning the position of the glacier at a particular time. However, if the remains are found on the surface of a feature or from within a deposit and the original growth location is not precisely known (Luckman 1996b), this material is referred to as detrital material. As the location where the tree was killed is unknown, the position of the ice front at that time cannot be precisely determined; so death dates from this type of evidence only indicate the timing of the advance. Therefore, when using trees killed by the glacier, *in-situ* remains provide both the time and place of a glacier advance whereas detrital material only provides information on the timing and minimum upvalley position of the advance.

4.3.2.4 Trees in the Forefield

As the glacier front recedes, new surfaces are continuously exposed and may be progressively colonised by trees. This implies that trees will colonise moraines in the order in which they were exposed so that older trees will be found on older surfaces. When using dendrochronology to determine moraine emplacement dates, the germination date of the oldest tree provides a minimum estimate of the age of the surface and therefore the landform (Heusser 1956, Luckman 1986, McCarthy & Luckman 1993, Luckman 1996b). Both living trees and snags can be used in this application providing the snags are crossdated.

There are several problems encountered using dendrochronology to date surfaces in glacial forefields (Luckman 1986, Luckman 1996b). Firstly, it is assumed that the oldest tree has been sampled. However, the oldest tree on a landform is often difficult to find on densely forested features. Although there is no guaranteed method for ensuring that the oldest tree has been sampled, tree size, morphology, branch spacing, bark characteristics and simple common sense are useful. Sampling a large number of trees also increases the possibility of finding the oldest tree. Generally, since the forefields encountered in this study were sparsely vegetated, the majority of trees on a particular feature could be sampled, reducing this source of possible error. However, if the oldest tree has not been sampled, the age of the moraine can be significantly underestimated (Luckman 1986).

The second major limitation is that the time between surface abandonment and tree colonisation is unknown. To date the surface accurately, this period of time should be added to the age of the oldest tree. This ecesis interval is difficult to estimate as it varies between and within sites depending on the environmental conditions (Table 4.2). Values typically range from ca. 10-20 yrs. but may be as long as 90 yrs. if substrates or local microclimates are not suitable (McCarthy & Luckman 1993). Several techniques have been employed to calculate this interval, the most common of which is to date trees growing on surfaces of known age. However, it is difficult to replicate the conditions present following formation of the oldest moraine. Documentation of former ice front positions is usually limited to the 20th century when glaciers were retreating rapidly and

Table 4.2

Engelmann Spruce Ecesis Values Estimated at Glacier Forefields in the Canadian Rocky Mountains (adapted from McCarthy & Luckman 1993)

Glacier	Reference	Estimated Ecesis	Dating Control
Angel	Heusser (1956)	10	Map
Saskatchewan		10	N.S.
Yoho		10	N.S.
Robson		12	N.S.
Southeast Lyell		12	N.S.
Freshfield		12	N.S.
Peyto		12	N.S
Bow		14	N.S.
Columbia		17	N.S.
Hilda		17	N.S.
Athabasca		17	N.S
Dome		17	N.S
Yoho	Bray and Struik (1963)	23-43	Photos
		20	
		25-26	
Hector	Brunger *et al.* (1967)	>27	
Peyto		>50-60	Photos, Map
Drummond		>80	Photos, Map
Angel	Luckman (1977)	15-30	Photos, Map
Bennington	McCarthy (1985)	9-20	
		13-40*	Photos
		11->61^{-}	
Athabasca -	Luckman (1988)	ca. 47	Tilted tree
lateral		ca. 50	
- terminal			
Dome - terminal		ca. 41	Scarred tree

* calculated using Pine, $^{-}$ calculated using Alpine fir, N.S. = not stated

conditions were generally warmer than the time of moraine formation (Luckman 1986). Ecesis intervals during earlier periods may have been longer as a result of cooler conditions. An additional complication for sites near the limit of tree growth is that colonisation may not be immediately after abandonment by the glacier but result from a subsequent climate change resulting in a shift in treeline (Luckman 1996b). Differences in substrate may also significantly impact the rate of colonisation on individual moraines. Blocky moraines, which are frequently the oldest, are a less favourable substrate for establishment than younger, finer texture moraines (Luckman 1986).

4.3.2.5 Determination of the Ecesis Interval

Sigafoos and Hendricks' (1969) classic study on seedling establishment in glacier forefields calculated the ecesis interval for a variety of surfaces near Mount Rainier, Washington by subtracting the age of the tree from the number of years that the seedbed had been exposed. The dating control in this study was obtained from historical photographs. This is considered the conventional method to determine ecesis intervals and has been widely employed in the Canadian Rocky Mountains (e.g. Heusser 1956, Luckman 1977).

Different methods have been used to date surface stabilisation. These include comparing the germination date of trees with independent estimates of moraine age determined using lichenometry (e.g. Burbank 1981) or growth suppression in trees beyond the forefield. Unfortunately, the error terms inherent in those techniques compound the total error for the moraine date. Another technique proposed by Luckman (1988) involves dating the moraine using trees damaged or killed by moraine emplacement. Unfortunately, since these situations are rare, it is unlikely that this technique will receive widespread application.

4.3.3 Radiocarbon Dating

Radiocarbon dating of organic material (usually wood) associated with LIA moraines has often been used as a dating technique. Since atmospheric ^{14}C has varied substantially over the last 1000 years, there is a non-linear relationship between calendar years and radiocarbon years (Stuiver *et al.* 1993). This results in error terms of ca. ±50-100 yrs. Given the well developed dendrochronological network in this area, dendrochronology is now the preferred technique in this time frame. Therefore, ^{14}C dating is only used when material does not crossdate, is too small or is beyond the range of dendrochronological dating.

4.4 Data Collection, Preparation and Analysis in the Present Study

In this study, several techniques were used to reconstruct glacial histories. Following a brief description of the mapping process, specific techniques used to date to the various glacial features identified in the forefields will be discussed.

4.4.1 Mapping

In the course of this study, relevant glacial landforms were identified largely through field investigations and mapped onto an appropriate aerial photograph prior to producing a base map. For the purpose of field mapping, air photos must be clear, well illuminated and of large scale but not necessarily the most recent. Following an inventory of the available photography, no suitable single set of photos that covered all sites was found, but photos with a nominal scale of 1:20,000 taken in 1968 and 1976 were selected as base maps for field work (Table 4.3). These were enlarged to scales ranging from 1:6,000 - 1:12,000 for use in field mapping. In the individual photos, the area under investigation was sufficiently small that the amount of distortion due to the enlargement process was negligible. Using the information obtained from field investigations, maps produced by previous researchers (e.g. Field 1949, Heusser 1956), existing topographic maps and additional aerial photographs, pertinent features could be identified on the field base maps. Once identified, these features were traced onto separate enlargements with scales ranging from 1:4,000 to 1:12,000 and then used to produce digital base maps with scales from 1:7,300 to 1:13,200.

Table 4.3

Aerial Photography Used In This Study

Flight Line	Photo Numbers	Date
A11729	12, 13, 22-24	1948
A12342	225, 366, 367	1948
A13068	221-223	1950
A14895	24-27, 69-71, 73-75, 117, 118	1955
A15535	29, 30	1956
A19684	1, 29, 66, 67	1966
A19685	104-106, 110, 111, 170, 171,	1966
A20887	142, 143[5], 144	1968
A20888	8, 9[3]	1968
A23010	196, 197	1970
A23011	167, 168	1970
A23015	37, 38, 40, 41	1970
A21540	186, 187, 188	1970
A23885	47, 48	1974
BC7876	61[1], 62, 196, 197[4], 198	1976
BC7877	56, 57[2], 58	1976
A40065	7-9, 26-30, 39-41, 62, 65-69	1977
A31609	6, 7, 25, 26, 37-39, 48-50	1992

Superscripts 1-5 indicate the photo used to produce a base map; 1 - Castleguard, 2 - Columbia, 3 - Kitchener, 4 - Saskatchewan, 5 - Stutfield

In addition to the production of base maps for landform mapping, the aerial photographs identified in Table 4.3 were also used to document recent ice front positions. For the purpose of ice front mapping, the most recent photo (1992) with the glacier at its furthest upvalley position was used as a base map. Those selected were photographically enlarged to scales ranging from 1:15,900 to 1:32,900. Using the relevant photos listed in Table 4.3, the corresponding ice front positions were sketched onto the enlargements using landmarks such as peaks, ridges, large boulders, prominent avalanche tracks, etc. as reference points. This method results in a considerable margin of error in the position of the ice fronts, possibly as high as 50 m. Therefore, the amount of retreat calculated using these maps are rough estimates.

4.4.2 Dating

Since the primary purpose of this research project was to date the LIA glacial maximum at the various sites and assess synchroneity, trimlines and outermost moraines were the primary target search areas. Approximately 600 living and dead trees were sampled over two field seasons, the location of each being noted on the appropriate base map. Sample identification followed standard UWO lab procedure; each sample label consisted of a single letter identifying the forefield from which the sample was taken (Table 4.4), followed by 4 digits. The first two digits represent the year the tree was sampled while the last two identify the exact sample. Since the majority of sites were sampled during a single field season, the year identification is omitted on summary maps. Within glacier forefields, the majority of samples were from living trees, although standing snags were also sampled. Search strategies on trimlines and outer moraines were focused on the sampling and recovery of evidence from damaged, tilted and killed trees, usually snags. However, this material was sparse and living trees were also sampled at these sites.

Table 4.4

Glacier Identification Key

Code	Glacier
C	Columbia
G	Castleguard
K and X	Kitchener
M	Manitoba
S	Saskatchewan
T and Y	Stutfield

4.4.2.1 Sampling of Living Trees

Living trees were sampled to obtain the oldest tree on a particular feature. This section reports the field and laboratory procedures used to collect and process this information.

4.4.2.1.1 Field Collection

Four hundred seventy three living trees were sampled to determine the maximum age for each using 16-inch Suunto increment corers with 5 mm diameter bores. The trees were cored as close to the base of the tree as possible and sampling height was recorded for all samples. However, McCarthy *et al.* (1991) found that errors due to sampling height were minor (1-4 yrs.) compared to those due to variation in germination dates for samples taken from within 10 cm of the root crown. Since the majority of the samples were taken from within this zone, a correction factor for sampling height was not applied in this study.

Pith dates are required to give the best estimates of establishment dates. At each tree selected for coring, the following procedures were carried out. Each core was examined in the field to determine a) the approximate age of the tree and b) whether the core contained or was near the pith. If the tree was young compared to others sampled on that feature, or if it contained the pith, no further cores were extracted. If the tree was old relative to others sampled on the feature and/or the core was significantly off pith, additional cores were taken until a pith date could be determined with some certainty. The position, coring height, species, basal diameter, tree height, approximate age and any other significant features of each tree cored were recorded. The cores were labelled and sealed in plastic straws for transport to the University of Western Ontario for analysis.

4.4.2.1.2 Laboratory Analysis of Cores

Each core was prepared in accordance with the procedures outlined in Stokes and Smiley (1968). Cores were placed in grooved blocks, surfaced by hand with progressively finer grades of sandpaper and polished with a soft cloth to produce a glossy finish. Each sample was sanded until the ring patterns were well displayed. All samples were counted at least twice from their outermost to the earliest observable ring with the aid of a Bausch and Lomb Stereozoom 7 microscope. Single, double and triple marks were used to identify decade, half century and century rings respectively on the cores.

After counting, distance to the pith for each sample was estimated in four categories (Table 4.5). The designations for individual samples are listed in Appendices I to VI. The appropriate correction factor was subtracted from the earliest ring to determine the estimated pith date. An ecesis interval correction (see section 4.4.2.2) was subtracted from the oldest estimated pith date to determine the age of each feature dated. In some cases, the oldest tree on a surface may not necessarily be from the sample containing the earliest dated ring because of the pith correction date.

Table 4.5

Correction Factors for Distance from the Pith

Proximity to Pith	Group Classification	Correction Factor
contains pith	P	0
< 2 yrs.	C	2
< 5 yrs.	NP	5
5-10 yrs.	OP	10
>10 yrs.	>OP	15

4.4.2.2 Ecesis Interval Estimate for the Columbia Icefield Area

Heusser (1956) determined ecesis intervals of 10-17 years for glaciers in the Canadian Rockies including Athabasca, Columbia, Dome and Saskatchewan Glaciers (Table 4.2). Based on dated ice front positions at selected glaciers, they remained as the only ecesis estimates for the Icefield area until the late 1980's. At Athabasca and Dome Glaciers, Luckman (1988) used dates obtained from tilted and scarred trees to determine dates of moraine emplacement. These dates indicated that the ecesis limits at these sites (where the moraines have a very sparse tree cover) are significantly longer than those used by Heusser (1956). With such a large discrepancy in these estimates, it was necessary to calculate an independent ecesis interval for this particular study.

One of the most important characteristics required to calculate ecesis intervals is the presence of precisely dated moraines. Columbia Glacier had both the best dating control and tree development of all the glaciers investigated in this study. Therefore, the traditional method outlined by Sigafoos and Hendricks (1969) was adopted to calculate the ecesis interval at Columbia Glacier.

In order to calculate the ecesis interval, twenty-two trees along the moraine that corresponds to Palmer's 1924 ice front position were sampled (Figure 4.1, Field 1949). Twelve trees with sufficiently wide bases to allow coring were sampled and ten narrower trees were cross-sectioned. The entire sectioned tree was retained in order to determine vertical growth rates. Each of these trees were sectioned at 1 cm intervals between the base and 10 cm *above the ground surface*. By determining the pith date for each section (Appendix I), the average growth rate for the lowest 10 cm of each tree was calculated. These rates were used to estimate the basal pith date for those trees not sectioned immediately at the ground surface. The mean growth rate for the 12 trees was 1.5 cm/yr, ranging from 0.6 to 3.5 cm/yr.

The two oldest trees cross sectioned in this study revealed pith dates of 1947 (CE7 and CE8) whereas the cores revealed much younger inner ring dates (Table 4.6). However, when corrected for proximity to pith, the ages were comparable, with the oldest cores revealing extrapolated pith dates of 1951 (CE9606 and CE9607). If all the samples are corrected for sampling height (using the growth rate determined for each individual tree for the sections and the average growth rate of 1.5 cm/yr for the cores), the oldest extrapolated basal pith date is 1940 (CE9603). This results in a minimum ecesis interval of 16 yrs.

While the absolute minimum ecesis interval was calculated to be 16 yrs. using basal dates, all the other samples taken in this study were obtained from coring heights equivalent to those used for the determination of the ecesis interval. The ecesis interval calculated is for the basal date and would require calculation of a correction for coring height. Therefore, in order to minimise the volume of calculations and sources of possible error, an ecesis interval that also incorporates the period necessary to grow to the sampling height was applied to the samples. The date of the oldest tree prior to correction for sampling height was used to calculate this value. Using the date 1947, determined for CE7 and CE8, the minimum applicable ecesis interval is 23 yrs.

An independent check on this value was undertaken at Stutfield Glacier. Along the northern lateral moraine, a reaction wood series indicated that a tree had been tilted in 1812 by the emplacement of the moraine. Approximately 2 m from the tilted tree, but directly on the moraine crest, a living tree was sampled 3 cm from the ground surface and

Figure 4.1 Palmer's 1924 ice front position at Columbia Glacier is the ridge bisecting the lacustrine sediments.

Table 4.6

Samples Obtained from Palmer's 1924 Moraine

			Cores		
Sample ID	Height (cm)	Core Height (cm)	Earliest Ring Date	Correction for Distance from Pith	Correction for Sampling Height[1]
CE9603	150	10	1960np	1955	1940
CE9606	150	6	1954c	1951	1942
CE9610	250	5	1952p	1952	1944
CE9607	NA	4	1961	1951	1945
CE9612	120	5	1953p	1953	1945
CE9602	150	10	1974	1964	1949
CE9605	130	7	1970	1960	1949
CE9608	NA	4	1961np	1956	1950
CE9609	100	4	1961np	1956	1950
CE9601	200	7	1972	1962	1951
CE9604	130		1970	1960	1960
CE9611	200	5	1976c	1973	1965

			Cross Sections		
Section ID	Height (cm)	Section Height (cm)	Pith Date	Average Growth Rate[2] (cm/yr)	Pith Date Extrapolated to Base
CE5	85	3	1950	1.8	1945
CE7	105	2	1947	1.0	1945
CE8	90	0	1947	1.4	1947
CE1	80	1	1952	0.8	1951
CE2	80	3	1961	3.5	1951
CE3	60	3	1955	0.6	1953
CE4	105	2	1956	1.6	1953
CE9	115	1	1954	1.5	1953
CE0	75	3	1958	1.4	1954
CE6	85	1	1961	1.8	1959

Cores classified following the conventions outlined in section 4.4.2.1, with the exception that c=<3 yrs. to the pith. All cross sections have pith dates.

Superscripts 1 - using an average growth rate of 1.5 cm/yr, 2 - to a height of 10 cm had a pith date of 1840. This suggests an ecesis interval of 28 yrs. for this particular location.

The two independently calculated ecesis values suggest that a period ranging from 20-30 years should be used as a representative applicable ecesis interval for this study. However, the ecesis interval determined at Columbia Glacier is probably the most favourable site (lowest elevation, greatest number of seedlings present, etc.). The appropriate value to be applied was determined on a "feature by feature" basis. If the moraine being sampled was located near a possible seed source, with an ameliorated microclimate suitable for tree growth, then an ecesis interval of 20 yrs. was used to determine surface ages. However, where the site was isolated from possible seed sources and had an unsuitable substrate i.e. blocky moraines or insufficient water supply, the upper limit of the range was applied. In the discussion, the ecesis value applied to a particular tree will be identified using the superscripts (a) for a 20 yr. interval and (b) for a 30 yr. ecesis interval.

4.4.2.3 Dating of Dead Trees

Dead material was sampled to recover evidence of tilting and/or the outer ring and pith date. In most cases, a complete radial section was obtained from each specimen using a bow saw. Where the intent was to recover kill dates the samples were taken from the portion of the tree which appeared to be the most complete and well preserved. This was to minimise the number of rings that might have weathered from the outer surface to achieve the most accurate death date. For damaged trees, it was imperative to sample the part of the tree which contained the evidence of damage, i.e. scar or reaction wood event. In the laboratory, the discs were sanded initially using a motorised belt sander, then using a Makita palm sander with progressively finer grades of sand paper until a polish of sufficient quality to count and measure the rings had been achieved.

The longest possible radii with the fewest growth abnormalities (e.g. reaction wood or breaks in the wood) were selected for measurement in order to maximise series length and the common signal between radii. Ideally, duplicate radii should be greater than 90° radially from each other. A minimum of two radii were measured on each cross section using the UWO measurement system[1]. As the date of the sample was unknown, the outer ring was arbitrarily assigned a date of 1000 AD. The software package used for measurement purposes was PJK6.DOS, developed by Paul Krusic from the Tree-Ring Laboratory, Lamont-Doherty Earth Observatory at Columbia University to be used with the Velmex measurement system (Grissino-Mayer *et al.* 1996). The data were then transferred to the VAX mainframe computers in standard Tucson format which is compatible with specialised computer programs in the Dendrochronology Program Library.

4.4.2.3.1 Principles of Crossdating

Crossdating is the procedure by which a series of rings from one tree are matched to those of another. Its major applications are to i) date unknown series and ii) verify dating between samples and identify dating and measurement errors using a known dated chronology. The master (dated) chronology is usually an average series created from several indexed ring-width series. Crossdating can be achieved by several methods; the simplest is based on visual matching of distinctive marker rings between an unknown ring series and a dated series. This "skeleton plot" method involves plotting an outline of the ring series under investigation (Stokes & Smiley 1968). However, this method requires highly sensitive species with a large number of marker rings. Unfortunately, the dominant species in the study area, Engelmann spruce (*Picea englemannii*), is unsuitable for this type of crossdating since it is a complacent species with relatively low interannual variation in ring-widths. Therefore, statistical techniques were used for crossdating.

Fritts (1963) developed the first computer program specifically for crossdating. The program calculates the correlation between ringwidth series of an unknown series and a dated chronology at all possible positions of overlap. The position with the highest correlation represents the most probable crossdate. Subsequent programs use a series of

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[1]The UWO measurement system is composed of a Bausch and Lomb high resolution camera mounted on a Leica Stereozoom 6 photo stereomicroscope, a Sony colour video monitor, a Zenith Systems 286 computer, a MicroLite FL2000 fibre optic light source, a Javelin converter, and an Absolute Zero II Acu-Rite digital encoder mounted on a Velmex Unislide measurement stage connected to a Metronics Quick Chek QC-1000-MAR which displays the measurements.

correlation coefficients to determine the appropriate crossdate (Parker 1971, Holmes 1983). Holmes (1983) used this approach to create a program called COFECHA where short segments of both the undated and the master series are compared at all locations until the best match is obtained.

4.4.2.3.2 COFECHA

The program COFECHA is part of the multi-purpose dendrochronological software package, the Dendrochronology Program Library. It was designed as a tool to identify portions of a series which may contain dating or measurement errors. First, each series is filtered with a cubic smoothing spline, then subjected to autoregressive modelling and log transformation (Grissino-Mayer *et al.* 1996). Correlations for segments of designated length, lagged a set number of years are calculated for the unknown series at every position in the master series. The ten highest correlating positions (best matches) are shown for each individual segment along with the number of years to add to the unknown series to obtain the indicated match. If the same number appears consistently in one of the highest correlating positions, the correct dating of the series is obtained by simply adding that number to the count of each ring (Holmes 1983, Grissino-Mayer *et al.* 1996).

The length of the segment to be tested is usually taken at the default value of 50 yrs. which is short enough to ensure several trials but long enough to minimise the chance of spuriously high correlations. However, with short series (<125 yrs.) or poorly dated sections, shorter segments can be used (Holmes 1983, Grissino-Mayer *et al.* 1996).

4.4.2.3.3 Crossdating Procedure

Crossdating was used in several capacities in the determination of the correct date for each sample. The various steps undertaken to obtain the correct date are outlined in Figure 4.2.

Once measured, the radii from each tree were crossdated against one another to identify possible inconsistencies in the tree. One radius from each tree was arbitrarily selected as the "master" (commonly the longest radius measured), while the other radius

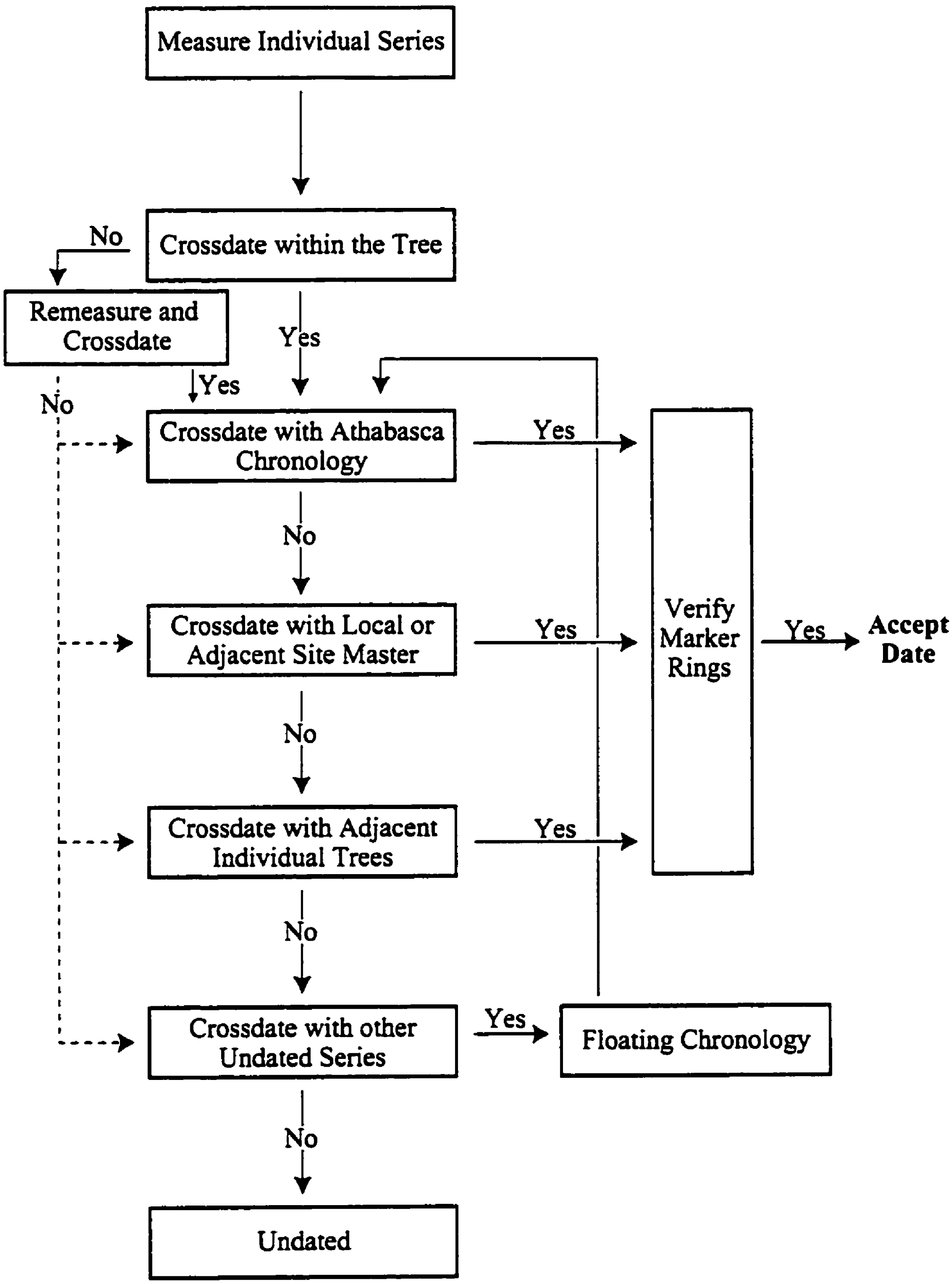


Figure 4.2 Flow Chart Describing the Crossdating Process

was input as an unknown series. Where problems were identified, each radius was re-examined until the error was located. If the problem could not be readily resolved, one or both radii were re-measured. When all attempts to match the two radii failed, both measurement series were retained for crossdating trials (Figure 4.2).

The key to successful crossdating is the strength of common interannual signal between the undated and the master series. Well replicated series usually have a stronger signal than chronologies from individual trees. Therefore, regional chronologies are the most useful for crossdating because they have better replication. However, there may be local variations in growth that are shared with other trees at the site which are not found in the regional chronology. Therefore, the strategy in crossdating is to begin with the regional chronology and, if unsuccessful, to use a series of progressively more local masters to attempt to obtain a successful crossdate.

The regional master chronology used in this study was a master ring-width chronology spanning the period 1072-1987 AD. This chronology was developed by Luckman (1993) using a composite sample of trees from sites adjacent to the Athabasca Glacier. Initially if there was no living tree chronology for the site being investigated, the initial crossdating trials used the regional (Athabasca) master. Trees that successfully crossdated with the regional master were used to develop a site master for the particular glacier forefield that may also have included living trees sampled at the site. If a tree did not crossdate with the regional master, it was subsequently correlated against the site master from the same or an adjacent forefield. If crossdating trials remained unsuccessful, the tree was crossdated against single series developed from adjacent dated and undated trees. If crossdates occurred between undated trees, floating masters could be developed which ultimately might be crossdated with calendar dated masters.

To determine the success of a crossdating trial, it is necessary to consult the output produced by COFECHA (Tables 4.7-4.9). Ultimately, a strong crossdate is obtained when the same calendar date is produced for the majority of 50 yr. segments consecutively throughout the series. This can be observed in Table 4.7. When all three radii of sample S9654 were compared to the Athabasca chronology, they indicate that the addition of 799 yrs. to each ring would result in strongest crossdate for that sample. This value was

Table 4.7

Example of a Very Strong Crossdate

Time span 1072 1987 916 years, best matches for 50-year segments lagged 25 years

Series	Counted Segment	Add # 1 Corr	Add # 2 Corr	Add # 3 Corr	Add # 4 Corr	Add # 5 Corr	Add # 6 Corr	Add # 7 Corr	Add # 8 Corr	Add # 9 Corr	Add #10 Corr	Add #11 Corr
S9654A	853 902	373 .49	314 .44	757 .41	**799** .39	494 .38	958 .35	256 .34	1076 .33	653 .33	899 .33	370 .32
S9654A	878 927	**799** .46	807 .44	749 .39	296 .37	256 .36	575 .35	874 .33	695 .33	417 .33	323 .32	500 .32
S9654A	903 952	**799** .64	621 .49	968 .46	195 .43	852 .42	323 .40	182 .39	343 .37	504 .35	634 .35	365 .34
S9654A	928 977	**799** .64	621 .54	323 .54	852 .49	634 .41	182 .40	706 .40	156 .37	877 .36	832 .34	556 .34
S9654A	951 1000	**799** .59	779 .56	814 .47	553 .46	303 .44	533 .44	419 .37	583 .36	568 .36	761 .35	518 .35

5 segments

Number of segments

Add No R_av Add No R_av Add No R_av Add No R_av Add No R_av Add No R_av Add No R_av Add No R_av

+799 5 .54 +323 3 .42

Series	Counted Segment	Add # 1 Corr	Add # 2 Corr	Add # 3 Corr	Add # 4 Corr	Add # 5 Corr	Add # 6 Corr	Add # 7 Corr	Add # 8 Corr	Add # 9 Corr	Add #10 Corr	Add #11 Corr
S9654B	853 902	373 .48	**799** .42	757 .37	314 .36	339 .36	958 .36	256 .35	970 .34	586 .33	806 .33	1013 .32
S9654B	878 927	**799** .51	807 .48	874 .42	575 .41	500 .41	484 .39	778 .37	695 .37	749 .35	365 .35	256 .34
S9654B	903 952	**799** .67	621 .48	345 .41	874 .40	365 .38	994 .37	543 .37	182 .36	778 .36	968 .34	636 .33
S9654B	928 977	**799** .68	621 .50	323 .45	441 .44	371 .43	395 .42	911 .42	182 .41	706 .38	303 .37	516 .35
S9654B	953 1002	**799** .81	553 .57	200 .50	779 .49	303 .42	814 .41	142 .39	568 .38	893 .38	518 .37	270 .34
S9654B	974 1023	**799** .71	568 .49	220 .47	746 .45	581 .40	553 .39	167 .39	304 .38	318 .38	408 .35	503 .34

6 segments

Number of segments

Add No R_av Add No R_av Add No R_av Add No R_av Add No R_av Add No R_av Add No R_av Add No R_av

+799 6 .63

Series	Counted Segment	Add # 1 Corr	Add # 2 Corr	Add # 3 Corr	Add # 4 Corr	Add # 5 Corr	Add # 6 Corr	Add # 7 Corr	Add # 8 Corr	Add # 9 Corr	Add #10 Corr	Add #11 Corr
S9654C	860 909	589 .38	225 .37	797 .37	586 .34	549 .34	438 .33	981 .33	623 .33	229 .33	1066 .32	679 .30
S9654C	885 934	**799** .60	1005 .43	527 .41	484 .39	889 .39	686 .38	666 .37	334 .37	575 .37	227 .36	614 .35
S9654C	910 959	**799** .68	968 .44	323 .44	343 .42	553 .41	262 .41	621 .41	213 .38	573 .37	963 .36	574 .35
S9654C	935 984	**799** .68	323 .65	621 .50	573 .48	553 .46	832 .44	214 .44	852 .44	289 .44	706 .40	536 .38
S9654C	960 1009	**799** .62	779 .46	976 .45	419 .45	536 .44	518 .44	686 .41	534 .41	303 .41	832 .40	196 .36
S9654C	980 1029	**799** .50	568 .45	220 .43	419 .42	715 .40	891 .37	540 .36	733 .36	144 .35	290 .35	887 .33

6 segments

Number of segments

Add No R_av Add No R_av Add No R_av Add No R_av Add No R_av Add No R_av Add No R_av Add No R_av

+799 5 .62

Table 4.8

Example of an Acceptable Crossdate

Time span 1072 1987 916 years, best matches for 50-year segments lagged 25 years

Series	Counted Segment	Add Corr # 1	Add Corr # 2	Add Corr # 3	Add Corr # 4	Add Corr # 5	Add Corr # 6	Add Corr # 7	Add Corr # 8	Add Corr # 9	Add Corr #10	Add Corr #11
K9659A	699 748	611 .44	429 .39	400 .36	1208 .34	1233 .34	374 .34	756 .33	513 .33	624 .33	850 .32	583 .32
K9659A	724 773	407 .41	535 .40	673 .37	896 .36	587 .36	610 .36	431 .34	1184 .34	1098 .34	374 .34	1031 .34
K9659A	749 798	1098 .38	842 .37	946 .37	673 .34	1183 .34	467 .34	963 .32	1188 .31	631 .31	501 .31	512 .31
K9659A	774 823	456 .43	842 .37	1096 .35	670 .33	689 .31	675 .31	817 .31	1065 .30	393 .30	1044 .30	369 .30
K9659A	799 848	909 .41	277 .39	453 .36	857 .36	322 .35	973 .35	815 .34	430 .34	894 .34	646 .33	625 .33
K9659A	824 873	755 .47	535 .45	1061 .41	**892** .41	646 .40	909 .38	309 .38	986 .36	383 .36	815 .35	973 .34
K9659A	849 898	755 .47	**892** .40	537 .39	1089 .38	491 .38	652 .37	416 .37	521 .37	962 .36	1059 .36	553 .36
K9659A	874 923	537 .49	278 .42	670 .40	322 .40	1001 .38	814 .38	652 .37	231 .36	1059 .36	574 .35	590 .34
K9659A	899 948	1001 .52	713 .49	330 .47	687 .44	859 .43	949 .41	594 .39	**892** .36	523 .35	613 .33	251 .32
K9659A	924 973	687 .48	949 .47	163 .47	304 .46	523 .41	352 .39	780 .39	330 .37	469 .35	**892** .35	859 .35
K9659A	949 998	833 .50	304 .42	780 .42	376 .41	315 .39	254 .39	137 .38	359 .36	**892** .35	163 .34	735 .34
K9659A	951 1000	Lag from prior segment 2 years - Insufficient										

11 segments -

Number of segments

Add No R_av	Add No R_av	Add No R_av	Add No R_av	Add No R_av	Add No R_av	Add No R_av	Add No R_av
+892 5 .37							

Series	Counted Segment	Add Corr # 1	Add Corr # 2	Add Corr # 3	Add Corr # 4	Add Corr # 5	Add Corr # 6	Add Corr # 7	Add Corr # 8	Add Corr # 9	Add Corr #10	Add Corr #11
K9659B	699 748	1098 .40	584 .39	934 .38	833 .35	402 .35	481 .34	852 .34	1182 .32	611 .32	1108 .32	560 .31
K9659B	724 773	430 .42	450 .41	407 .41	373 .39	535 .37	559 .37	431 .37	1030 .35	560 .35	673 .34	729 .34
K9659B	749 798	764 .57	1010 .45	501 .44	1030 .39	745 .37	865 .37	990 .37	390 .35	846 .34	450 .34	330 .34
K9659B	774 823	711 .39	411 .39	838 .38	817 .37	412 .37	990 .35	709 .35	310 .34	812 .34	1010 .34	390 .34
K9659B	799 848	917 .38	857 .38	669 .37	406 .37	350 .35	1046 .33	909 .33	838 .33	886 .32	1018 .32	590 .30
K9659B	824 873	350 .52	646 .50	461 .42	**892** .40	272 .39	909 .39	683 .38	535 .36	504 .36	578 .35	1073 .34
K9659B	849 898	**892** .45	1073 .44	272 .39	488 .37	293 .37	347 .36	831 .35	1019 .35	353 .35	591 .34	565 .34
K9659B	874 923	**892** .50	857 .41	518 .40	383 .40	646 .38	544 .36	859 .36	633 .36	1039 .35	718 .35	1059 .35
K9659B	899 948	544 .54	**892** .48	646 .42	331 .41	512 .38	560 .38	528 .37	383 .37	914 .36	633 .34	297 .34
K9659B	924 973	850 .45	813 .45	476 .44	661 .42	261 .38	396 .34	321 .33	840 .32	509 .32	**892** .32	671 .32
K9659B	942 991	193 .61	850 .50	476 .45	632 .41	661 .41	**892** .36	840 .35	594 .34	721 .34	523 .34	534 .33

11 segments -

Number of segments

Add No R_av	Add No R_av	Add No R_av	Add No R_av	Add No R_av	Add No R_av	Add No R_av	Add No R_av
+892 6 .42	+560 3 .35	+646 3 .43					

Chronological order

Add No	Add No	Add No	Add No	Add No	Add No	Add No	Add No	Add No	Add No	Add No	Add No	Add No	Add No
+560 3	+646 3	+892 6											

Table 4.9

Example of an Inconclusive Crossdate

Time span 1072 1987 916 years, best matches for 50-year segments lagged 25 years

Series	Counted Segment	Add Corr # 1	Add Corr # 2	Add Corr # 3	Add Corr # 4	Add Corr # 5	Add Corr # 6	Add Corr # 7	Add Corr # 8	Add Corr # 9	Add Corr #10	Add Corr #11
C9652A	842 891	936 .53	407 .51	883 .46	899 .46	718 .46	373 .42	357 .42	240 .41	296 .39	803 .37	882 .37
C9652A	867 916	408 .43	340 .42	902 .41	538 .40	671 .38	569 .38	964 .37	866 .35	477 .34	434 .34	282 .33
C9652A	892 941	927 .43	310 .38	836 .38	655 .37	989 .35	757 .34	234 .34	282 .34	805 .34	709 .33	205 .33
C9652A	917 966	177 .43	837 .40	205 .40	509 .37	156 .36	234 .36	820 .36	1014 .35	343 .35	344 .34	473 .34
C9652A	942 991	574 .45	465 .44	200 .42	845 .42	472 .41	190 .41	182 .36	783 .36	776 .36	130 .36	369 .36
C9652A	951 1000	Lag from prior segment 9 years - insufficient										

5 segments; no pattern found

Series	Counted Segment	Add Corr # 1	Add Corr # 2	Add Corr # 3	Add Corr # 4	Add Corr # 5	Add Corr # 6	Add Corr # 7	Add Corr # 8	Add Corr # 9	Add Corr #10	Add Corr #11
C9652B	842 891	936 .60	407 .54	899 .47	884 .45	373 .45	718 .43	297 .41	240 .40	671 .39	357 .39	995 .36
C9652B	867 916	786 .42	297 .40	321 .39	569 .39	1047 .37	1011 .37	917 .35	1032 .35	656 .35	406 .33	838 .33
C9652B	892 941	1031 .41	716 .40	946 .37	557 .36	952 .34	608 .34	573 .33	310 .33	248 .33	838 .33	669 .32
C9652B	917 966	273 .46	473 .40	946 .40	211 .36	239 .36	190 .35	511 .35	294 .35	837 .34	999 .34	551 .33
C9652B	942 991	190 .48	649 .41	776 .38	574 .37	637 .37	862 .36	257 .35	629 .35	130 .34	872 .33	222 .33
C9652B	943 992	Lag from prior segment 1 years - insufficient										

5 segments; no pattern found

generated as one of the "top eleven correlating" positions for all but one segment (the period 860-909 on radius C), and the highest correlating position for 81% of the segments. Unfortunately, this type of occurrence is rare. More frequently, situations such as those observed in Table 4.8 occur where the same value was produced at several locations in the record, with correlations higher than the next possible date. In this case, +892 is only selected as a first choice for the 849-923 interval on radius B of K9659 but is selected at other positions for other segments and also from trials with radius A. In addition, it has the best average correlation for both radii, 0.4, statistically significant if $r \geq 0.3281$. Although stronger correlations between the dated and the undated series can be observed if other values were added (e.g. the segment 749-798 on radius B indicates that a correlation of 0.57 will be achieved if 764 is added to the undated series), it should be recognised that many of these values appear only for an individual segment. It is more important that the same value is obtained for several segments than a high correlation is obtained for an individual segment. Finally, if there was no pattern in the highest correlating positions for all the segments, the crossdate was deemed unsuccessful (Table 4.9).

Ultimately, the same crossdate had to be obtained from all radii of a particular tree, although only one needed a high correlation to be accepted. Where tentative crossdates were produced, weak correlations in all radii or a crossdate only obtained from one radius, the sample was re-examined to evaluate the quality of the series, i.e. to identify any ring distortion within the series or the presence of a reaction wood event. The calendar date resulting from the series with no aberrations present was more prone to acceptance than those with distorted rings or reaction wood. However, if conflicting crossdates were produced and none of the radii contained any growth aberrations, none of the prospective crossdates were accepted. As a final test, regardless of the strength of the crossdate obtained, each sample was checked for the presence of major marker (narrow) rings known from the regional master chronologies. Between 1700 and 1850, for example, these narrow rings are 1701, 1723, 1746, 1762, 1779, 1799, 1814, 1824, 1832 and 1844. If the majority of the appropriate markers were not present, the crossdate was not accepted in the suggested crossdating position.

The quality of the crossdate depends on the strength of the common signal between two series which, often, is predicated by climate. Acceptance of a crossdate is based on the matching to a regional, site or local chronology that has been previously established. Absence of a successful crossdate is an inconclusive result which provides no information about the age of the sample. However, it should be recognised that an inconclusive result does not necessarily mean that the undated sequence is different in age than the master being used.

4.5 Summary

In order to reconstruct an accurate chronology of glacial events, it is necessary to consider both the spatial and temporal context of the evidence. After the creation of a landform map, the ages of pertinent features are required. These may be determined using several techniques which vary considerably in temporal resolution. However, it is the type of evidence available which determines the best dating method to be applied and therefore controls the resolution obtained. While historical documentation is the most precise, it has a limited applicable range and is not available for events that occurred prior to the 20th century. Dendrochronology is most effective over the last 3-400 yrs., but varies in precision depending on the type of evidence available. Kill and damage dates usually have superior precision over ecesis based estimates. However, the presence of suitable evidence does not guarantee a successful outcome. Crossdating may be inconclusive because the length and nature of the preserved record is inadequate.

Despite the limitations of dendrochronology, it remains one of the most accurate and frequently employed techniques to date glacial events in alpine environments. In the present study, the three types of dendroglaciological evidence: minimum surface ages, damage to living trees and death dates for trees were used to reconstruct the glacial histories at five sites at the Columbia Icefield.

CHAPTER V
CASTLEGUARD GLACIER

5.1 Introduction

The Castleguard Glacier is the most southerly glacier examined in this study and is part of the Castleguard glacier complex located at the south end of the Columbia Icefield. Unlike the other glaciers investigated in this study, the Castleguard Glacier is the only glacier not confined to a deeply incised valley. On either side of the glacier there are minor icefields that are presently detached from the Columbia Icefield. However, it appears that the small icefield to the south was connected to the main Icefield during the LIA, and coalesced with Castleguard Glacier to form a wide lobe extending almost 4 km downvalley from its present position. At its maximum position, the glacier to the north appears to have expanded to within 800 m of the Castleguard Glacier forefield.

The Castleguard Glacier gradually descends from the Icefield plateau, without an icefall, to a toe elevation of 2000 m.a.s.l.. Most of the 8.2 km^2 area (based on 1992 aerial photography) was below the 2450 m.a.s.l. 1992 snowline (Robinson 1996) which possibly accounts for the gently sloping, retreating front (Figure 5.1). Meltwater flows from the terminus into a small proglacial lake carved into the bedrock that was first evident in aerial photographs in 1966. The single stream draining this lake is the main surface source of the Castleguard River and is confined to a bedrock gorge that bisects the forefield (Figure 5.2).

The Castleguard Glacier forefield lacks conspicuous glacial landforms, but most of the forefield has a thin veneer of till over bedrock slopes, creating an unsuitable environment for tree growth and resulting in sparse vegetation cover. Distinct moraine ridges are concentrated at the eastern end of the forefield near the former glacier limits and the central and western portions are almost exclusively bedrock (Figure 5.3). On the north side of the valley, the moraines are low (1-2 m high), discontinuous ridges, running parallel to the downvalley axis of the forefield. At their upvalley limit, approximately 2.5 km from the present ice front they are truncated by debris flow deposits (Area A, Figure 5.3) originating from the glacier to the north. On the western edge of these

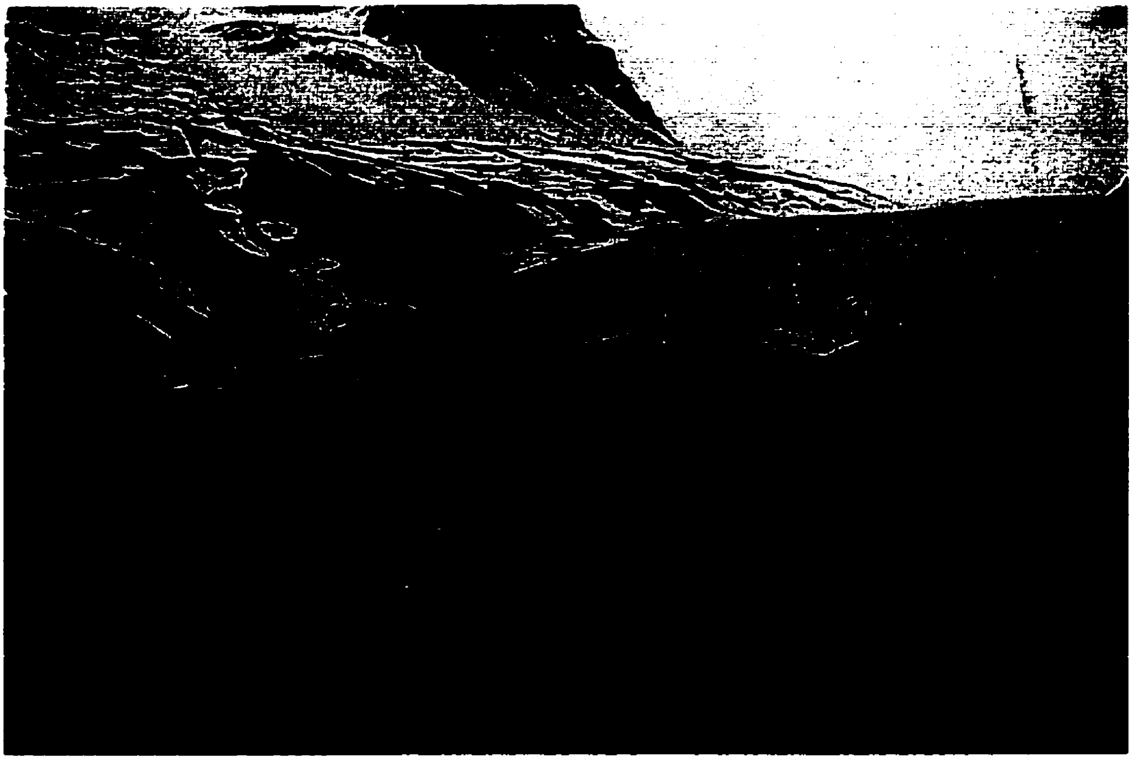


Figure 5.1 The Castleguard Glacier (Photo BJR, August 2, 1996).

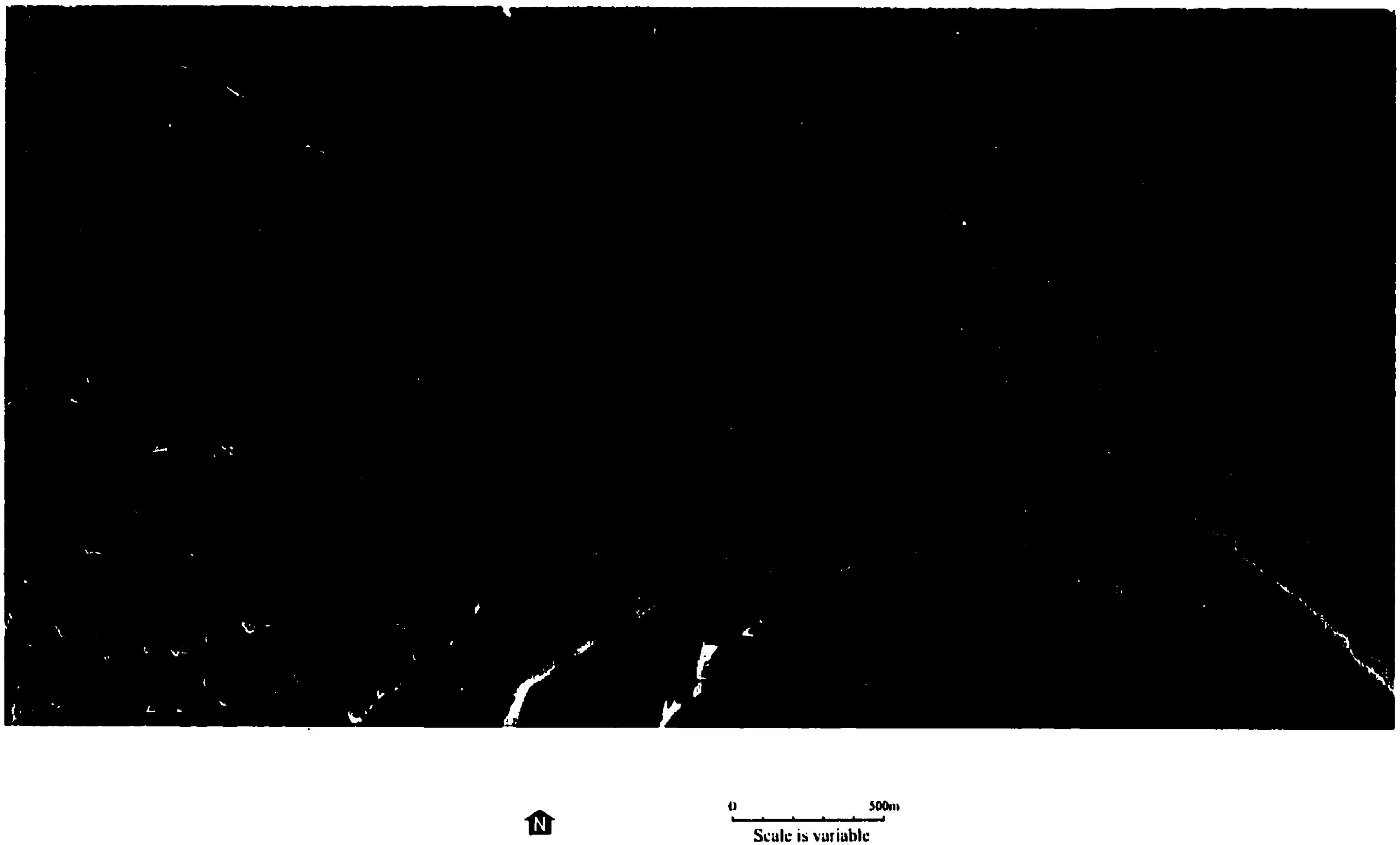


Figure 5.2 Aerial Photo of the Castleguard Glacier site (Province of British Columbia, Photo Number BC7876-61, Sept., 1976).

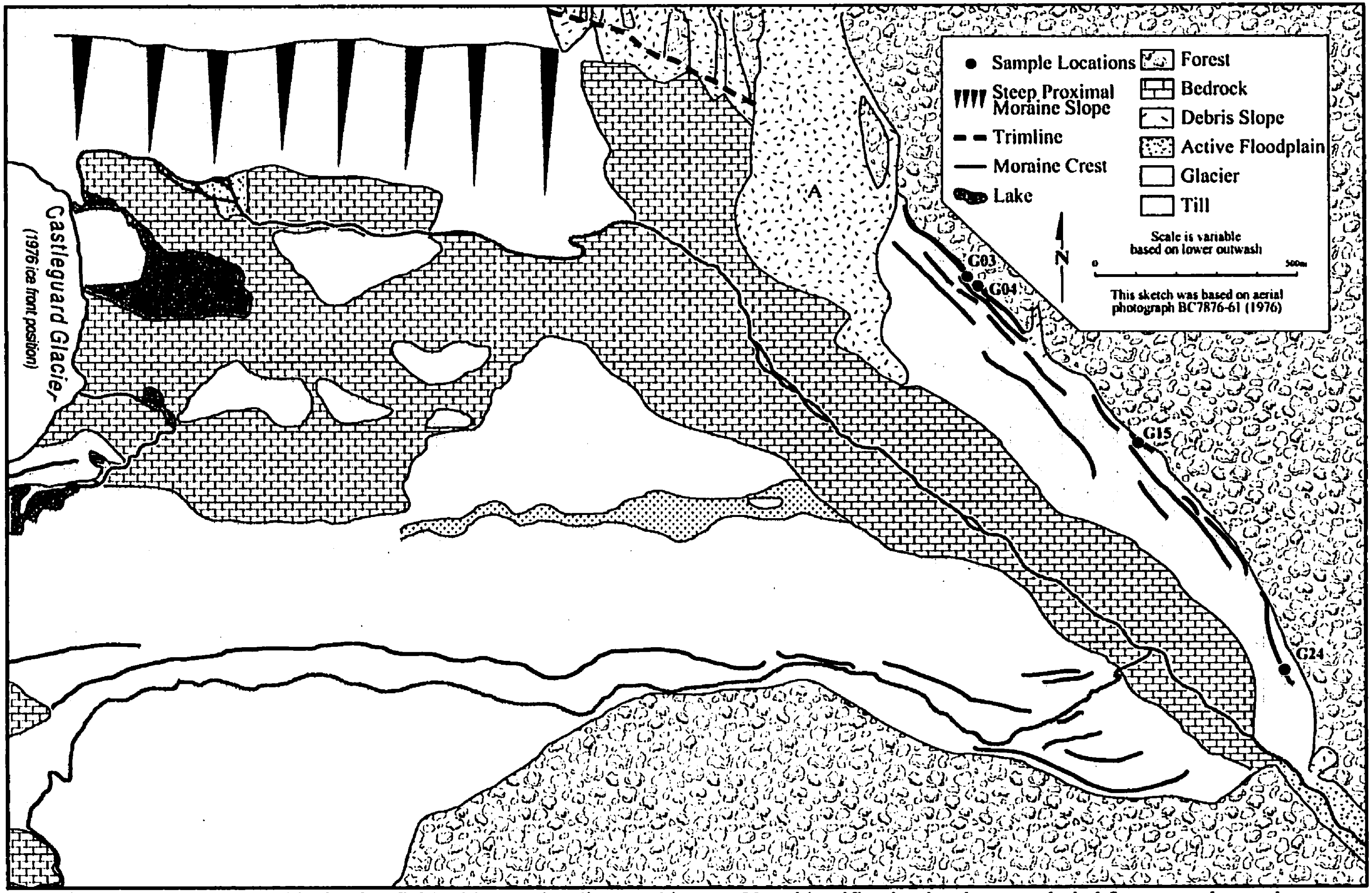


Figure 5.3 Castleguard Glacier forefield with samples discussed in text. Year identification has been excluded from sample numbers (e.g. G03=G9603).

deposits, a 600 m section of trimline dissected by abandoned meltwater channels is evident. The remaining section of the forefield nearest the glacier consists of steep valley walls and bedrock.

On the south side of the valley, a lateral/interlobate moraine feature extends from near the current ice front to within 500 m of the downvalley limit of the forefield, cutting across the forefield of the glacier to the south (Figures 5.2 and 5.3). One of the main streams draining the glacier to the south runs parallel to the distal side of this moraine along its entire length. On the east side of this stream at the downvalley limit of the forefield, the moraines curve downslope into the centre of the valley.

5.2 Results

Based on preliminary analysis of aerial photos, there appeared to be a well defined trimline at this site which would offer considerable scope to provide dendrogeomorphic dates for the LIA maximum event. Field investigations however revealed this was not the case. Despite extensive searching, no tilted or killed trees were found along this trimline to provide precisely limiting dates for glacial events. Therefore, the investigation at this site was restricted to the sampling of living trees. As the steep bedrock gorge in the centre of the valley prevented access to the southern forefield and there was little vegetation near the centre of the forefield, sampling was limited to the outer moraines on the northern side of the forefield. Complete results from all samples collected at this site are presented in Appendix II.

Most of the trees sampled provide earliest ring dates in the early 20th century (Appendix II), corresponding to germination dates in the late 19th century. Two of the oldest trees were located within several metres of one another on the outer moraine near the upvalley limit of till (Table 5.1 and Figure 5.3). A tree (G9604) located midway up the proximal slope had a pith date of 1888. At the base of the distal slope, a tree growing on the moraine revealed a near pith date of 1894 (G9603). Since there is limited access to water and restricted soil development resulting from the proximity of the bedrock to the surface, the maximum ecesis interval of 30 yrs. was applied, resulting in germination dates of 1858 and 1859, respectively. This suggests a minimum surface age of 1858 for the outermost moraine. Supporting evidence for this date was found throughout the forefield. A tree growing on the distal slope of a 7 m long moraine ridge 650 m downvalley of these two trees indicates that the moraine was deposited ca. 1860 (G9615, Table 5.1). Furthermore, near the eastern end of the forefield, a tree growing on the crest of the outermost moraine indicates that the surface had stabilised by ca. 1866 (G9624). The dates obtained from the northern forefield all suggest the outermost moraine and trimline date from the mid to late 19^{th} century, ca. 1858-60.

Table 5.1

Limiting Ages for Moraines in the Castleguard Forefield

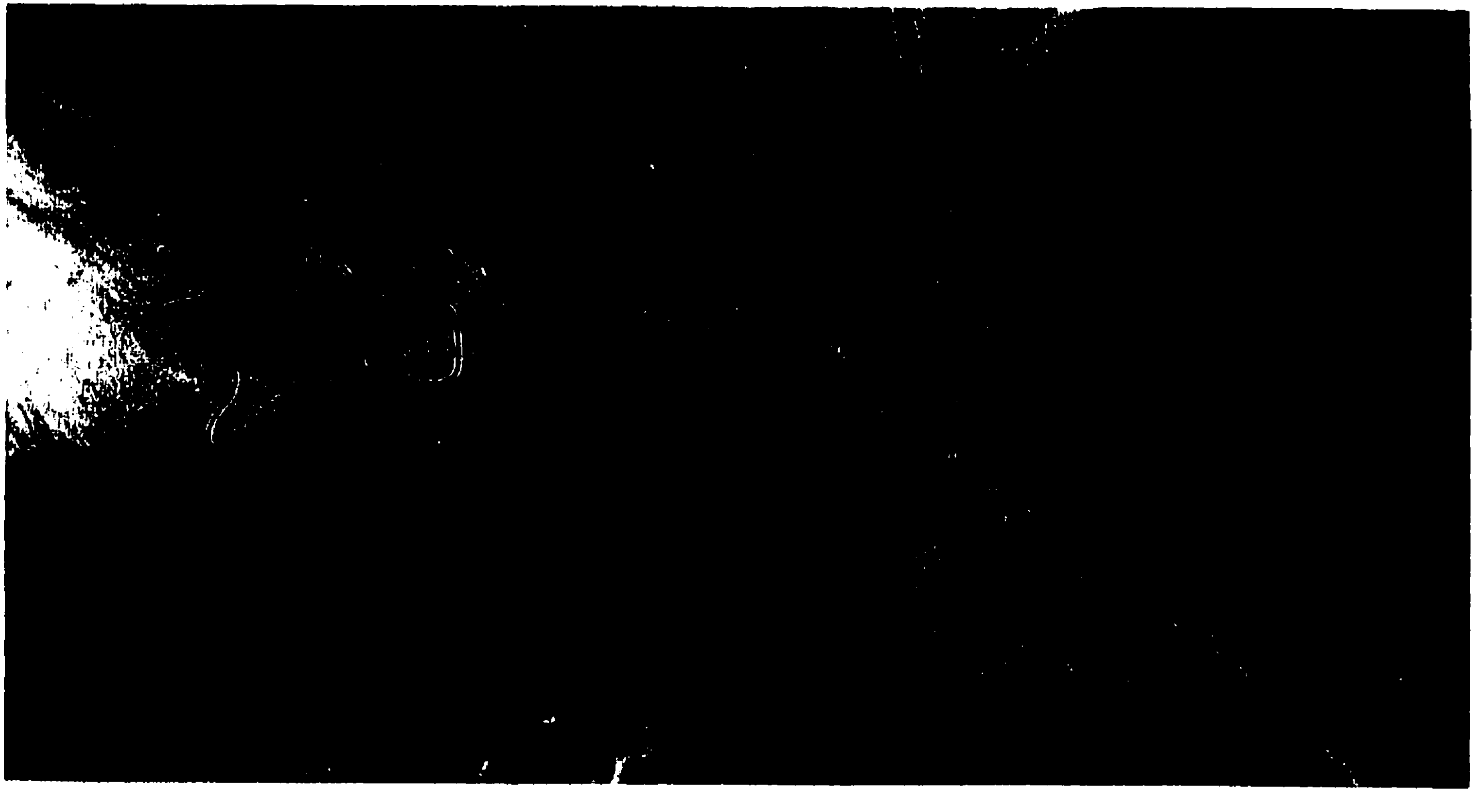
Location	Core ID	Inner Ring Date	Surface Date
250 m downvalley of the westernmost limit of moraines	**G9604**	**1888**	**1858**
250 m downvalley of the westernmost limit of moraines	G9603	1894	1859
1000 m upvalley of the maximum position	G9615	1890	1860
250 m upvalley of the maximum position	G9624	1901	1866

Note: Sample ID's include year identification. Surface date includes an ecesis of 30 yrs. and a pith correction. Bold indicates the limiting date.

5.3 Documented Ice Front Positions

The Castleguard Glacier has undergone rapid and continuous retreat throughout its documented history. At its LIA maximum position, the Castleguard Glacier extended almost 4 km downvalley of its present position. The only historical record of the ice front position prior to the first available aerial photographs in 1955 was an oblique photo taken from Watchman Peak by the Interprovincial Boundary Survey in 1918 (Cautley & Wheeler 1924). Although partially obscured, the snout position appears to have retreated approximately 1300 m from its maximum position (Table 5.2 and Figure 5.4). Between this date and the first aerial photograph in 1955, the glacier had receded an additional 1.1 km. This general pattern of retreat continued until the period 1966-1970 when the rate of retreat reached an estimated high of 74 m/yr and remained high until 1976 at which time it returned to previous values (Table 5.2).

When compared to glaciers in this and other studies (e.g. Field 1949, Field & Heusser 1954, Luckman 1988), Castleguard Glacier has experienced the greatest amount



0 500m
Scale is variable based on lower outwash

A - 1992 C - 1970 E - 1955
B - 1976 D - 1966 F - 1918

N

Figure 5.4 Recent Ice Front Positions at Castleguard Glacier (NAPL A31609-6).

of net recession during the 19th century. Historical evidence suggests that most other glaciers remained close to their maximum positions until the early 20th century when retreat gradually increased, reaching a peak in the 1950's-60's. Castleguard Glacier differs from this pattern in that recession has proceeded at a high rate throughout its history.

Table 5.2

Recession of Castleguard Glacier

Period	Net Retreat (m)	Rate of Retreat (m/yr)
1860 - 1918	1330	-
1918 - 1955	1130	31
1955 - 1966	225	21
1966 - 1970	300	74
1970 - 1976	330	55
1976 - 1992	460	29

5.4 Summary

Unlike the other glaciers investigated in this study, there is only dated evidence for a single glacial event at this site. The limited evidence indicates that the maximum was achieved here ca. 1858. Additional moraines documenting later, minor readvances exist within this limit. Unfortunately, the lack of dateable material prevents the age determination of these features. However, photographs from the early 20th century indicate that Castleguard Glacier rapidly retreated from its maximum position and was over 1.3 km upvalley by 1917.

CHAPTER VI
KITCHENER GLACIER

6.1 Introduction

Kitchener Glacier is the name informally applied to the glacier fed by ice from the main Icefield that avalanches down the steep northeast face of Mt. Kitchener. Unlike the other glaciers discussed in this thesis, Kitchener Glacier has no direct connection with the Icefield (Figure 6.1). The glacier is a heavily debris-covered lobe dissected by supraglacial drainage channels and has an approximate area of 5 km^2 (Robinson 1996). Three-quarters of the area lies below the 2500 m 1992 snowline (Robinson 1996). The 2 km long glacier has a relatively gentle downvalley gradient and terminates at approximately 2100 m.a.s.l. (Robinson 1996), almost 400 m above the adjacent Stutfield Glacier. In the centre of the valley, the toe of the glacier gently grades into till. Along its western flanks, a near vertical ice cliff is exposed adjacent to a presently abandoned channel.

On both sides of the valley, steep, well-developed lateral moraines 50 m high extend from the present ice front to approximately 2 km downvalley (Figure 6.2). The moraine on the eastern side of the valley has a single crest that begins near the present ice front and extends approximately 1 km downvalley before splitting into two ridges. The outer ridge has sharp crested sections, approximately 5 m high and more subdued crested sections 1-2 m high. The inner ridge has a distal slope of 5 m for most of its length. However, the outer ridge is partially obscured in places by talus from the steep valley walls above. About 300 m from their downvalley limit, the distance between the two crests gradually increases from ca. 20 to 130 m and they begin to grade into low, subdued ridges that extend down to the present outwash. Furthermore, a distinct difference in lithology is evident. There is a sharp contact between the materials of the outer dark grey ridge and the inner beige ridge.

The features on the west side of the valley are slightly more complex and five ridges can be recognised (Figure 6.3). The outer two (Ridges a and b) have subdued crests approximately 10-20 m high with a well developed forest cover for most of their length. Neither of these features have equivalents on the east side of the valley. At its

Figure 6.1 Kitchener Glacier from Tangle Ridge (Photo BJR, August 7, 1996).

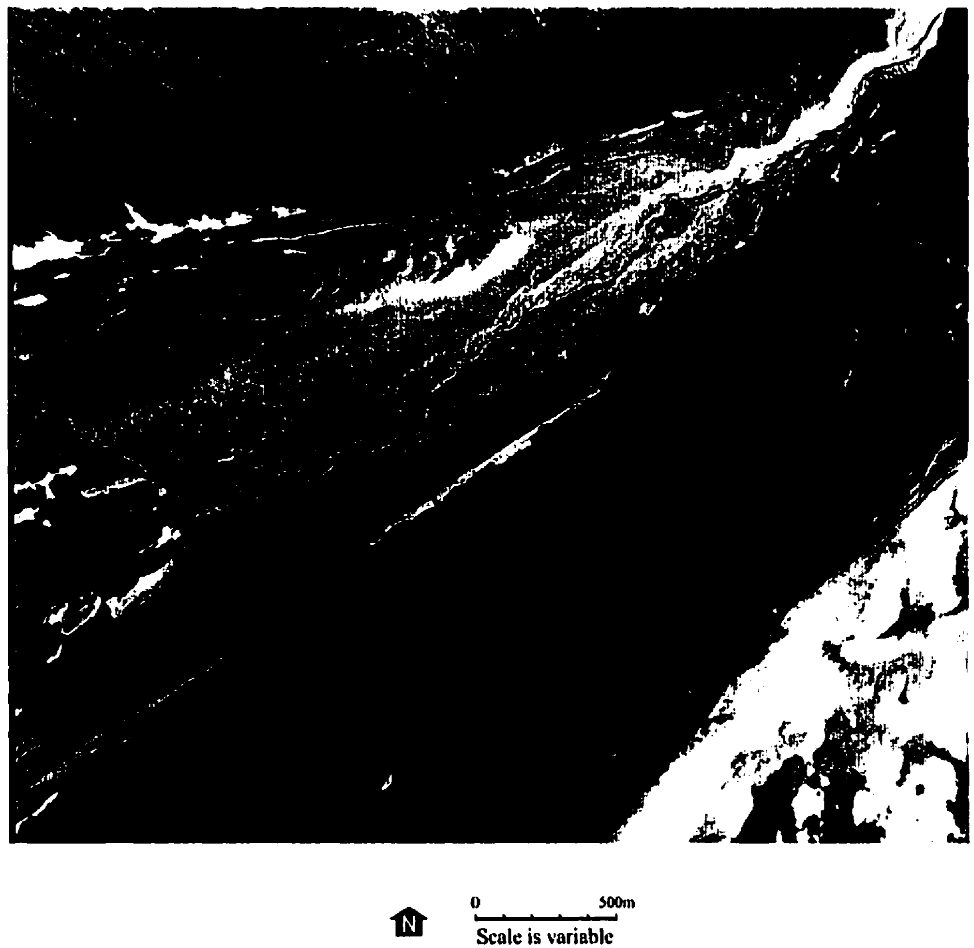


Figure 6.2 Aerial Photo of the Kitchener Glacier site (Natural Resources Canada, NAPL Photo Number A20888-9, July 1968).

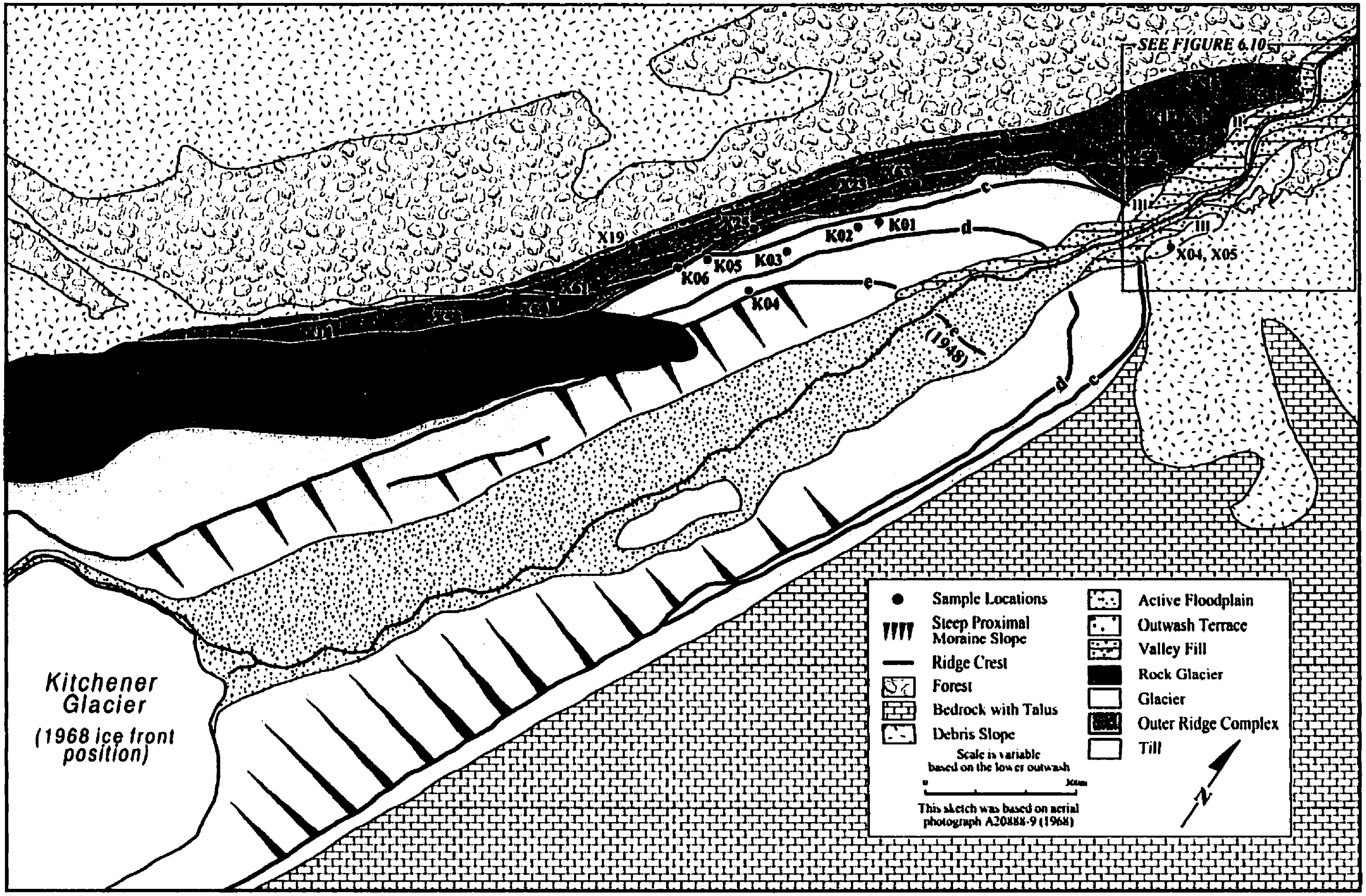


Figure 6.3 Kitchener Glacier forefield with samples discussed in text. Year identification has been excluded from sample numbers (e.g. K01=K9601).

upglacier limit, ridge b is truncated by younger rock-glacierised moraines. Downvalley ridges a and b become indistinguishable and grade into an area of undulating "kettle-like" topography downvalley of the limit of fresh moraines.

The three inner ridges (ridges c-e) can be distinguished from the outer features based on differences in substrate, morphology and vegetation cover (Figure 6.2) and they define the forefield for this glacier. The inner features are sharp crested ridges up to 15 m high that are composed of blocky material with minimal vegetation cover. At their upvalley limits, they are truncated by a 1.6 km long rock glacier along the western lateral moraine that begins above the present ice front. At their downvalley limits, the outer two unforested ridges widen and form subdued ridges that are continuous with the two moraines on the east side of the valley. The innermost feature is a small ridge with a distal slope of only several metres. Aerial photographs from 1948 and 1955 show that it formerly extended across the valley floor as a small ridge. This ridge has been subsequently destroyed by the fluvial activity of the stream in the centre of the forefield.

Downvalley of ridge d, the stream is confined to the centre of the valley for approximately 900 m and has cut down into an alluvial fill (Figure 6.3). At three locations along this section there are well developed terraces. Just downstream of the outermost unforested ridge (ridge c) there is a paired terrace about 20 m above the present stream. About 200 and 300 m downvalley of this position there are two terrace fragments 14 and 10 m above the stream respectively.

6.2 Results

The area studied at Kitchener Glacier can be divided into three different units: 1) the outer, mainly forested ridges and the hummocky terrain, 2) the inner, unforested ridges (fresh moraines) and 3) the downvalley terrace sequence. The results for each of these areas will be presented separately. Complete results from all samples collected at this site are presented in Appendix III.

6.2.1 Outer Forested Features

Beyond the fresh moraines, there are two distinct ridges approximately 10-20 m high with weathered crests (Figure 6.3). The inner ridge (ridge b) is approximately 1.6 km in total length, although approximately 100 m of the middle section has been overridden by the outermost unforested ridge (ridge c, Figures 6.4 and 6.5). The two sections of ridge b are correlated on the basis of morphology and slope angle. The upvalley portion of ridge b has minimal soil cover with only sparse patches of dryas and a few trees (Figure 6.4). Downvalley of the truncation, vegetation is initially confined to small islands on the ice proximal side and patchy forest on the distal side but grades into closed forest at its extreme downvalley limit (Figure 6.5). The outermost ridge (ridge a) is forested for its entire 1 km length and is truncated upvalley by ridge b (Figure 6.4). At their downvalley limits, both ridges grade into an area of "kettled" topography with closed depressions up to 20 m deep that is confined to the area west of the stream.

A variety of dates were obtained for these features and are presented in Table 6.1. The oldest living tree on ridge a occurs just below the crest on the proximal slope and had an inner ring date of 1722 (X9619, Figure 6.3). Since this feature was located within the current forest, it was likely that favourable conditions existed for germination and an ecesis interval of only 20 yrs. was used. Although several other trees on this feature were cored, they yielded considerably younger inner ring dates (Appendix III).

Ridge b can be subdivided into two separate units; the unforested upvalley portion of the ridge and the forested section downvalley of the truncation. The unforested portion of the ridge has a scattered tree cover, the oldest of which yielded a pith date of 1814 (X9614). However, standing snags sampled on the same surface revealed much older pith dates; namely 1576, 1637 and 1712 (X9661, X9658 and X9659). Downvalley, in the forested portion of the moraine the oldest tree sampled had a pith date of 1591 (X9663). These values indicate that ridge b was created prior to 1546 (Table 6.1).

Sampling was also undertaken in the kettled area at the limit of the forefield (Figure 6.3, Table 6.1). The range of inner dates of living trees cored was 1715 (K9616) to 1897 (K9614). However, a large log found lying on the surface just beyond the outermost unforested ridge (ridge c) was crossdated to reveal a pith date of 1543

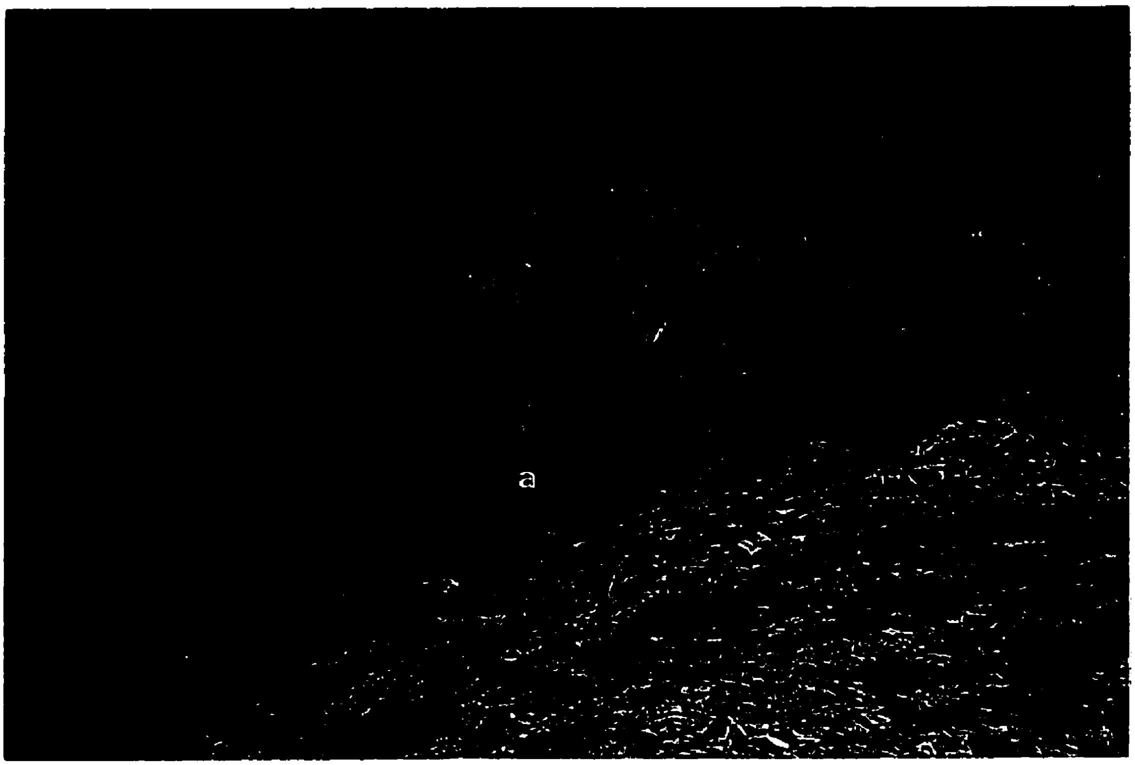


Figure 6.4 The outer ridges at Kitchener Glacier. The photo was taken immediately upvalley of the beginning of ridge a, shown as the forested ridge on the left. Ridge b is shown in the centre of the photo and ridge c is located on the right. Note how ridge b tapers into ridge c and can then be observed extending out from it further downvalley (Photo BJR, July 27, 1996).

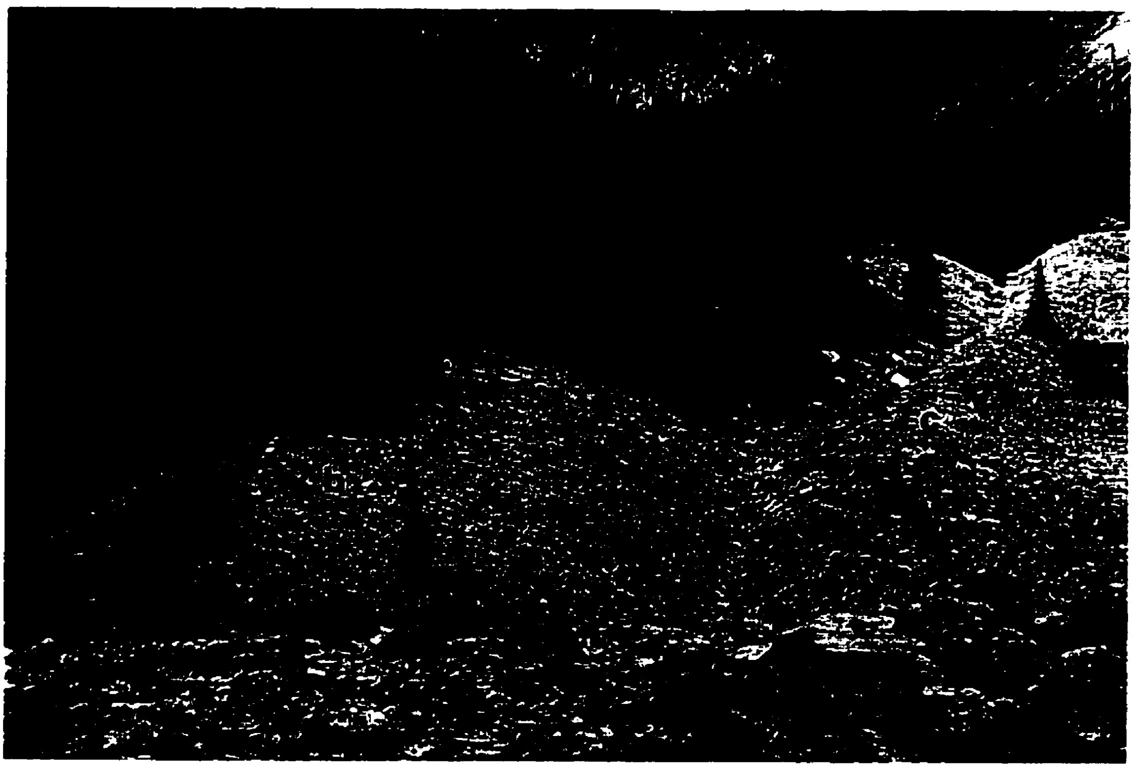


Figure 6.5 Ridges b (left) and c (right) at Kitchener Glacier. Taken from position where ridge b has been overridden by ridge c. Note the difference in vegetation cover between the two ridges (Photo BHL, July 2, 1996).

(K9650), corresponding to a germination date of 1523. Although this log was not found in growth position, it could not have been transported to this position by avalanches or the river. The antiquity of this surface is supported by pith dates from two standing snags (1595, K9658 and 1591, K9659) also located in this area. These values are consistent with the ages of the outer ridges upvalley, suggesting that all three features are of similar age.

Table 6.1

Limiting Surface Ages for the Outer Forested Ridges

	Cores			Cross Sections		
Location	Core ID	Inner Ring Date	Surface Date	Section ID	Section Range	Surface Date
ridge a	X9619[a]	1722	1692			
ridge b	X9614[b]	1814	1779	X9658[b]	1637-1995	1607
	X9623[a]	1642	1617	X9659[a]	1712-1970	1682
	X9625[a]	1761	1739	**X9661[a]**	**1576-1906**	**1546**
				X9663[a]	1591-1885	1571
Kettled Area	K9616[a]	1715	1685	**K9650[a]**	**1543-1803**	**1523**
	K9613[a]	1733	1703	K9658[a]	1595-1898	1575
	K9612[a]	1734	1704	K9659[a]	1591-1893	1571

Note: Sample ID's include year identification. Surface date includes ecesis and pith corrections; superscripts: a - an ecesis interval of 20 yrs., b - an ecesis interval of 30 yrs. Bold indicates the limiting date.

Although these features form well developed moraine-like ridges at their upvalley end, the morphology is more ambiguous downvalley. This ambiguity results in uncertainty concerning their origin. Although they appear glacial in morphology, similar ridges may also be found along the margins of landslide deposits. The kettled area beyond the limit of the forefield differs from contemporary terminal glacial deposits in other forefields with heavily debris covered glaciers because of the size of the depressions. At Dome and Stutfield Glaciers, the dead ice topography does not extend far downvalley and the large kettle holes are eventually destroyed or infilled by the active outwash.

Several explanations are suggested to account for the origin of these features; they may be: 1) purely glacial features, 2) exclusively landslide deposits or 3) some combination of both. Under the glacial hypothesis, the outer ridges are older, weathered lateral moraines and the kettled area represents an area of "dead ice" topography in the terminal area of the glacial advance. Alternatively, they may document a landsliding event that occurred either before or after the glacier had reached its maximum position. The parallel nature of the outer ridges, between each other and the fresh moraines, suggests that they were shaped by the glacier, possibly as a reworked landslide deposit. The landslide may have occurred prior to occupation by the glacier and as the glacier advanced, the material was "pushed up" along its margins. Alternatively, the ridges may represent a landslide that was deflected downvalley along the margin of the glacier. The kettled area downvalley may also be the result of reworked landslide deposits or the margin of an earlier rockslide that ran down the axis of the valley prior to the glacial event. Unfortunately, the morphology of these features is ambiguous and further detailed sedimentological investigations would be necessary to resolve this question.

Minimum dates from the surface of the outer features indicates that they predate the mid-1600's. Other glacial features of this age reported in the Canadian Rockies also have poorly constrained dating control (Luckman 1995, Smith *et al.* 1995); it is uncertain whether the trees used to date these surfaces were the primary colonisers or survivors of a much older forest. Therefore, this material may have been deposited by an event that substantially pre-dates the available dating. Sections from trees in the alluvial fill inset into the kettled area suggest an older age.

Approximately 150 m upvalley of the downvalley terminus of the kettled area, an alluvial valley fill deposit is inset into the terrain (see section 6.2.3). Two trees found about 4 m below the surface of terrace II reveal inner ring dates of 1139 and 1334 (X9651 and X9653, Table 6.2). Although the lower portion of the tree from which section X9651 was taken was rotten, its vertical orientation suggests that it was in growth position. This suggests that the surface in which they were rooted had stabilised by the early 12th century. However, the age and length of X9651 place it completely beyond the limit of the site master and during a period lacking marker rings. Therefore, although the section produced an acceptable crossdate with the regional master, the dates obtained for this sample are tentative. However, the length of sample X9653 ensures a stronger crossdate and since the sample was taken from just above the root crown it provides a closely limiting date. This results in a slightly younger surface dating to the early 14th century. Since these trees were located within the kettled area, the surface of this feature must predate at least 1334. This suggests that the outer features were created before the LIA.

Table 6.2

Outer Ring Dates for Trees in the Valley Fill

Section ID	Calendar Dates
K9653	1539-1781
K9654	1405-1783
K9656	1417-1778
K9657	1672-1794
X9651	1139-1317
X9653	1334-1704
X9656	1625-1790

Note: Sample ID's include year identification.

In an attempt to provide a more closely limiting date for ridges a and b, two soil pits were dug to observe whether tephra was present. During the Holocene, three tephras have been deposited in the Columbia Icefields region (Osborn 1993): Mt. Mazama (6850 yrs. BP), Mt. St. Helen's Yn (3400 yrs. BP) and Bridge River (2400 yrs. BP). However, since tephra was not detected on either of the ridges, the result is chronologically inconclusive.

6.2.2 Fresh Moraines

The inner ridges (ridges c-e) differ considerably in morphology and vegetation cover from the outer forested features. They are sharp crested ridges that begin at the toe of the rock glacier and are predominantly composed of boulders. They have minimal soil development and trees are limited to sheltered depressions. Therefore, a maximum ecesis interval of 30 yrs. was applied to all samples taken from this area. The outermost of these ridges, ridge c, is composed of dark grey, blocky material and is approximately 15-20 m high. It forms a 1 km long continuous ridge on the west side of the valley and is continuous with the grey ridge east of the stream. At its downvalley limit, it forms a wide

(~15 m) ridge with a subdued crest but tapers upvalley into a narrow ridge with steep slopes.

Between ridges c and d (Figure 6.3), there is an abrupt change in surface composition and colour from blocky, dark grey material to a finer grained, beige-coloured debris that comprises the inner two ridges. The outer "beige" ridge (ridge d) is a 700 m long feature that runs parallel to ridge c for its entire length and is continuous with the inner ridge on the east side of the stream. Similar to the outer ridge, the upvalley portion of this feature has a sharp crest approximately 10 m high but grades to a subdued ridge about 5 m high at its downvalley limit. The innermost feature (ridge e) is a small (1-2 m high), 400 m long ridge that has been overridden by the rock glacier at its upvalley end. The small size of this feature (compared to the other unforested ridges) suggests that this ridge was deposited as a minor readvance whereas the others represent significant readvance positions.

The almost complete lack of vegetation on these three unforested ridges allowed the sampling of all living trees on these features (Table 6.3). Only two trees were located on the crest of the outer moraine (Figure 6.3): one on the portion of the moraine that overrode the earlier moraines (K9606) and the other several metres downvalley of the truncation (K9605). However, since all the basal cores obtained from tree K9606 were twisted and/or very poor quality, none were retained. Therefore, the tree was cored 9" from the base and the earliest ring provides a limiting date of 1806. The other tree (K9605) provides a much more closely limiting date (Figure 6.6). This 2 m high flagged and skirted spruce had a severely irregular growth form which hindered attempts to obtain a core containing the pith. A pith date of 1759 was obtained from a height of 6" but a basal core provided a near pith date of 1745. This suggests that ridge c was deposited before ca. 1713. Three trees located in two small depressions between ridges c and d yield inner ring dates of 1846, 1865 and 1944 (K9601-K9603) respectively. The only other tree located on these moraines was found approximately 1 m outside the inner ridge (ridge e). A core taken 5 cm from the base gave a near pith date of 1847 (K9604, Figure 6.7). This provides a minimum limiting date of 1812 for ridge d. As there were no trees inside ridge e, a minimum age could not be determined. Although moraine e may predate

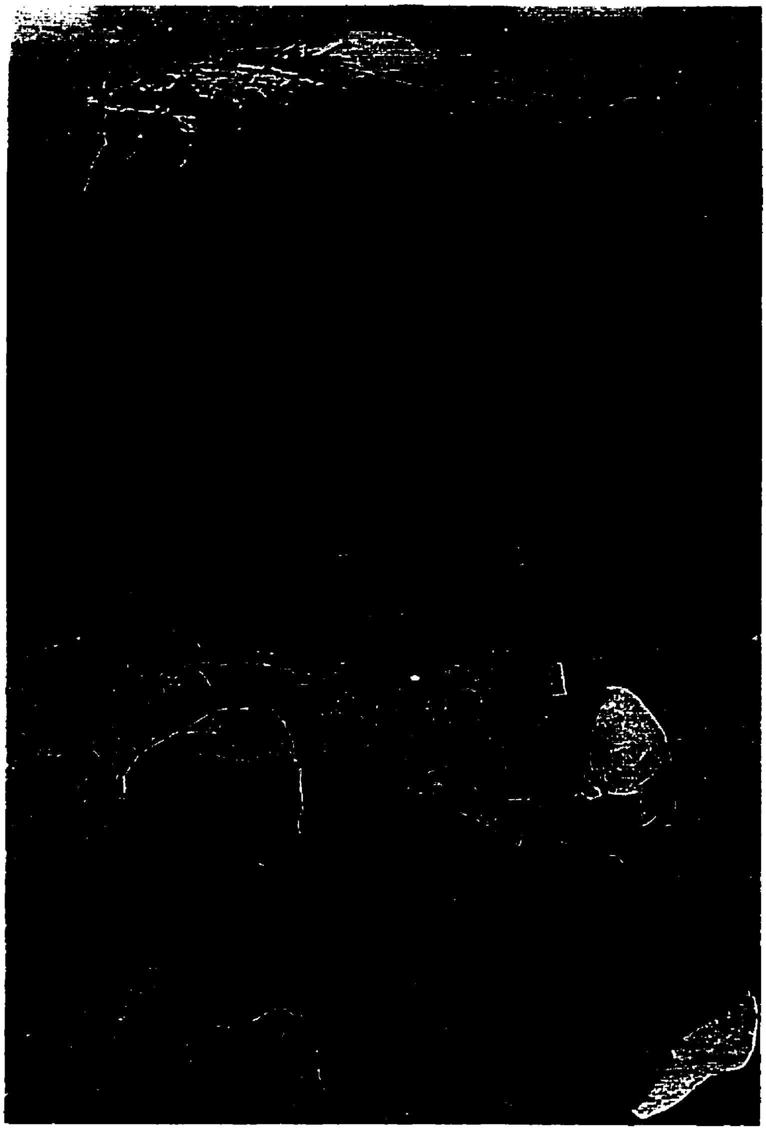


Figure 6.6 Tree K9605. Located on the crest of ridge c. Trees K9601 and K9602 can be observed in the small depression in the background. The ridge to the left is ridge b (Photo BJR, July 2, 1996).

Figure 6.7 Tree K9604. Located just beyond ridge e which is the slight ridge on the left (Photo BJR, July 2, 1996).

the germination of K9604, it is unlikely. The dates obtained for these features indicate that the two major readvances represented by ridges c and d predate ca. 1713 and between 1713 and 1812, respectively. The innermost ridge (ridge e) documents a readvance that likely occurred after 1812.

Table 6.3

Limiting Ages for the Inner Unforested Ridges and Terrace

Location	Core ID	Inner Ring Date	Surface Date
ridge c	**K9605**	**1745**	**1713**
	K9606	1806	1774
	K9601	1846	1816
	K9602	1865	1830
	K9603	1944	1904
ridge d	**K9604**	**1847**	**1812**
terrace	**X9604**	**1850**	**1820**
	X9605	1871	1831

Note: Sample ID's include year identification. Surface date includes an ecesis of 30 yrs. and a pith correction. Bold indicates the limiting date.

6.2.3 Outwash Terrace and Associated Valley Fill

Immediately downvalley of ridge c, the stream is flanked by a conspicuous paired terrace remnant with an almost horizontal surface approximately 20 m above the present stream (terrace III, Figures 6.3, 6.8 and 6.9). On the western side of the valley, this feature is sparsely vegetated whereas a grove of trees extending down from the slope above has colonised its eastern counterpart. On the eastern feature, trees sampled at the junction between the terrace and the slope above indicate that the terrace was formed prior to ca. 1820 (X9604).

Downvalley, there are two more small terrace fragments with steeper downvalley gradients than terrace III that are completely barren of surface vegetation. Furthermore, neither have counterparts on the eastern side of the valley. Terrace fragment II is approximately 14 m above the present stream while terrace I is only about 10 m above the channel. The stream section in valley fill comprising these two terraces revealed several snags in vertical, growth positions protruding from the fill (Figure 6.10). The death dates

Figure 6.8 Paired terrace just beyond ridge c (Photo BJR, July 2, 1996).

Figure 6.9 Outwash terrace on the east side of the valley located just beyond the eastern equivalent of ridge c (grey ridge). Also visible is ridge d (beige ridge) (Photo BHL, July 2, 1996).

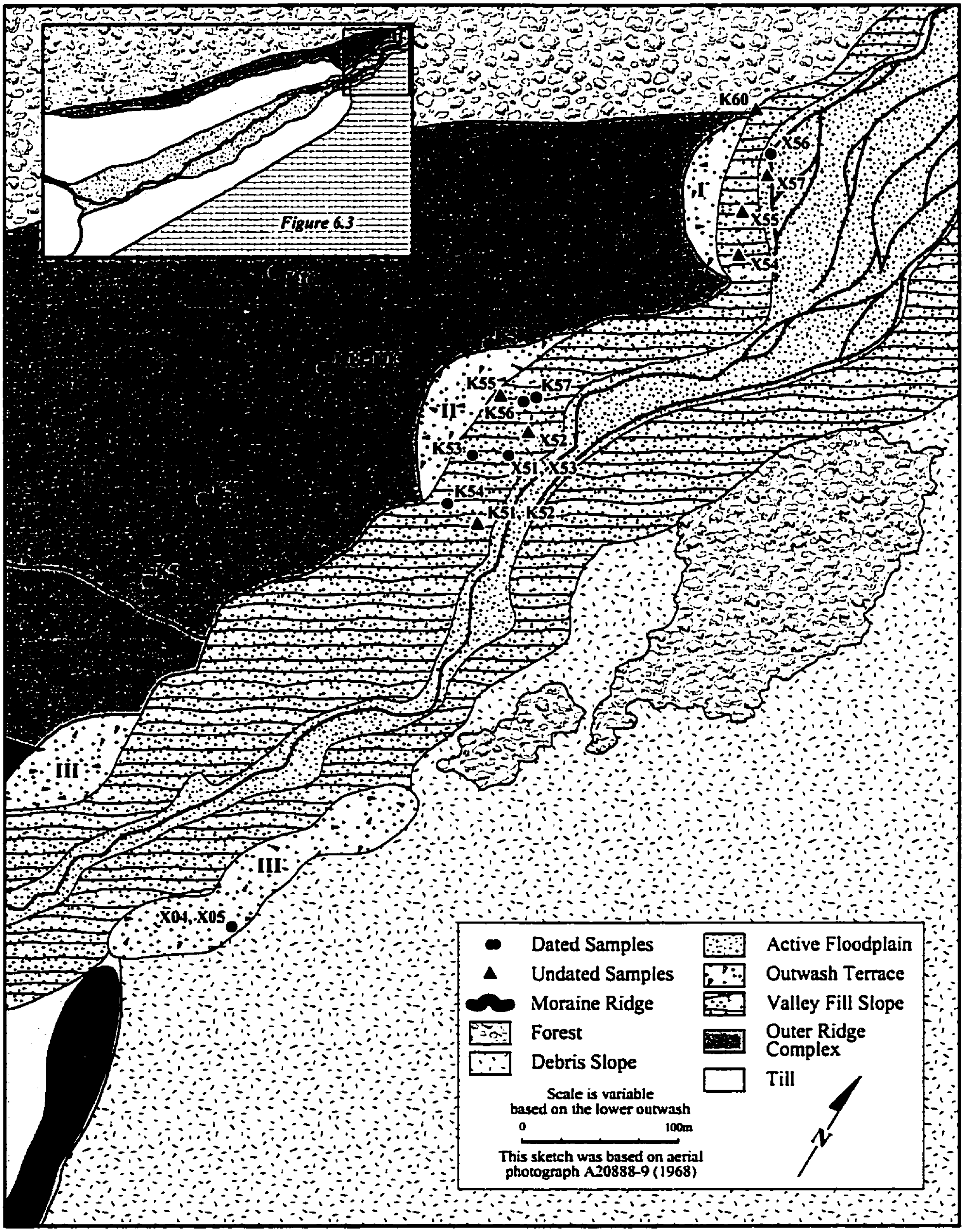


Figure 6.10 Location of valley fill snag samples at Kitchener Glacier. Year identification has been excluded from sample numbers (e.g. K50=K9650).

for these trees provide limiting dates for the enclosing sediments. Fifteen trees were sampled within these sediments but only 7 were successfully crossdated and only 1 of these was located below terrace I (X9656). Although X9656 was found in the present stream bed and not on the slope, it was rooted below the surface and the top of the tree had been sheared off. Therefore, it was likely killed by a filling event which buried the tree and the top subsequently fell. Although all of these trees likely have some outer rings missing, 5 (of 7) show similar death dates, between 1778 and 1794 (Table 6.2) while their inner ring dates suggest a stand of mixed age. Since the highest tree sectioned was 3 m below the terrace fill surface, it suggests that a minimum of 3-4 m of deposition postdated ca. 1790. This is a minimum estimate because the rooting depth of the buried trees is unknown. However, the narrow spread of death dates suggests that most of the trees were growing on the same surface.

Both trees with older death dates (X9651 and X9653) were found within one metre of each other in the bottom third of the slope, below the other samples in terrace II. The discrepancy between the 1317 and 1704 death dates of the lower samples and the late 18th century kill dates for the majority of the samples can be accounted for in several ways. The first is that these two trees were dead before the valley was filled. After the trees died, they remained standing and were incorporated in the sediments. Another explanation suggests that these trees were killed during earlier fill events prior to the 18th century. The final possibility is errors in crossdating. Although both sections produced acceptable crossdates with the regional master, the age and length of X9651 place it completely beyond the limit of the site master and during a period lacking marker rings. Although the dates obtained for X9651 are tentative, the length of sample X9653 ensures a stronger crossdate. This indicates at least one tree in the assemblage was not killed during the fill event of the late 18th - early 19th century, resulting in speculation concerning earlier events.

The main filling event that killed most of the trees found in the valley fill occurred during the late 18th - early 19th century. This is consistent with the surface age of ca. 1820 for terrace III. This suggests that the event that killed the trees and filled the valley also created terrace III. When the date of this event is compared to the ages of the unforested moraines, it appears that the filling event was concurrent with the formation of moraine d between 1713 and 1812. Therefore, the most recent death date from the trees in the valley fill provides a maximum limiting date, resulting in a more precise estimate of the age of the event that created moraine d, 1794-1812. As the glacier advanced to this position over previous till deposits, it deposited considerable debris downvalley, aggrading the channel and forming a "fan-like" terrace deposit beyond the moraine limits.

6.3 Documented Ice Front Positions

The downvalley limit of the kettled area is 2.5 km downvalley of the present snout of Kitchener Glacier. If this limit marks a glacial event, it was approximately 450 m downvalley of moraine c, the only terminal moraine with a closely limiting age of 1713 (ridge c, Table 6.4, Figure 6.11). The next dated position is from the 1948 aerial photograph. Although there is morphological evidence for two readvances between 1713 and 1948, neither could be closely dated as a result of the lack of dateable material. In the 1948 photograph, the glacier had experienced a net retreat of almost 2050 m from its 1713 position marked by the moraine crest. By 1955, the glacier had retreated to within approximately 300 m downvalley of the present ice front. The period between 1955 and 1968 was characterised by a readvance totalling 250 m, with an increase in the rate of advance between 1966 and 1968. During this advance phase, the snout of the glacier changed from a tapered tongue into a broad terminus. The glacier then retreated the 525 m to its present position.

Table 6.4

Recession of Kitchener Glacier

Period	Distance of Retreat (m)	Rate of Retreat (m/yr)
1713 - 1948	2038	-
1948 - 1955	308	44
1955 - 1966	-67	-6
1966 - 1968	-188	-94
1968 - 1974	308	51
1974 - 1992	215	12

0 500m
Scale is variable based on lower outwash

A - 1992 C - 1968 E - 1948
B - 1974 D - 1955

N

Figure 6.11 Recent Ice Front Positions at Kitchener Glacier (NAPL A31609-50).

6.4 Summary

Evidence from the Kitchener Glacier forefield indicates four periods of moraine construction (Table 6.4). Minimum surface ages for ridges a and b and the kettled region suggest that they were formed in the early 16th century. However, snags within the valley fill suggest that the feature predates the beginning of the 12th century or at least the early 14th century. Therefore, if they are glacial features, the maximum Holocene extent of Kitchener Glacier was achieved prior to the LIA. Three events in the LIA were documented. The LIA maximum moraine (ridge c) predates 1713. The next moraine (ridge d) represents a major readvance that can be dated to between 1713 and 1812. However, death dates from trees within alluvium indicate a filling event in the late 18th century and the minimum surface date for one of the terraces is early 19th century. Therefore, moraine d is most probably early 19th century (1794-1812). The final event documented by morphological evidence probably postdates 1812 and based on a correlation with other areas, probably dates from the mid 19th century.

Table 6.5

Major Periods of Moraine Development at Kitchener Glacier

Event Date	Supporting Evidence
pre-LIA	Oldest trees on ridges a and b and in the kettled area, inner dates of trees buried in valley fill
ca. 1713	Oldest tree on ridge c
1794-1812	Outer dates of trees buried in valley fill, oldest tree on ridge d and terrace
post 1812	Oldest tree on ridge e

CHAPTER VII

COLUMBIA AND MANITOBA GLACIERS

7.1 Introduction

The Columbia Glacier occupies the Upper Athabasca Valley and is the only major glacier on the western side of the Icefield. It is also the only outlet glacier that is presently calving into a proglacial lake. Several glaciers formerly coalesced with the Columbia Glacier at its maximum position during the Little Ice Age. The largest of these flows northwards from Mount King Edward and is referred to as the Manitoba Glacier (after Habel 1902). The termini of the Columbia and Manitoba glaciers merged at their Little Ice Age maxima forming a complex series of moraines on the south side of the main valley floor. Investigations at this site attempted to reconstruct the history of both glaciers.

7.1.1 Columbia Glacier

Columbia Glacier is the only major outlet on the west side of the Icefield and Robinson (1996) estimated that approximately 75% of its total area of 38 km^2 lies above the major icefall that divides the accumulation area from the ablation area. The glacier flows approximately 4 km from the icefall and terminates in a large proglacial lake at 1503 m. (Figure 7.1). Although the glacier tongue has a very gentle downvalley gradient (approximately 5°), the glacier descends 700 m in the first 2.5 km from the main Icefield in a spectacular icefall characterised by classic ogives and heavy crevassing. The central portion of the glacier is relatively clean ice, whereas the northern and southern margins are debris covered, resulting in an uneven front through most of its historical record (Figure 7.2). The terminal lake is first seen in historical and aerial photographs taken in 1948 (Field 1949) and expanded to a size of 1.6 km^2 by 1996. The Athabasca River is the only outlet of the lake and is incised slightly into the outwash fan that fills the valley floor downstream of the lake. Further west, the Athabasca River braids freely across the entire valley floor.

Near the ice front, the moraines on either side of the valley are built up to height of 100-125 m against the flanking cliffs and are dissected by steep gullies. On the northern

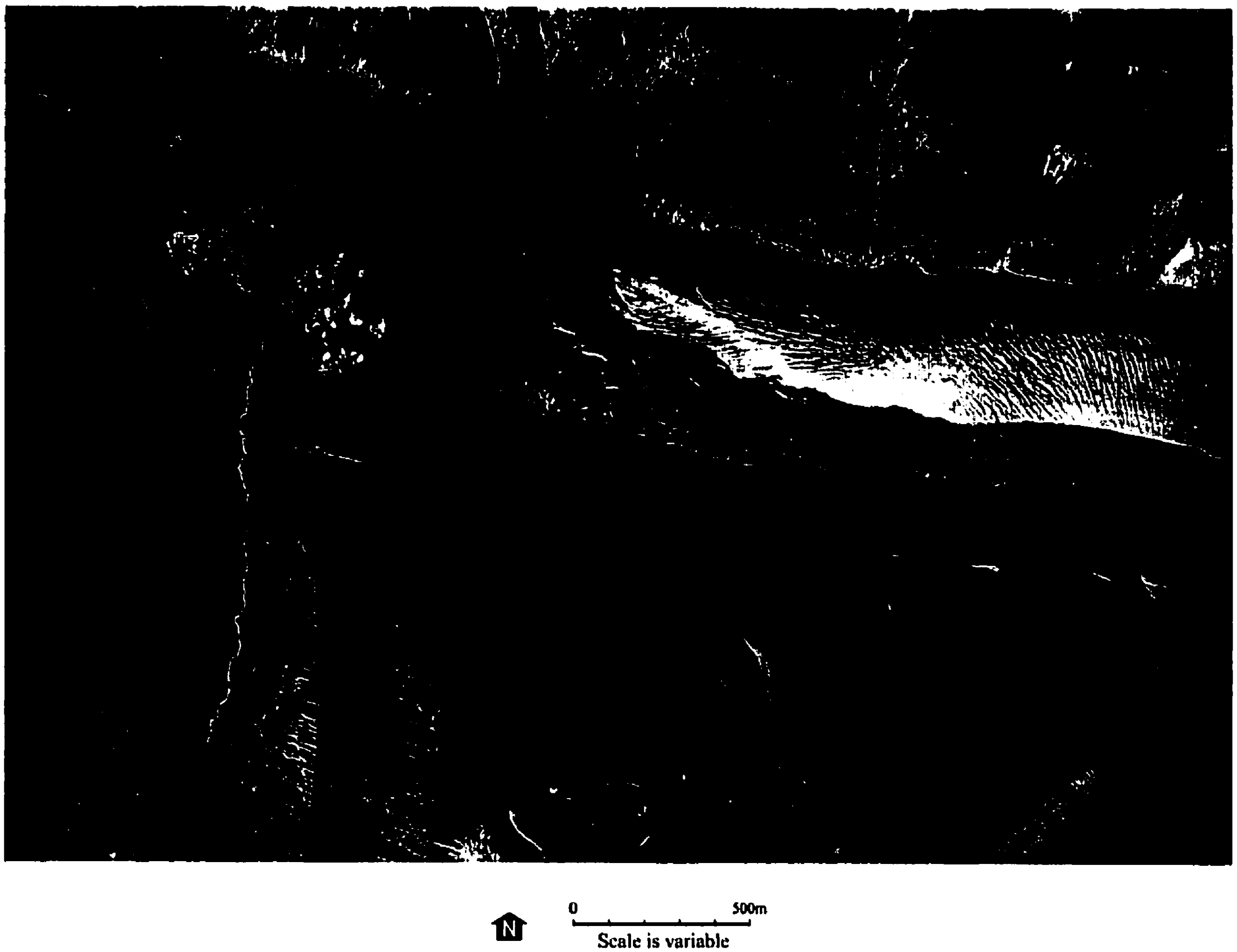


Figure 6.2 Aerial Photo of the Columbia and Manitoba Glaciers site (Province of British Columbia, Photo Number BC7877-57, Sept. 1976).

Figure 7.2 The Columbia Glacier. The tributary valley on the right is the Manitoba Valley (Photo BHL, June 22, 1996).

valley wall, lateral moraine remnants extend 750 m downvalley beyond the present ice front before grading into till plastered against the valley wall and an indeterminant trimline at the limit of the forefield. On the southern valley wall, the lateral moraine is evident until the junction with the Manitoba Valley. As a result of their steepness and instability, both lateral moraine crests were extremely difficult to access and were not visited in the field.

The terminal moraines on the valley floor are low, fluted ridges less than 15 m high. At their downvalley limit, 2.5 km from the 1992 snout, they can be divided into three areas by stream channels cut through the moraine complex (Figure 7.3). The moraines are bisected by the present Athabasca River in the centre of the valley and the south side is further dissected by an abandoned channel that was probably formed by the Athabasca River. The moraines north of this abandoned channel are conspicuous, arcuate ridges that are sparsely vegetated (Figure 7.4, areas B & C in Figure 7.3). On the south side of the valley (Area A in Figure 7.3), the interaction between the Athabasca and Manitoba Glaciers has resulted in a complex pattern of cross-cutting moraines. Downvalley of the moraines, the whole width of the valley is filled with outwash that heads in abandoned fans or channels cut through the moraines. The presence of large boulders and scattered till islands on this outwash terrace, plus the downvalley limits of the lateral trimlines suggest that the moraines marking the Little Ice Age limit of Columbia Glacier were destroyed by fluvial activity and lay slightly downvalley of the present outer moraine complex.

7.1.2 Manitoba Glacier

The Manitoba Glacier is the informal name applied to the heavily debris-mantled glacier fed by ice from the main Icefield that avalanches northwards down the ridge between Mt. King Edward and Mt. Columbia (Figures 7.1 and 7.5). Although the total area of the glacier is approximately 8 km^2 (Robinson 1996), only 2.8 km^2 is below the 2500 m.a.s.l. 1992 snowline (Robinson 1996). The glacier is 2 km long and dissected by supraglacial drainage channels. It has a relatively gentle downvalley gradient and terminates at approximately 1700 m.a.s.l. (Robinson 1996), 200 m above the main Athabasca Valley.

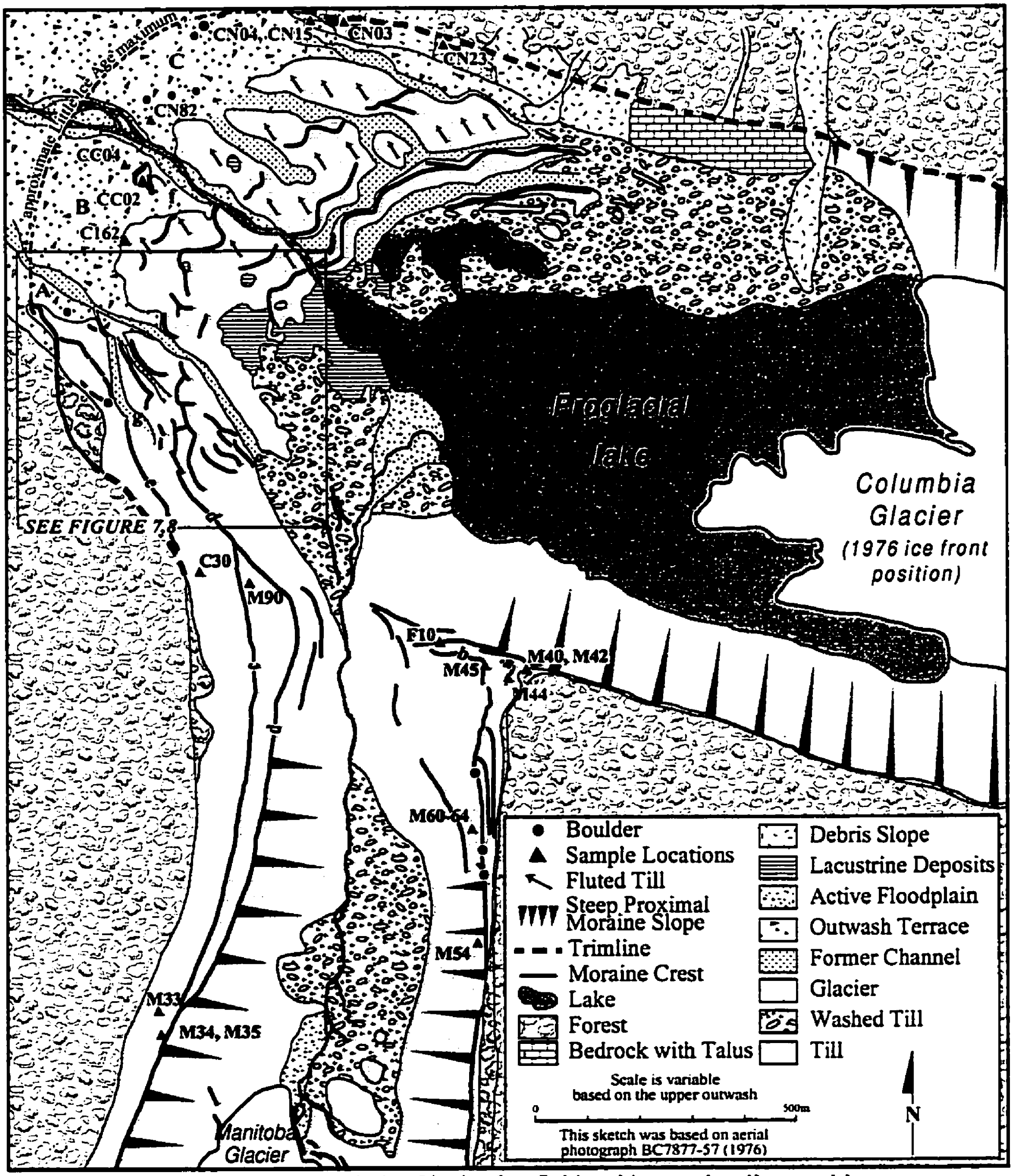


Figure 7.3 Columbia and Manitoba Glacier forefields with samples discussed in text. Year identification has been excluded from sample numbers (e.g. C30=C9630).

Figure 7.4 Oblique view of the Columbia Glacier forefield (Photo BHL, June 22, 1996).

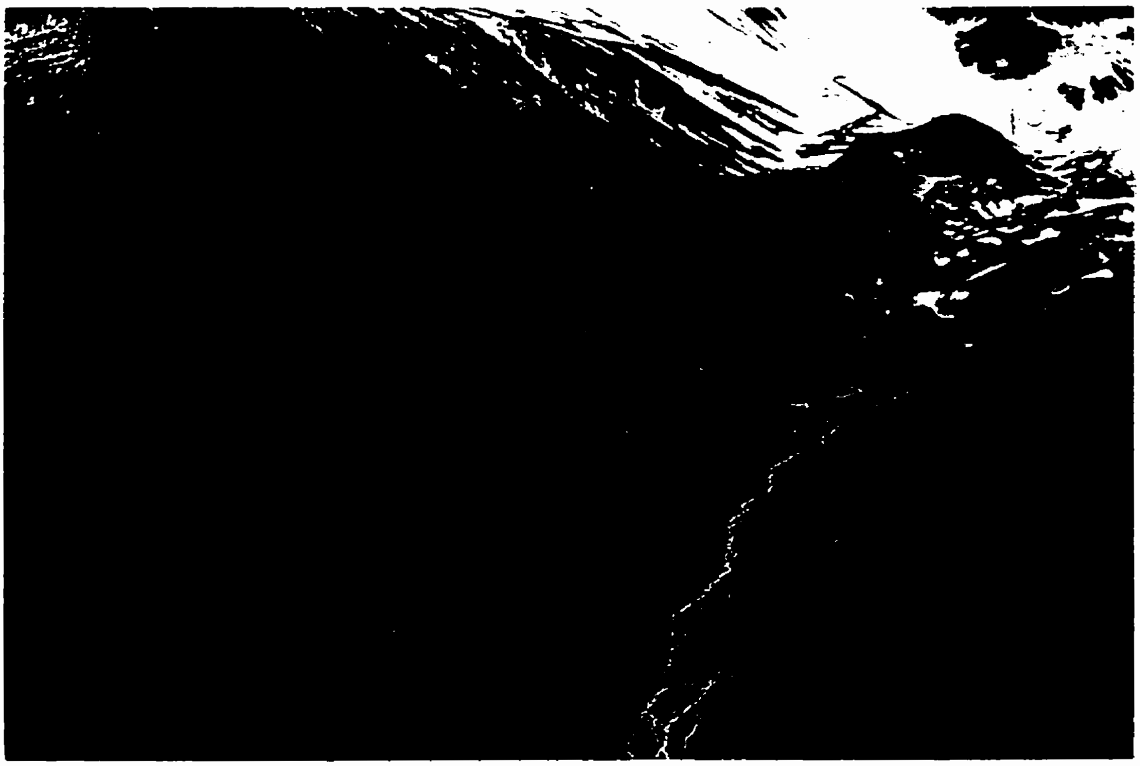


Figure 7.5 The Manitoba Valley (Photo BHL, June 22, 1996).

During the LIA, the Manitoba Glacier extended 1 km downvalley of its 1992 position and coalesced with the Columbia Glacier. Flanking the glacier, there are well developed lateral moraines that rise in steep, gullied slopes 75-100 m above the main valley floor but have gentler, undissected distal slopes that rise 15-20 m above the adjacent valley sides. Both lateral moraines have relatively minor downvalley gradients within the Manitoba Valley, but descend abruptly near the junction with the Athabasca Valley. In addition, at their downvalley limits, moraines on both sides of the Manitoba Valley are deflected westward parallel to the formerly coalescent Columbia Glacier. On the eastern side of the Manitoba Valley, the lateral moraine crest rises above the adjacent forested valley side and there is no forest trimline. At the northern (downvalley) limit of this moraine, as it drops down to the Columbia, it subdivides into several smaller ridges that are truncated by a lateral moraine from the Columbia Glacier. On the west side of the valley, the main lateral moraine divides into two ridges approximately 425 m downvalley of the current ice front. These ridges gradually diminish in height downvalley and become barely perceptible ridges within the trimline which becomes the dominant feature along the western valley wall. At the junction with the Athabasca Valley, the trimline and moraines turn westward and continue uninterrupted into the Athabasca Valley (Figure 7. 6).

7.2 Previous Investigations

Glacier fronts of the Columbia Glacier during the 20th century have been documented by a variety of researchers (e.g. Schäffer 1908, Thorington 1925, Field 1949) and will be discussed in detail in section 7.4. Glacier variations prior to these historical accounts were determined by Heusser (1956) who visited the site in 1953. Employing dendrochronology, he found that the ice began retreating from its maximum position in the early 18th century. Using an ecesis interval of 14 yrs., trees found near the terminal position along the northern and southern trimlines suggest that recession began in approximately 1724 and 1763 respectively. The later date from the southern trimline was attributed to the presence of the Manitoba Glacier coalescing with the Columbia. Dates from several groups of trees within the limit of the valley floor indicate that the ice had receded from that position in about 1796. Heusser (1956) suggested that the recession

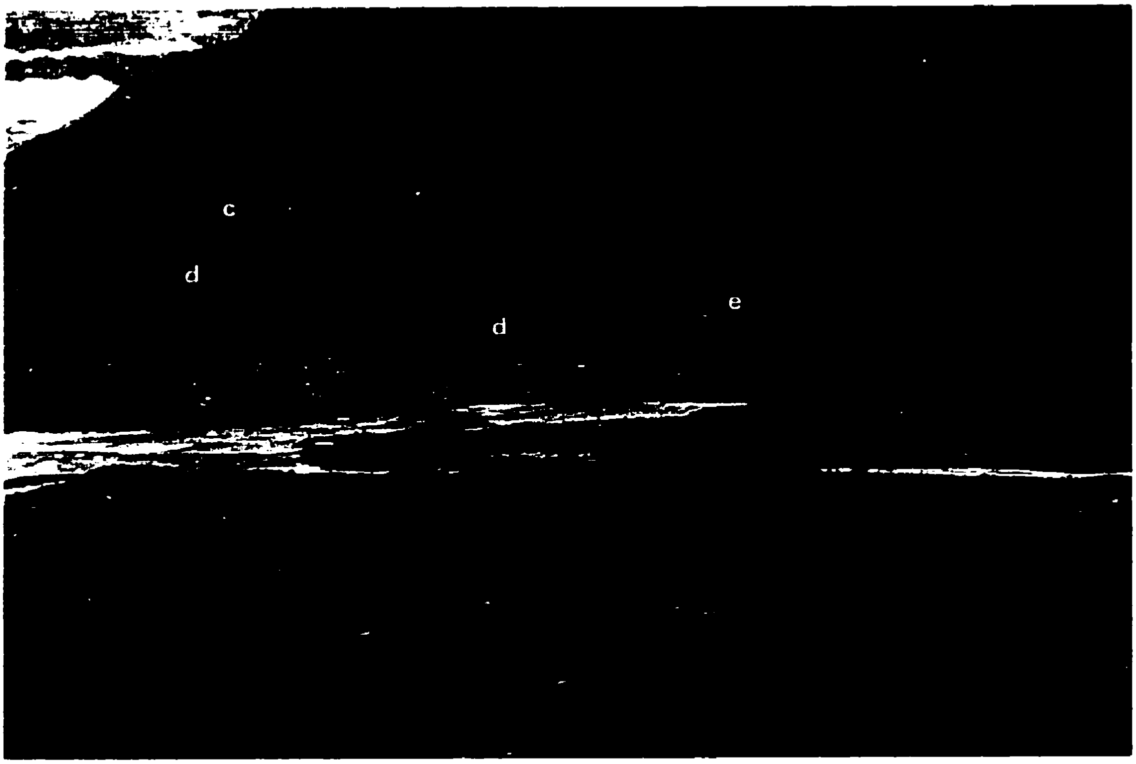


Figure 7.6 Western junction of the Athabasca and Manitoba valleys. Note the trimline and moraines continuing out from the Manitoba Valley and the truncation of ridge c by ridge d (Photo BHL, June 28, 1996).

occurred first along the northern trimline, then from the southern trimline and finally from the centre of the valley. This implies that the glacier began thinning prior to its retreat.

Heusser (1956) reported that Field sampled a tilted tree in 1948 that indicated a glacier advance in 1842 that was of similar magnitude to the maximum. Unfortunately, the precise location of this sample is unknown (Heusser, personal communication 1996) and, despite extensive searching, was not relocated during field investigation in 1996. Along the northern trimline, Heusser dated retreat from a position near the maximum to 1854 whereas a small moraine located close to the "presumed" maximum mid-valley dated to 1864. In the period between 1864 and the first documented ice front position in 1901, the Columbia Glacier formed a single moraine in 1871 (Heusser 1956).

7.3 Results of the Present Study

As the 20th century history of the Columbia Glacier is well documented, this study focused on reconstructing earlier events. In this discussion, the site will be divided into three areas, namely: 1) the Manitoba Valley, 2) the confluence zone of the Columbia and Manitoba Glaciers on the south side of the Athabasca Valley (Area A in Figure 7.3) and 3) the central and northern portions of the Athabasca Valley (Areas B & C in Figure 7.3). Following a discussion of each individual area, the evidence will be summarised in the overall site chronology presented in section 7.3.4. Complete results from all samples collected at this site are presented in Appendix IV.

In association with the present study, G. Osborn collected several trees and *in-situ* stumps that were incorporated into the main lateral moraines of Columbia Glacier. As the length of record of these stumps was too short to permit crossdating, all the samples were retained by G. Osborn for radiocarbon dating. However, results from this analysis are not yet available and therefore will not be discussed.

7.3.1 Manitoba Valley

The Manitoba Glacier extended out of its valley and coalesced with the Columbia Glacier during the LIA. There are no dateable, terminal features inside the Manitoba Valley. The glacial events in the Manitoba Valley were determined from evidence found along the 75-100 m high lateral moraines that extend the entire length of the valley. The evidence found on each side of the valley will be discussed separately.

7.3.1.1 Manitoba East

The eastern lateral moraine is a single, sharp-crested ridge with a gentle downvalley gradient for ca. 500 m immediately downvalley of the current ice front. At the point marked by three large boulders on the crest of the lateral moraine (Figure 7.3), the main ridge grades from a 15- 20 m high ridge down to a subdued crest less than 2 m high and descends steeply to the Athabasca Valley. About 30 m below the main crest on the proximal slope, there is a single ridge that curves into the valley. Several minor ridges found between this inner ridge and the main lateral are the result of debris flow activity that has affected the area (Figure 7.5). The other two ridges, less than 2-3 m high, are found between the trimline and the main lateral moraine. Near the limit of the Manitoba Valley, several small ridges parallel to the Manitoba Valley are deflected to the west. The remaining moraines in this area run parallel to the Athabasca Valley.

Near the junction with the Athabasca Valley, all of the moraine ridges are sparsely vegetated. However, at its upvalley limit, the distal slope of the main lateral is 15-20 m high and most of it is covered by dense forest. As this moraine was built up between the ice front and the adjacent forest, the glacier was not in direct contact with trees, eliminating the possibility of tilted or killed trees. Several trees sampled in this area were young with wide rings indicating a sheltered environment. The young trees and advanced state of decomposition in this area suggested it was unlikely that dateable older material would be found and further sampling was abandoned.

Evidence for the dating of glacial events was found on the proximal slope of the main lateral moraine below the three large boulders that mark the break in slope (Figure 7.3). Between the middle and furthest downvalley of these boulders, detrital wood was found on and partially buried in the moraine surface. These logs ranged from 10 to 50 cm in diameter and from 0.5 to 3 m in length, and were obvious remnants of larger trees. As the present vegetation cover is sparse and predominantly stunted growth forms up to 10 cm maximum trunk diameter and 3 m maximum height, these larger logs must have been transported to these positions. As they were found on the proximal slope of the main moraine, the only viable method of transportation and burial is glacial. Unfortunately, many of the logs were rotten, shattered or almost completely buried under the large boulders and impossible to excavate with the available equipment. Five logs were cross sectioned and dated (Table 7.1). There was no geographic pattern noted in the distribution of these logs but the outer ring dates fall into two groups, 1690-1698 and 1796-1808. These suggest derivation and deposition by two separate advances: one ca. 1700 and the other in the early 1800's.

Table 7.1

Outer Ring Dates for Logs Found on the Surface of the Eastern Manitoba Main Lateral Moraine

Section ID	Calendar Dates
M9660	1420-1690
M9661	1658-1796
M9662	1661-1798
M9663	1426-1698
M9664	1593-1808

Note: Sample ID's include year identification.

Subsurface material indicating an earlier advance was also found protruding from the inner face of the east lateral moraine. The trunk was lying at 90° to the moraine crest but parallel to the distal slope. Although two large logs were retrieved from the main lateral moraine, about 1 m below the crest (Figure 7.3), only one has been tentatively cross-dated and has an outer ring date of 1474 (M9654, Figure 7.7). This indicates a glacial advance in the late 15th century. The orientation of this log suggests that it may have been deposited on the distal face of the moraine and subsequently buried by accreting debris from the glacier topping the lateral moraine. Therefore, the moraine was built up to its present height by the 1400's and stabilised relatively soon after that (following the deposition of 1 m of sediment); accounting for the well developed forest cover on the distal slope. The dates obtained from logs found within and on the lateral moraine suggest that although this moraine was re-occupied in the 18th century, the glacier was reworking trees killed and/or buried by earlier glacial events and probably did not overtop the moraine crest over its entire length.

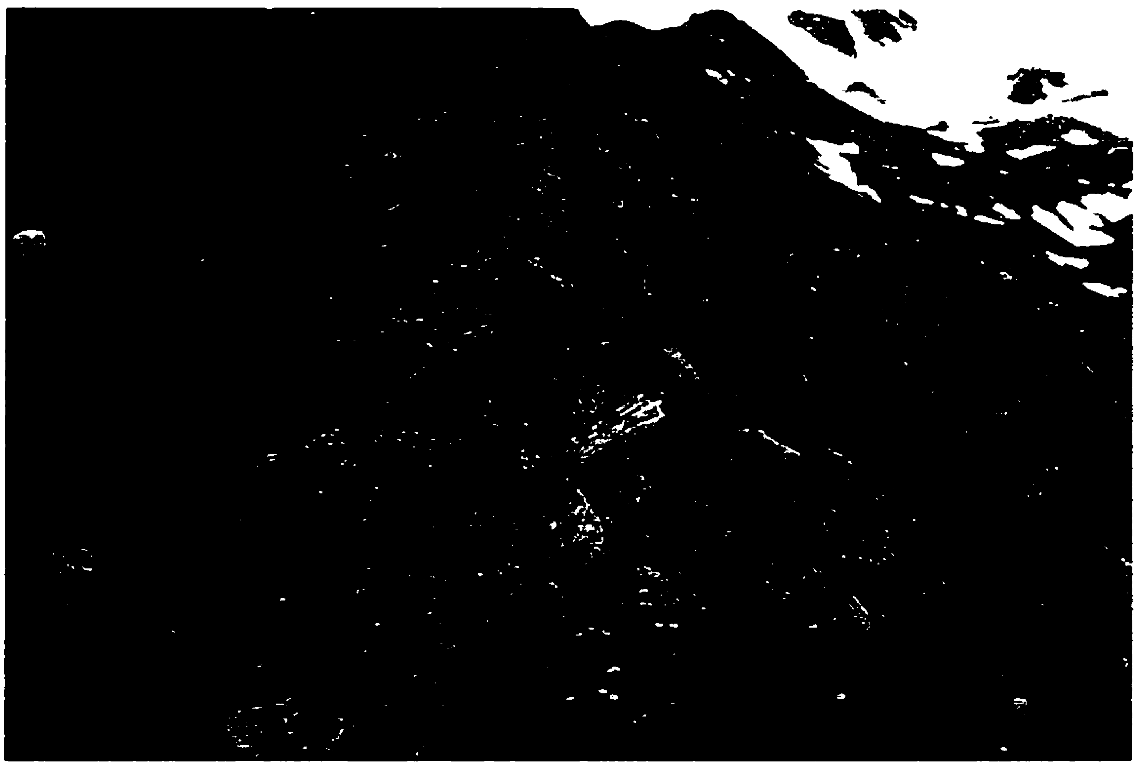


Figure 7.7 Log imbedded in eastern lateral moraine of the Manitoba Glacier (Photo BHL, June 23, 1996).

Table 7.2

Limiting Ages for the Moraines at the Eastern Junction
Between the Manitoba and Athabasca Valleys

Location	Cores			Cross Sections		
	Core ID	Inner Ring Date	Surface Date	Section ID	Section Range	Surface Date
Below trimline	**M9640**	**1777**	**1755**	M9642	1777-1986	1757
Ridge a	**M9644**	**1840**	**1818**			
Ridge b	**M9645**	**1862**	**1837**			
Columbia moraine	**F9610**	**1901**	**1871**			

Note: Sample ID's include year identification. Surface date includes an ecesis of 20 yrs. and pith correction. Bold indicates the limiting date.

The area east of the junction between the Manitoba and Athabasca Valleys was sampled in an effort to establish a date for the maximum position of the Manitoba Glacier (Table 7.2, Figure 7.3). The oldest living tree sampled had a near pith date of 1777 (M9640) and was located approximately 10 m below the trimline at the valley junction. This date was supported by the age of a standing snag (M9642) less than 10 m across the slope from M9640 which revealed a pith date of 1777. At the downvalley limit of the main lateral moraine, there are two small (<2 m high) ridges (ridges a and b, Figure 7.3) on either side of its crest. At their upvalley limits, these moraines are parallel to the Manitoba Valley for approximately 50 m before turning abruptly westward. The oldest tree sampled on the north-south trending portion of the outermost ridge (ridge a) was found at the base of its distal slope and revealed a near pith date of 1840 (M9644). The oldest tree on the inner ridge (ridge b) was located in a similar position, at the base, on the distal slope of the north-south trending moraine and M9645 revealed a near pith date of 1862. The remaining moraines in this area are parallel to the Athabasca Valley and lack significant vegetation cover. A single tree located on the 2.5 m high distal slope of the main Columbia lateral moraine revealed an off pith date of 1901 (F9610), corresponding to a germination date 1871. Results from this area indicate that the trimline within the Manitoba Valley was formed prior to 1755 while the two outer moraines were deposited after 1818 and 1837 respectively, thereby suggesting that the Manitoba Glacier had retreated from this area by the mid 19th century. However, the date from the Columbia moraine indicates that the Columbia Glacier remained close to this position until the late 19th century.

7.3.1.2 Manitoba West

The west side of the Manitoba Valley above the current ice front closely resembles the lateral moraine east of the glacier. It is a single, sharp crested ridge with a distal slope approximately 10-15 m high and a gentle downvalley gradient. However, unlike its eastern counterpart, the distal slope of the western lateral moraine is only sparsely vegetated. Approximately 425 m downvalley from the current ice front, the lateral moraine divides into two distinct ridges (Figure 7.3). These moraines gradually diminish to barely perceptible ridges downvalley and the trimline becomes the dominant feature. As the moraines descend steeply towards the Athabasca Valley, the distance between them increases and a colour difference becomes apparent (Figure 7.6). The area between the trimline and the upper moraine (ridge c) is a predominantly "buff" coloured material steeply sloping into the Manitoba Valley with a scattered tree cover. Ridge c is clearly identified by a row of trees and the contact between the "buff" and the dark grey material. The lower moraine (ridge d) represents the transition to the predominantly fine grained, light grey material with scattered large boulders that forms the inner slope of the lateral moraine. Near the downvalley limit of the Manitoba Valley, the trimline and these two moraines are deflected westward along the wall of the Athabasca Valley (Figure 7.6).

As a result of the limited dateable and morphological evidence on the west side of the valley, fewer glacial events are identified. Approximately 400 m downvalley of the 1992 ice front, immediately upvalley of the location where the lateral moraine splits into two ridges, several trees were sampled on the moraine and in the trench between the trimline and the moraine (Figure 7.3). The oldest tree sampled in the trench at the base of the moraine revealed an off pith date of 1771 (M9633), indicating that the trimline was formed prior to 1741 (Table 7.3). However, where the Manitoba Glacier entered the Athabasca Valley, the oldest tree approximately 20 m below the trimline yielded a pith date of 1790 (C9630, Table 7.3). The large difference in limiting ages for a single feature is likely a result of the environmental factors affecting both of the sample locations. While the Manitoba Glacier may have retreated from the trimline, the Columbia Glacier likely remained close to its maximum position. Therefore, the junction between the Manitoba and Athabasca valleys would still be adjacent to ice, hindering the colonisation of trees, increasing the ecesis intervals for this area and resulting in an apparently younger surface age.

A tree sampled on the crest of the 10 m high main lateral moraine, approximately 400 m from the 1992 ice front suggests that the moraine was created ca. 1784 (M9635). This date is supported by a second tree located several metres from M9635, which revealed a near pith date of 1822. These dates provide a limiting date of 1797 (Table 7.3) for the single moraine ridge. Downvalley, a single tree with a pith date of 1884 (M9690) was found between ridges c and d. This suggest that the outer moraine at the junction between the two valleys wasn't deposited until ca. 1864.

Table 7.3

Limiting Ages for the Western Side of the Manitoba Valley

Location	Core ID	Inner Ring Date	Surface Date
Below trimline	**M9633**	1771	**1741**
Main lateral	**M9635**	**1806**	**1784**
	M9634	1822	1797
Below trimline	**C9630**	**1790**	**1770**
Between ridges C and D	**M9690**	**1884**	**1864**

Note: Sample ID's include year identification. Surface date includes an ecesis of 20 yrs. and pith correction.

7.3.1.3 Manitoba Valley Summary

Wood found on and in the eastern lateral moraine indicates three periods of glacial advance: at the end of the 15th century, the late 17th/early 18th century and the beginning of the 19th century. The death dates of these trees indicate that the glacier was advancing into the forest at those times and they provide the maximum limiting dates for the formation of the eastern lateral moraine (since the tree was killed before burial). Trees growing in the forefield also provide evidence for an early 1700's advance. Dates obtained from the trimlines indicate that both were formed prior to 1741. Surfaces with slightly younger limiting dates, such as those from the junction of the Manitoba and Athabasca Valleys, were also believed to be part of this event. The conditions these trees were exposed to following initial deglaciation justifies a younger surface age. Unless both glaciers retreated rapidly and synchronously from their maximum positions, the junction would still be in close proximity to at least one of the glaciers. This would adversely affect the local environment and increase the amount of time it would take for the trees to invade the area.

Limited evidence for a late 1700's advance exists in the Manitoba Valley. Although the western main lateral moraine appears to have stabilised during this period, only a small moraine near the junction (likely with a longer ecesis interval) was found on the east side of the valley. Both maximum and minimum dates for a mid 1800's advance exist within the Manitoba Valley. Trees on the eastern lateral moraine indicate that the glacier was advancing in 1808, indicating that the moraine was deposited shortly thereafter. This value is corroborated by the surface ages on many of the inner moraines within the valley.

7.3.2 Zone of Coalescence

The south side of the Athabasca Valley, west of the Manitoba Valley (Area A) is a complex assemblage of moraines created by the coalescence of the Columbia and Manitoba Glaciers (Figure 7.8). Many of the features along the southern valley wall are continuous with landforms along the west side of the Manitoba Valley. Along the valley floor, the coalescence of the two glaciers has resulted in a complicated pattern of perpendicular moraines and fluvioglacial channels.

7.3.2.1 Southern Valley Wall

The trimline along the southern side of the Athabasca Valley west of the Manitoba Valley continues for approximately 250 m from the junction before being obscured by the forest (Figure 7.3). While the trimline is not readily observable in the forest, a small ridge less than 3 m high continues through the forest and terminates at the valley floor. The

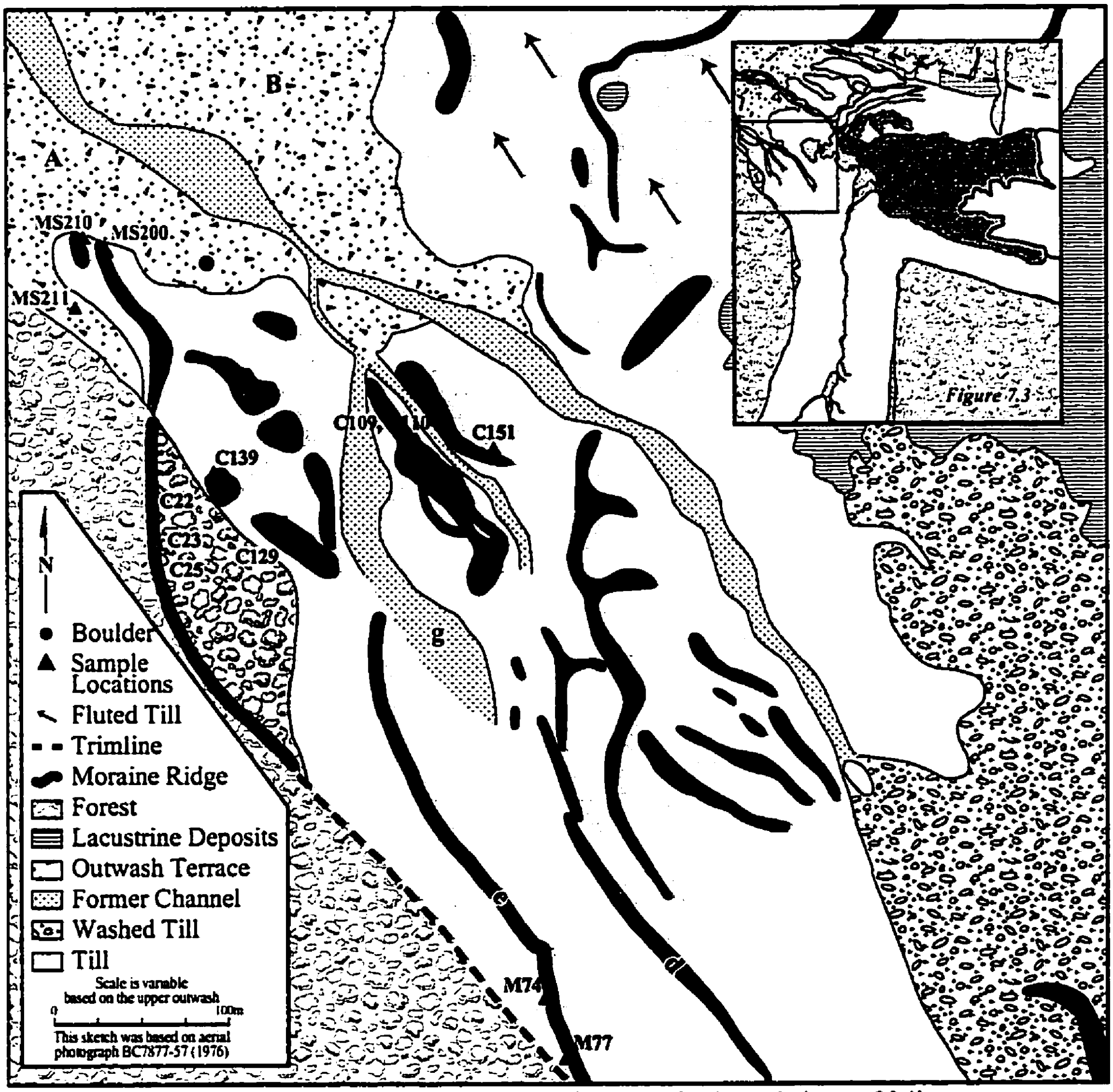


Figure 7.8 Sample locations in the zone of coalescence in the Athabasca Valley. Year identification has been excluded from sample numbers (e.g. M74=M9674).

second moraine along the valley wall (ridge e) is a ridge approximately 1 m high that begins at the junction and extends down to the valley floor. Approximately one third of the way down the slope, this moraine sharply levels out before continuing to the valley floor where it ends at the head of an outwash channel. The final feature along the valley wall is ridge d (Figure 7.3) continuing out from the Manitoba Valley. Where ridges c and d enter the Athabasca Valley, both are deflected to the west. However, ridge d is diverted almost parallel to the valley axis and truncates ridge c (Figure 7.6). Downvalley, this ridge continues obliquely downslope and merges into a 10 m high, unvegetated ridge on the valley floor.

Table 7.4

Limiting Ages for the Southern Valley Wall of the Athabasca Valley

Location	Core ID	Inner Ring Date	Surface Date
Moraine in the forest	**C9625**	**1770**	**1740**
	C9323	1796	1766
	C9622	1789	1769
Between trimline and ridge e	**M9674**	**1818**	**1793**
	M9677	1832	1807

Note: Sample ID's include year identification. Surface date includes an ecesis of 20 yrs. and pith correction. Bold indicates the limiting date.

There were no trees along ridge d and sampling along the southern valley wall was limited to the few groves of trees upslope of ridge e and the outer moraine in the forest. In a grove of trees 20 m below the trimline and several metres above ridge e (Figure 7.8), the oldest tree sampled yielded a near pith date of 1818 (M9674), suggesting a minimum surface age of 1793 (Table 7.4). Sampling along the outer moraine in the forest indicate that this feature was deposited in the early 18th century (Appendix IV). An off pith date of 1770 (C9625) obtained from a large spruce growing on the crest of the moraine about 80 m from its downvalley limit indicates that the moraine was formed ca. 1740 (Table 7.4). These dates suggest that the furthest downvalley extent of ice on the southern valley wall was reached in the mid 18th century and experienced a net retreat of approximately 250 m by the end of the 18th century.

7.3.2.2 Southern Valley Floor

The valley floor in the zone of coalescence is characterised by cross cutting moraines, fluvioglacial channels and fans resulting in a rather complex pattern (Figure 7.8). The moraines just west of the Manitoba Valley are large (up to 10 m high), discrete ridges void of vegetation, although moss and a few prostrate shrubs exist in the "inter-moraine" areas (Figure 7.9). They are composed predominantly of limestone clasts up to 0.5 m in diameter although a downvalley fining is noticeable. At ridge d there is a distinct transition in morphology and vegetation cover. Downvalley, in addition to distinct ridges with subdued crests, there are several small moraine fragments or "islands". Compared to the moraines upvalley, this area appears to be composed of finer grained material with moss, prostrate shrubs and moderate tree cover (Figure 7.10). Therefore, only the downvalley features could be sampled.

The zone of coalescence is bisected by an abandoned channel (channel g in Figure 7.8). Three distinct moraines downvalley of this feature (moraines I-III) and one moraine upvalley (moraine IV) were sampled. The outermost feature is a moraine fragment (moraine I) less than 3 m high. This hummocky mound is completely detached from all other ridges and is bordered by a small terrace (maximum width of 1 m, too narrow to be depicted in Figure 7.8) approximately 0.75 m above the main outwash surface (Figure 7.11). The second target was the first continuous ridge (moraine II) several metres upvalley of the outermost moraine fragment. At its downvalley limit, it is a 3 m high ridge with a very subdued crest (Figure 7.12) which increases to an almost 7 m high moraine with steep slopes and sharp crest at the forest edge (Figure 7.13). Although it appears that this ridge is continuous with the outermost moraine along the valley wall, there is a clear break between the ridges as well as a distinct difference in morphology. The ridge along the valley wall is a small, subdued ridge where is reaches the valley floor, while moraine II is a sharp crested feature roughly 7 m high near the forest edge.

The two outer features (moraines I and II) reveal similar dates of 1788 and 1789. While there were no trees growing on moraine I, several trees on the small terrace were cored, the oldest yielding a near pith date of 1814 (MS210). As the moraine was truncated by the terrace, it suggests that the terrace postdates moraine and provides a

Figure 7.9 "Unvegetated" moraines in the upvalley portion of the zone of coalescence (Photo BJR, June 25, 1996).

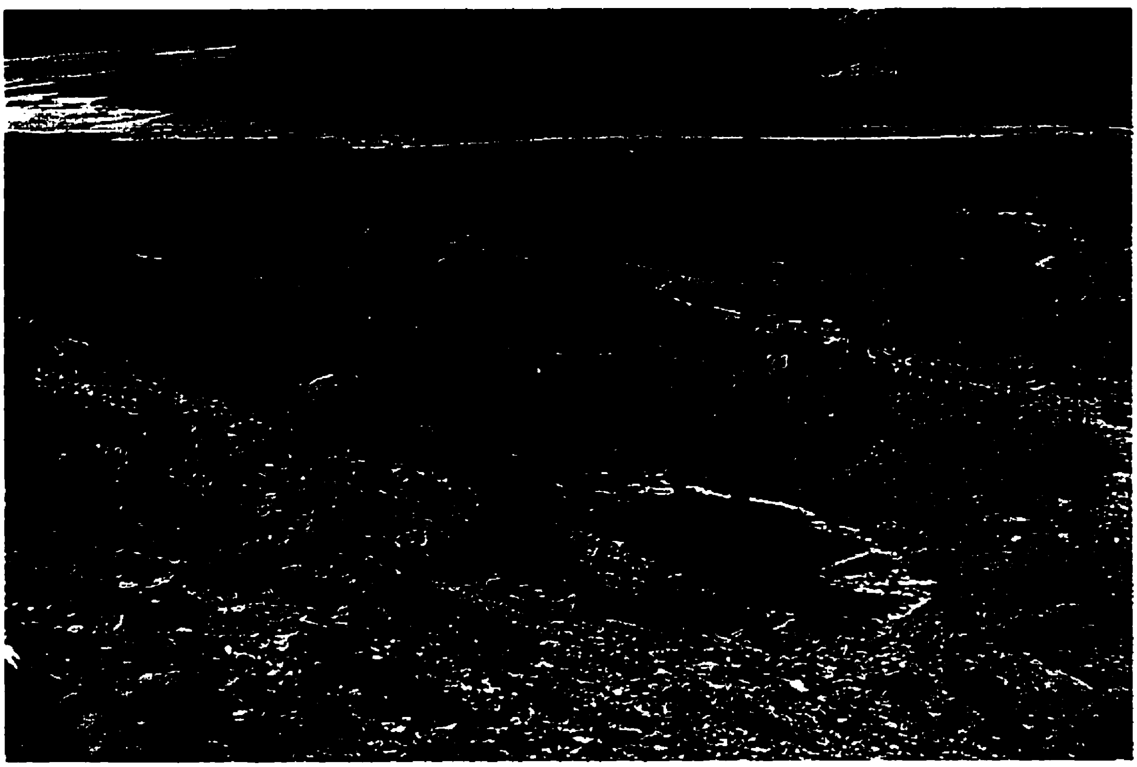


Figure 7.10 Downvalley moraines in the zone of coalescence (Photo BHL, June 27, 1996).

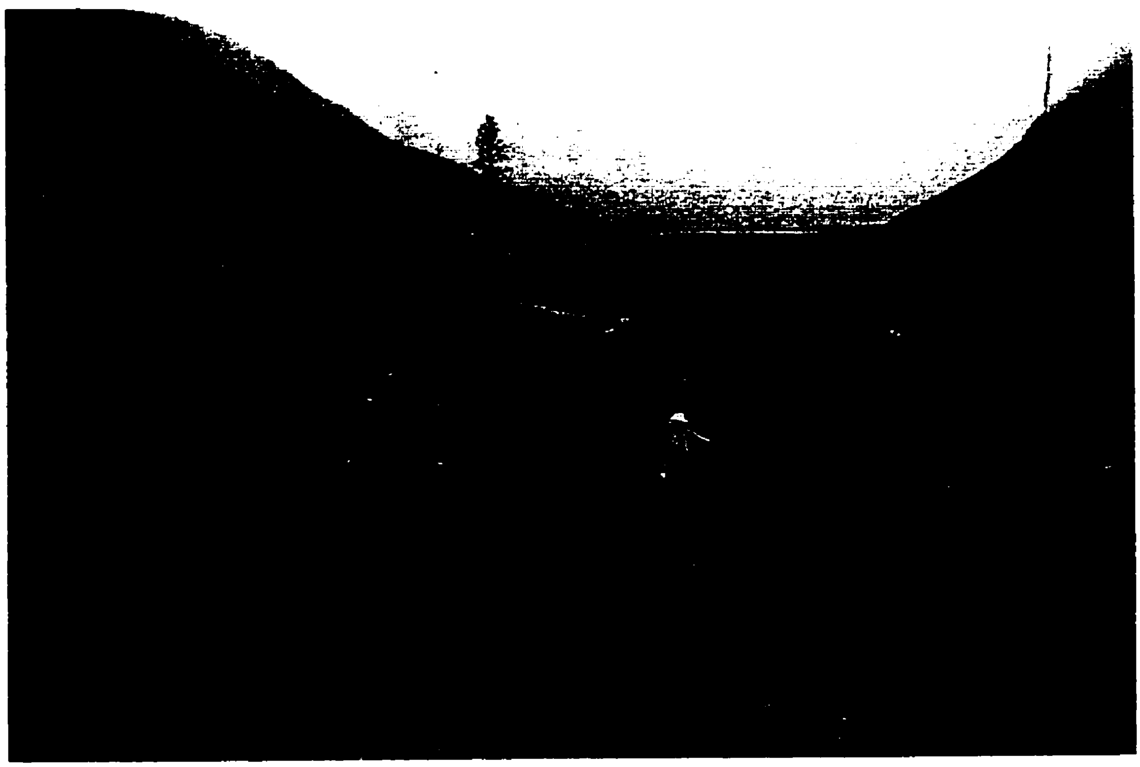


Figure 7.11 Moraine I – located in the centre of the photo. Note the small terrace near the base of the moraine (Photo BJR, June 24, 1996).

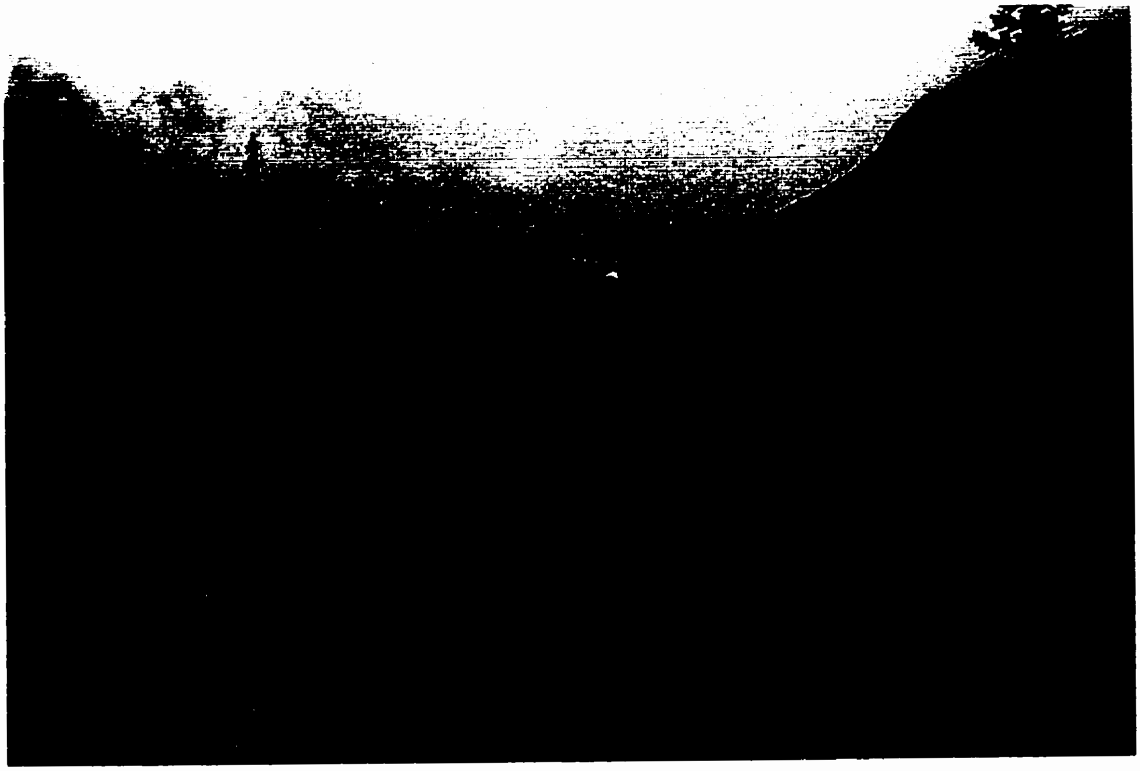


Figure 7.12 Western limit of moraine II (Photo BJR, June 24, 1996).

Figure 7.13 Eastern limit of moraine II (Photo BJR, June 24, 1996).

minimum date of 1789 for the formation of the ridge (Table 7.5). Three trees sampled near the downvalley limit of moraine II on its distal edge revealed similar germination dates of 1788, 1790 and 1792 (MS200, MS203, MS201). These trees were located within 5 m of one another, two halfway up the slope (MS201 and MS203) while the other was growing at the base of a slope which had been undercut by the outwash. Similar ages were also obtained for features surrounding these ridges. A pith dated tree (C96139) on a small moraine fragment upvalley of moraine II yielded a germination date of 1788 (Table 7.5). The similarity in age of these features indicates that they were likely created during a single event in the late 18th century.

The oldest tree sampled on the outwash between moraine I and the valley wall revealed that the surface had stabilised by 1793 (MS211, Table 7.5). However, the stratigraphic relationship between these two features indicates that the moraine was deposited after the outwash surface was formed. These dates suggest that the outwash was likely formed as the glacier advanced to the position marked by moraine I but was not colonised until the ice had retreated from that position.

The third sampling location is a composite feature made up of two nearly perpendicular ridges (moraine III), easily identified in aerial photos as a result of the large boulder at its upvalley limit (Figures 7.1 and 7.8). It is composed of a short ridge 6-7 m high lying parallel to the valley side and a small ridge about 1-2 m high extending northwards out of the moraine slightly upvalley of the boulder. This "boulder moraine" is only about 30 m long and 10 -15 m wide in some places with an eroded crest. A small spruce tree mid way down moraine III, just 1 m off the crest yielded a pith date of 1838 (C96129), indicating that this surface had stabilised by 1818 (Table 7.5). The age of this feature may indicate an advance during the early 19th century. However, the lack of corroborating evidence and similarity in age with the surrounding features suggests that moraine III was formed concurrently with or shortly after the other features in this area, during the late 18th century, and has an attenuated ecesis interval.

The only feature targeted for sampling upvalley of channel g was moraine IV; informally referred to as "moraine 100" and located at the upvalley limit of the vegetated moraines. It is a 120 m long complex of ridges, trending generally downvalley bounded by a large meltwater channel to the north and a smaller one to the south. The most significant characteristic of this feature is the 30 m section of moraine which is perpendicular to the main ridge at its eastern end. Both portions of the moraine are approximately 5 m high, although at the intersection, the crest of the southerly trending section decreases slightly (Figure 7.14). Unfortunately, there were no trees growing on this section of the moraine to determine if both sections were deposited synchronously. On the north side of the main moraine, a slightly lower ridge (3-4 m high) runs parallel to the main crest before cutting across the moraine. In addition, there is a 1 m high, arcuate extension on the south side where the inner ridge cuts across the main ridge.

Table 7.5

Limiting Ages for the Zone of Coalescence

Location	Core ID	Inner Ring Date	Surface Date
Outer moraine fragment (moraine I)	**MS210**	**1814**	**1789**
Continuous moraine (moraine II)	**MS200**	**1811**	**1788**
	MS203	1815	1790
	MS201	1817	1792
Moraine fragment upvalley of moraine II	**C96139**	**1808**	**1788**
Outwash south of moraine I	**MS211**	**1813**	**1793**
Boulder moraine (moraine III)	**C96129**	**1838**	**1818**
	C96133	1847	1827
Moraine 100 (moraine IV)	**C96108**	**1902**	**1877**
	C96109	1898	1878
	C96110	1906	1881
Ridge north of moraine IV	**C96151**	**1910**	**1888**

Note: Sample ID's include year identification. Surface date includes an ecesis of 20 yrs. and pith correction. Bold indicates the limiting date.

Sampling along moraine IV indicated that most of the trees germinated in the latter half of the 19th century (Appendix IV), with maximum ages of 1877 and 1878 (C96108 and C96109, Table 7.5) obtained from two trees in the centre of the moraine, one at the crest and the other at the base of the slope. The oldest tree located on a small ridge located across the meltwater channel north of moraine IV revealed a similar surface age of 1888 (C96151, Table 7.5). Schäffer's (1908) picture of the snout of the Columbia shows

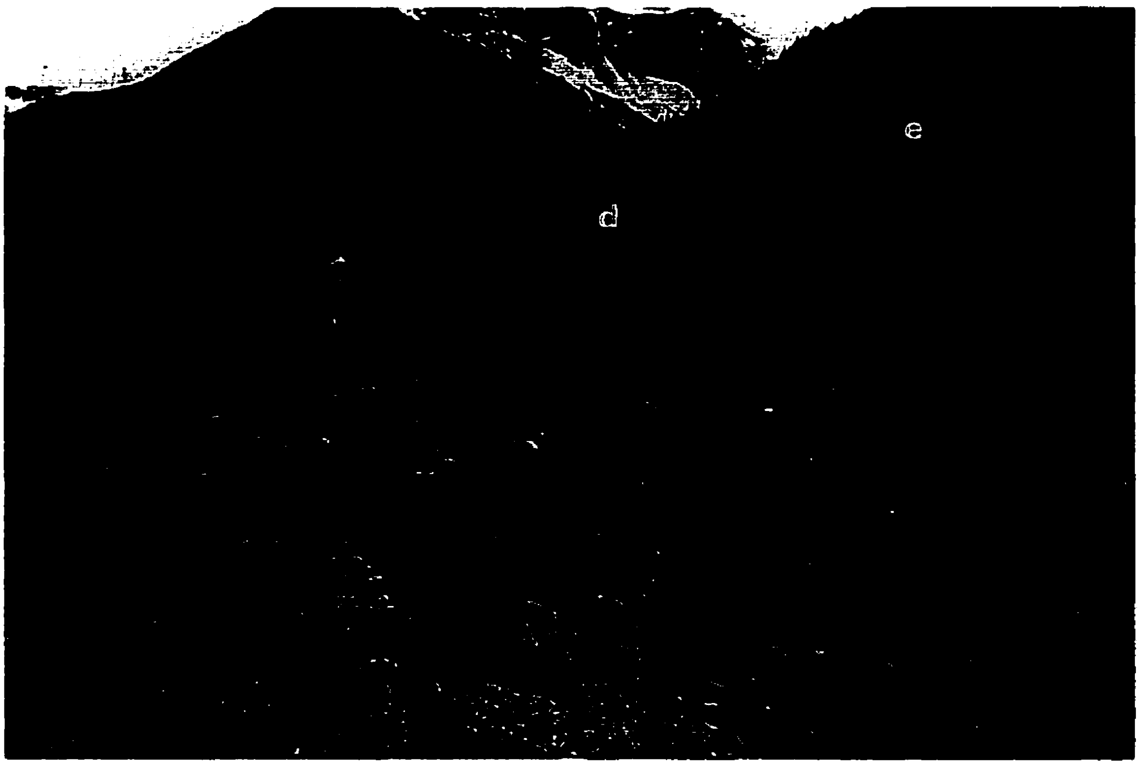


Figure 7.14 Western end of moraine IV. The ridge on the right side of the photo is the perpendicular extension of the moraine on its south side. Note morianes d and e on the valley wall (Photo BJR, June 27, 1996).

a debris covered ice front close to this position which confirms the recent age of these features.

7.3.2.3 Zone of Coalescence Summary

Evidence from the zone of coalescence indicates three distinct periods of moraine construction: 1) the early 18th century, 2) the late 18th century and 3) the late 19th century. The earliest phase of glacial activity is recorded by the outermost moraine along the valley wall. Dates from this forested moraine indicate that the maximum position along the valley wall was achieved prior to 1740. A subsequent advance in the late 18th century resulted in the deposition of moraine e along the valley wall and the entire moraine complex downvalley of channel g. The landforms upvalley of channel g post-date the mid 19th century.

7.3.3 Athabasca Valley

The discussion of the results in the remainder of the Athabasca Valley can be divided into two sections: the valley floor area (Areas B and C) and the northern valley wall. Areas B and C consist of sparsely vegetated arcuate moraine systems from the Columbia Glacier that have been dissected by meltwater streams and are conspicuously fluted (Figures 7.4 and 7.15). Beyond the limit of these contiguous features, isolated "till islands" and several huge boulders occur in the outwash indicating that the LIA maximum position of the Columbia Glacier was some distance downvalley of the present glacial deposits. The landforms along the valley wall differ considerably from the terminal moraines in the centre of the valley. Near the downvalley limit of the discontinuous trimline, there are two small terraces lying against the north valley side (Figure 7.3, Figure 7.16) about 5 m above the valley floor which appear to have truncated a small 2-5 m high moraine approximately 20 m downslope of the trimline. The central 140 m of this moraine has been eliminated by a well developed avalanche track.

Trees were cored on many of these surfaces and establish minimum ages. The oldest tree (CN9603) revealed a minimum surface age of 1739 for the area between the trimline on the north valley side and the moraine at the foot of the slope (Table 7.6). The

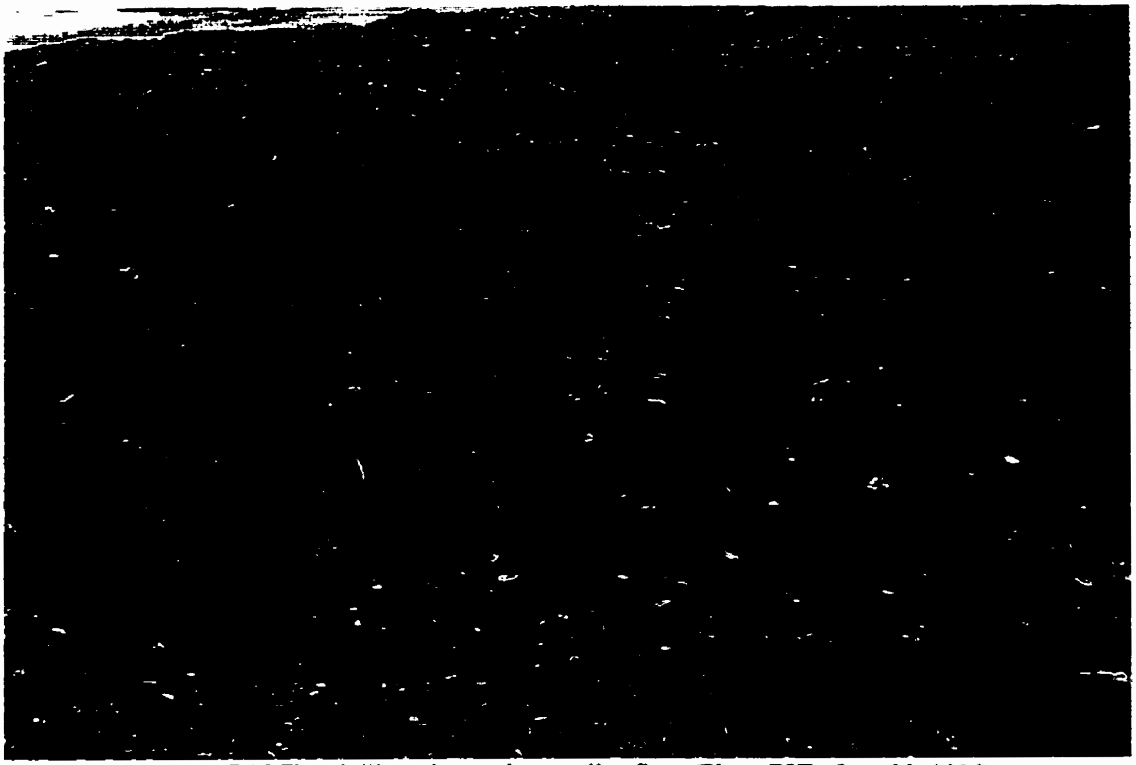


Figure 7.15 Fluted till on the northern valley floor (Photo BHL, June 28, 1996).

Figure 7.16 Outwash terrace located on the northern valley wall (Photo BHL, June 28, 1996).

oldest tree on this moraine and the adjacent terraces indicates a germination date of 1855 (CN9615). A similar age of 1852 (CC9602) was found for the oldest tree on the outermost "till island" on the main valley floor. This suggests that the most extensive, preserved terminal moraine remnants were deposited concurrently with the valley side moraine in the mid 19th century. Finally, the oldest tree sampled at the downvalley limit of contiguous till (Figure 7.3) yielded a pith date of 1907 (C96162), indicating that the moraine was deposited ca. 1887 (Table 7.6).

Table 7.6

Limiting Ages for the Central and Northern Athabasca Valley Floor

Location	Core ID	Inner Ring Date	Surface Date
Valley Side			
Between trimline and moraine	**CN9603**	**1761**	**1739**
Valley side terrace	**CN9615**	**1877**	**1855**
	CN9604	1886	1861
Valley side moraine	**CN9623**	**1887**	**1862**
Valley Floor			
Till island	**CC9602**	**1882**	**1852**
Outwash north of the river	**CN9682**	**1842**	**1822**
Outwash south of the river	**CC9604**	**1886**	**1866**
Downvalley limit of continuous till	**C96162**	**1907**	**1887**

Note: Sample ID's include year identification. Surface date includes an ecesis of 20 yrs. and a pith correction. Bold indicates the limiting date.

Several samples were also obtained from the few trees growing on the outwash between the moraine island (island f) and the downvalley limit of till on the south side of the river (Figure 7.3). These trees reveal an intermediate surface age of 1866 (CC9604), suggesting that the outwash was likely created prior to the glacier's advance to the limit of contiguous till. Several trees sampled on a similar outwash surface north of the river revealed that the oldest yielded a pith date of 1842 (CN9682). This indicates that the surface stabilised ca. 1822, providing a limiting date for those features downvalley. As the only notable feature downvalley of this position are the boulders marking the downvalley limit of glacier extent, this indicates that the ice has not reached this position since at least the early 19th century.

7.3.4 Evaluation of the Site Chronology

Results for the confluent parts of the Columbia-Manitoba forefield suggest four distinct periods of glacial advance: 1) Advance A in the late 15th century, 2) Advance B between 1698 and 1739, 3) Advance C prior to 1784 and 4) Advance D between 1808 and 1852. Most of these events are supported by several lines of evidence summarised in Table 7.7. The principal exception to this is Advance A which is based on an outer ring date of 1474 for a single log buried in the lateral moraine of Manitoba Glacier.

Evidence for Advance B, corresponding to Heusser's early 18th century advance, is preserved in many locations in both valleys. Outer ring dates from detrital logs found on the crest of the Manitoba lateral moraine indicate that the glacier was advancing into forest until at least 1698. This provides a maximum age for the surface as the trees were killed during advance and likely predates the moraine. Minimum surface ages between 1739 and 1770 based on trees within the trimlines are consistent with this early 18th century event. The earliest date was obtained from the northern trimline and indicates that the ice had retreated from this position by 1739. The maximum and minimum dates obtained suggest the advance occurred between 1698 and 1739. Despite all efforts, no damaged or killed trees were found at the trimline so a more precise date can not be determined.

The furthest downvalley limit of glacial advance preserved on the Athabasca Valley floor dates is attributed to Advance C. Unfortunately, the evidence for this event is limited to a small area of the Athabasca Valley and the Manitoba Valley. However, many of the limiting dates for this event are tightly clustered around 1785, corresponding to Heusser's (1956) advance of 1796. Several surfaces dating to the beginning of the 19th century were also believed to be part of this advance. Since the moraines corresponding to a mid 19th century event were found only a short distance upvalley of the late 18th century position, it suggests that the glaciers remained close to their earlier positions. The proximity of the ice likely had an adverse effect on the local environment, thereby increasing the ecesis interval.

The final major period of moraine building was Advance D, with maximum and minimum limiting ages of 1808 and 1852, respectively. Detrital wood from the Manitoba

Table 7.7

Summary of Data at Columbia and Manitoba Glaciers

Location	Surface Date
Advance A	
Imbedded in eastern lateral moraine of Manitoba Glacier	>1474
Advance B	
On surface of eastern lateral moraine of Manitoba Glacier	> 1698
North trimline (Heusser 1956, Athabasca Valley)	<1724
Northern trimline (Athabasca Valley)	<1739
Forested valley wall moraine (Zone of Coalescence)	<1740
Western trimline (Manitoba Valley)	<1741
Trimline at eastern junction (Athabasca Valley)	<1755
South trimline (Heusser 1956, Athabasca Valley)	<1763
Above moraine c (Manitoba Valley)	<1770
Advance C	
Main western lateral moraine of Manitoba Glacier	<1784
Moraine II (Zone of Coalescence)	<1788
Moraine fragment upvalley of moraine II (Zone of Coalescence)	<1788
Moraine I (Zone of Coalescence)	<1789
Above moraine e (Zone of Coalescence)	<1793
Inside outermost terminus on Athabasca Valley floor (Heusser 1956)	<1796
Moraine a (Manitoba Valley)	<1818
Moraine III (Zone of Coalescence)	<1818
Advance D	
On surface of Manitoba Glacier's eastern lateral moraine	>1808
Moraine b (Manitoba Valley)	<1837
"Unknown" (Heusser 1956)	<1842
Moraine island (Athabasca Valley)	<1852
Near north trimline (Heusser 1956, Athabasca Valley)	<1854
Northern moraine and terrace (Athabasca Valley)	<1855
Moraine c (Manitoba Valley)	<1864
Valley floor south of the river (Heusser 1956, Athabasca Valley)	<1864
Columbia moraine at eastern junction (Athabasca Valley)	<1871
Moraine IV (Zone of Coalescence)	<1877
Limit of till in the centre of the Athabasca Valley	<1887

lateral moraine crest provides a maximum date of 1808 for the advance. Moraines dating from between 1852-1877 are located inside the forefield limit defined by trimlines at numerous locations throughout the site. The single exception to this is the "till island" in the centre of the valley, beyond the limit of contiguous till. The timing of this event is consistent with the readvance reported by Heusser (1956) in 1842. Unfortunately, the location of his sample was not revealed and can therefore only be used for general comparative purposes.

Evidence from the Athabasca Valley indicates three main glacial advances of similar magnitude. Unfortunately, due to the fragmentary nature of the evidence, it is uncertain which was the definitive downvalley limit. On both valley walls, the trimline was created by an early 18th century event. However, the outermost moraines on the valley floor extend further downvalley than the trimlines. Furthermore, the most extensive glacial and fluvioglacial features throughout the valley are dated to different periods. The outermost moraine (moraine I) in the zone of coalescence is attributed to an event in the late 18th century, whereas the "till island" in the centre of the valley, located at an equivalent position downvalley, was created prior to 1852. However, trees growing on the outwash slightly downvalley of these features indicate that the glacier had not occupied such an extensive position since before 1822. Several explanations can be used to interpret this evidence.

The simplest explanation would be that the glacier reached its most extensive position at different places in the valley at different times. Another would be that each successive advance was slightly more extensive, eliminating most of the evidence from the previous ones. The evidence for the most recent advance was then eliminated by the extensive outwash. Unfortunately, dates from terraces on the valley floor indicate that they stabilised prior to the mid 19th century, which is inconsistent with this interpretation.

The most probable explanation is that the glacier achieved its maximum during one of the 18th century advances, likely the early 1700's. As the glacier retreated from its maximum position and subsequently readvanced to a slightly less extensive position, the meltwater destroyed the evidence of the earlier advance on the valley floor. A similar explanation can be used to account for the sparcity of features from the late 18th century event. As the glacier readvanced to its mid 19th century position, the outwash eliminated features from the previous events in the centre of the valley. This explanation is favoured for several reasons. Primarily, it accounts for the difference in the ages of the outwash terraces on the valley floor. It also explains the relationship between the outermost moraines and outwash terraces on the south side of the valley and their similar limiting ages. At their leading edges, it appears that the moraines have truncated the outwash surface; as would be expected if a glacier advanced over its own outwash. In these situations, it would be expected that the deposits beyond the toe of the glacier would be a similar age or slightly older. Finally, the "till island" in the centre of the valley has been completely isolated from the contiguous till through fluvial erosion, supporting the possibility of the elimination of moraines via fluvial activity in this valley.

7.4 Documented Ice Front Positions

Coalescence of the Columbia and Manitoba glaciers has prevented the production of separate LIA histories for the two glaciers. Historical documentation of 20th century ice front positions has enabled the identification of differences in the recessional patterns of these two glaciers. Therefore, each glacier will be discussed separately.

7.4.1 Columbia Glacier

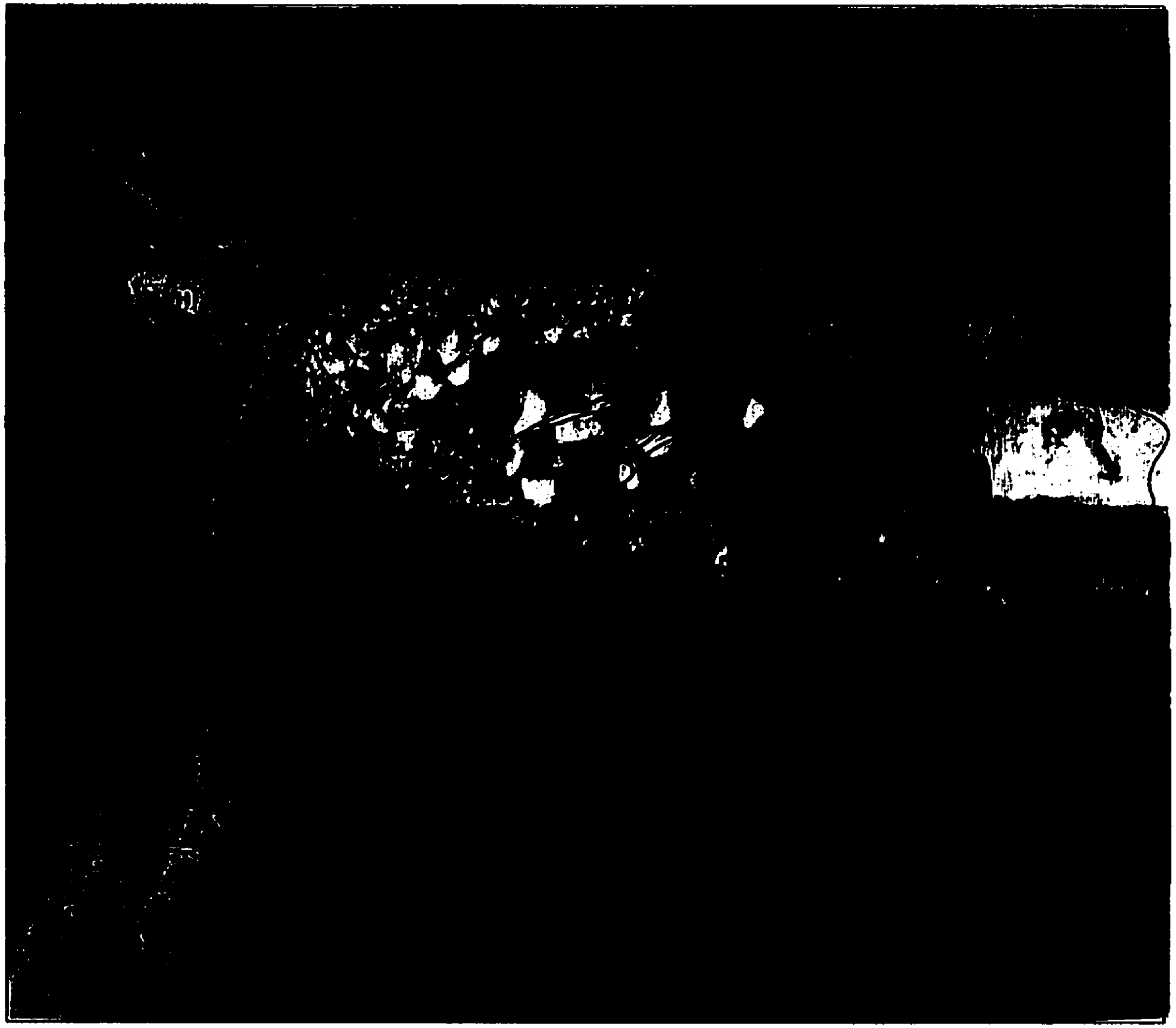
The first written account of the Columbia Glacier was given by Jean Habel, who observed it in 1901 (Habel 1902). His photo shows the glacier near its maximum position as defined by terminal moraines and the trimline on the south side of the valley. Photographs taken 6 years later by Schäffer (1908) show the glacier in a similar position, with the glacier less than 100 m back from its maximum position (pg. 40 in Schäffer 1980). Thorington (1925) used photographs taken in 1919 by the Interprovincial Boundary Commission to show that the glacier had retreated from its 1907 position but remained close to the trimline along both valley walls. Photographs taken by Howard Palmer in 1920 and 1924 show that the glacier had retreated approximately 75 m (in Field 1949). Furthermore, the ice surface had begun to lower significantly so that the valley sides below the trimlines were visible. Comparisons between Palmer's photographs and those taken by Field from the same position in 1948 indicate that Palmer's 1924 terminus was about 400 m from the outermost terminal moraine and approximately 825 m from the 1948 terminus (Field 1949). Additional observations in 1953 (Field & Heusser 1954) indicate that Columbia Glacier receded about 200 m in the 5 year period between these two visits. This suggests that Columbia Glacier has experienced a substantially greater amount of net retreat during the present century than it did in the 19th century.

Table 7.8

Recession of Columbia Glacier

Period	Distance of Retreat (m)	Rate of Retreat (m/yr.)	Source
early 18th century – 1924	400	-	Field & Heusser 1954
1920-1924	75	19	Field 1949
1924-1953	1025	35	Field & Heusser 1954
1948-1955	175	25	this study
1955-1966	330	30	this study
1966-1974	160	20	this study
1974-1976	-71	-36	this study
1966-1977	-1000	-91	Baranowski & Henoch (1978)
1976-1992	925	52	this study
1992-1996	400	100	this study

Beginning in 1948, aerial photography of the Columbia Glacier has been taken at irregular intervals. For most of this interval, the glacier has an irregular ice front that makes it difficult to provide simple measurements of ice front recession. Detailed ice front positions are shown in Figure 7.17. The values shown in Table 7.8 represent the average rate of retreat over the entire ice front. Between 1948 and 1974, the snout of the Columbia Glacier retreated about 650 m at rates ranging from 20 to 35 m/yr. (Figure 7.17). Although Baranowski & Henoch (1978) reported a frontal advance in the order of 1000 m between 1966 and 1977, their map indicates an advance of only 260 m. Furthermore, analysis of the air photos revealed a maximum advance of 300 m in the central portion of the glacier between 1974 and 1976. Following this brief readvance, the glacier retreated 925 m to its 1992 position. Between 1992 and 1996, the Columbia



0 500m
Scale is variable based on lower outwash

A - 1996	C - 1976	E - 1966	G - 1948	I - 1907
B - 1992	D - 1974	F - 1955	H - 1924	

N

Figure 7.17 Recent Ice Front Positions at Columbia and Manitoba Glaciers (NAPL A31609-38).

Glacier rapidly retreated an additional 400 m. This is a much more rapid change than previously documented.

The rapid rate of retreat of the Columbia Glacier during the 20th century may be related to the environment in which it terminates. Prior to 1924, the glacier was terrestrial based. However, all photographs taken after 1948 indicate that the glacier is actively calving large pieces of ice into a large terminal lake at the toe of the glacier. This shows that the glacier changed from a land to an aquatic based glacier between 1924 and 1948. This coincides with the increase in the rate of retreat (Table 7.8). However, ice front positions of floating glaciers do not necessarily reflect climatic influences since the amount of calving is largely controlled by the water depth (Sugden & John 1976).

7.4.2 Manitoba Glacier

Early observations of the Manitoba Glacier are considerably less frequent than for the Columbia as it is frequently only mentioned as a side note in discussions of the recessional history of the larger glacier. However, Manitoba's pattern of recession is substantially different than Columbia's (Table 7.9). By 1907 (pg. 40 in Schäffer 1980), the Manitoba Glacier had already receded approximately 915 m from the maximum terminal position marked by the downvalley limit of till in the Athabasca Valley. This was followed by a further 460 m retreat between 1907 and 1948 (Field 1949).

A historical photo taken in 1924 (Figure 7.18) shows the Manitoba Glacier approximately 250 m from the end of the Manitoba Valley. This photograph was used in the present study to calculate the amount of retreat between 1924 and 1948 (Table 7.9). Between 1924 and 1955, the glacier retreated a total of 400 m at an average rate of 13 m/yr. Recession then decreased to half its previous rate over a period of 11 yrs. From 1966 to 1976, the Manitoba Glacier retreated faster than during any other period in its history. Between 1976 and 1992, the glacier only receded about 50 m, reflecting downwasting rather than retreat.

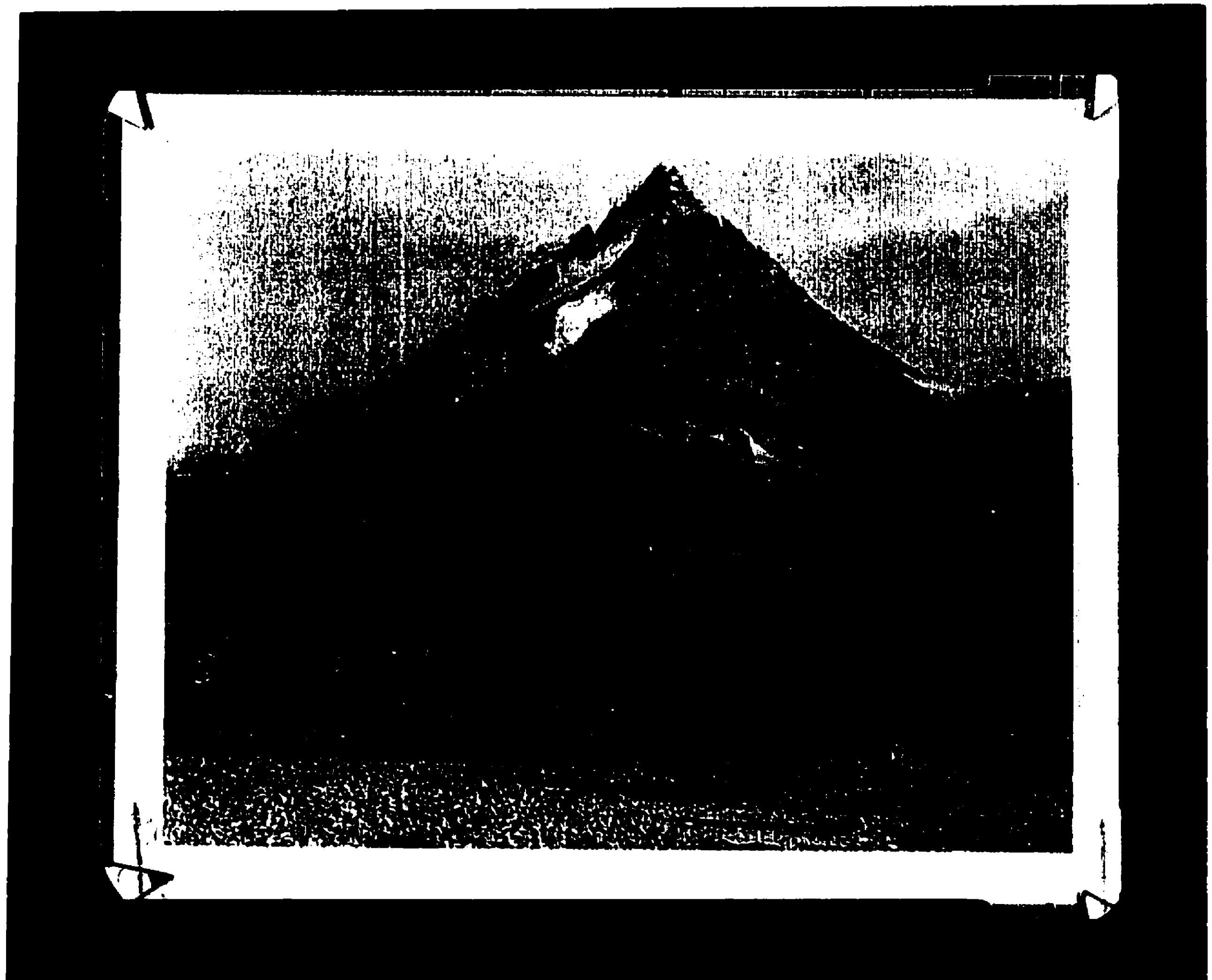


Figure 7.18 Columbia Glacier in 1924 (V14/ACOP-697, Whyte Museum of the Canadian Rockies).

Table 7.9

Recession of Manitoba Glacier

Period	Distance of Retreat (m)	Rate of Retreat (m/yr.)	Source
early 18th century - 1907	915	-	Field 1949
1907-1948	460	11	Field 1949
1924-1948	290	12	this study
1948-1955	105	15	this study
1955-1966	70	6	this study
1966-1976	220	22	this study
1976-1992	55	3	this study

7.5 Summary

The Columbia and Manitoba Glaciers coalesced at their maxima and it is impossible to isolate the history of the individual glaciers during the LIA. Dates obtained from both valleys indicate at least three advances of similar magnitude (Table 7.10): the early 18th century, late 18th century and the mid 19th century. Although the relative extent can not be conclusively determined due to the fragmentary nature of the evidence, it is likely that the earliest advance was the most extensive based on evidence from the trimlines. Trees killed by the glacier in 1698 and seeding within trimlines by 1739 in the Athabasca and Manitoba Valleys provide the maximum and minimum limiting dates for the early 18th century event. These dates are consistent with the early 1700's advance reported by Heusser (1956). Evidence corresponding to Heusser's (1956) late 1700's event was limited to the zone of coalescence on the southern valley floor of the Athabasca Valley and is defined by a minimum surface age of 1788. Many moraines in the Athabasca Valley can be dated to a mid 19th century advance. Heusser (1956) indicated a minimum estimate of 1842 for this event, however, detrital logs from the crest of the Manitoba lateral moraine provide a maximum age of 1808 for this event. This confines this advance between 1808 and 1842. Although the general timing of glacial events determined in this study does not differ significantly from those found by Heusser (1956), maximum limiting dates for two of the events were obtained from this study. Furthermore, evidence of an earlier advance of unknown extent was found in the lateral moraine of the Manitoba Glacier. This late 15th century event went unreported by Heusser (1956).

Although there are dated moraines that reflect early 20th century ice front positions, direct observations by researchers in the early 1900's are the most accurate sources for documenting retreating ice fronts. Furthermore, these type of observations enable the identification of differences in the recessional histories of the Columbia and Manitoba Glaciers. While the Manitoba Glacier had undergone considerable retreat between its maximum position and the first historical photos, the Columbia Glacier remained within several hundred metres of its downvalley limit. Furthermore, recession of the Manitoba Glacier has proceeded at a moderate rate, whereas the Columbia Glacier has undergone several distinct phases of retreat, with rapidly increasing rates of recession. However, these contrasts are likely a result of the difference in their terminal environments. The Manitoba Glacier is a heavily debris covered, terrestrial glacier that predominantly responds to changes in mass balance. On the other hand, the Columbia Glacier is a floating glacier with an actively calving terminus with water depth being a dominant factor controlling the amount of retreat.

Table 7.10

Major Periods of Glacial Advance at Columbia and Manitoba Glaciers

Date of Advance	Supporting evidence
after 1474	Outer date of overridden tree in east Manitoba lateral moraine
1698-1739	Outer dates of trees buried under large boulders on east Manitoba lateral moraine, oldest trees below 4 trimlines and oldest tree on valley side moraine
ca. 1785	Oldest trees on 5 moraines
1808-1852	Outer dates of trees buried under large boulders on east Manitoba lateral moraine, oldest trees on moraines

CHAPTER VIII
STUTFIELD GLACIER

8.1 Introduction

Stutfield Glacier is the large debris-covered glacier flowing northeastward from Stutfield Peak near the north end of the Icefield. The total area of the glacier is 21 km^2, although less than 25% of it makes up the outlet tongue below the 1992 snowline of 2050 m.a.s.l. (Robinson 1996). This is the lowest snowline observed on the Icefield (Robinson 1996). The accumulation area comprises most of the northern wing of the Icefield and the ice is transferred via avalanching and a single icefall (between 2100 and 2900 m.a.s.l) on the western flank of Mt. Kitchener to the lower, apparently stagnant, outlet portion of the glacier. The 3.5 km long glacier terminates at 1715 m.a.s.l., is heavily debris covered and dissected by many supraglacial drainage channels (Figure 8.1).

In the centre of the valley immediately beyond the glacier terminus there is a proglacial lake (Figure 8.2) draining eastward. Beyond the limit of the glacier forefield, approximately 1.3 km downvalley of the present ice front, the entire width of the valley is filled with coarse, unvegetated outwash. At the former limit of the glacier, the stream is incised slightly into this outwash. Flanking the glacier upvalley, there are well developed lateral moraines that rise steeply over 75 m above the main valley floor but have gentler distal slopes that rise 5-10 m above the adjacent valley sides. Downvalley of the current ice front, the north and south sides of the valley differ morphologically (Figure 8.3). On the north side of the valley, the lateral moraine passes downvalley into a noticeable trimline at a similar elevation before steeply "dropping off" to the valley floor. South of the river, however, the main lateral moraine divides into numerous smaller ridges and gradually descends to the valley floor.

The lateral moraine along the northern valley wall can be divided into two sections by a small creek (located on the western edge of Figure 8.3), approximately 350 m upvalley of the current ice front. Upvalley of this stream, the lateral moraine is a single ridge with a distal slope of 5 m (Figure 8.2). Downvalley of the creek, the landforms along the valley wall are more complex. For 350 m immediately downvalley of the creek,

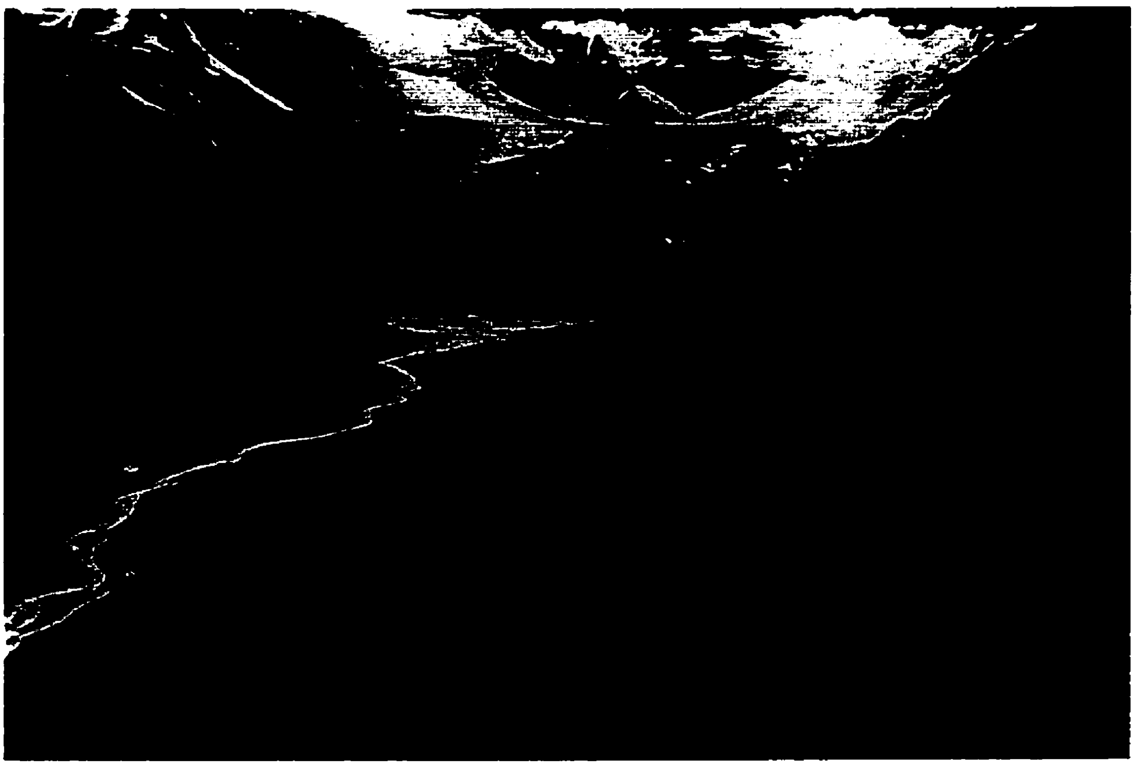


Figure 8.1 The Stutfield Glacier (Photo BHL, June 30, 1996)

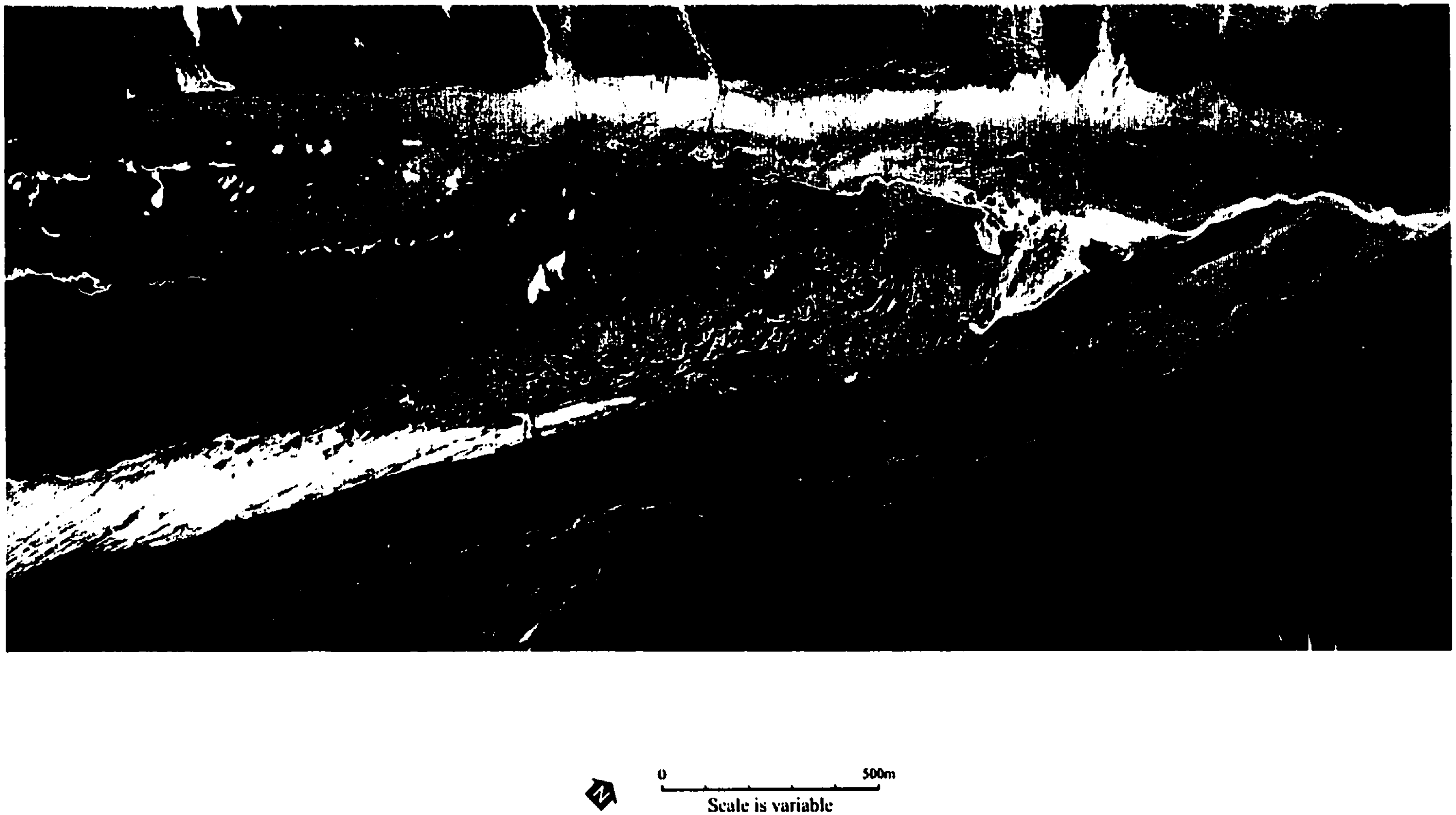


Figure 8.2 Aerial Photo of the Stutfield Glacier site (Natural Resources Canada, NAPL Photo Number A20887-142, July 1968).

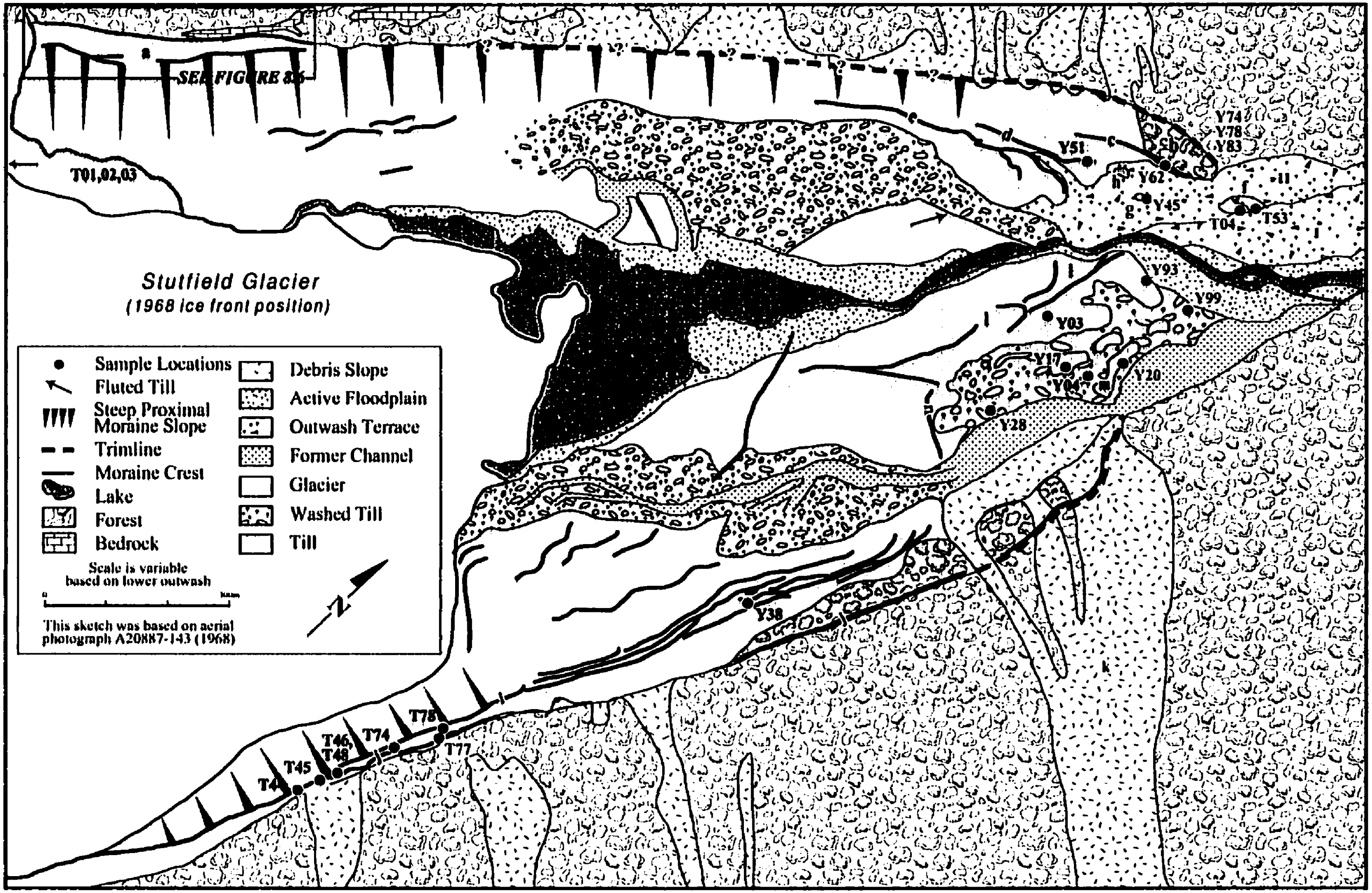


Figure 8.3 Stutfield Glacier forefield with samples discussed in text. Year identification has been excluded from all sample numbers. All samples shown were collected in the 1996 field season (e.g. T77=T9677).

there are two moraine ridges separated by a 75 m wide till "ledge" (Area a in Figure 8.3). The inner ridge is the main lateral moraine crest with a distal slope of about 1 m, while the outer ridge is 1-3 m high and its crest marks the limit of the forest. Downvalley, this "ledge" grades into till plastered against the valley wall with an indeterminant trimline obscured by avalanche tracks, minor stream channels and slope failures. Approximately 350 m upvalley from the limit of the forefield, the trimline reappears as a distinct feature and continues through the forest to the limit of the forefield.

Almost all glacial features on the northern part of the valley floor have been modified or eroded by fluvial activity. Three moraines below the trimline at the limit of the forefield (ridges c, d and e in Figure 8.3) have been truncated at their downvalley limits by the upper of two outwash terraces. The only remaining glacial features on the valley floor are three "islands" of till (f, g and h in Figure 8.3) that appear to have been isolated from the outer moraines by fluvial activity.

Upvalley of the current ice front on the south side of the valley, the lateral moraine is a single-crested feature with a distal slope about 8 m high. Downvalley of the present ice front, the lateral moraine splits into two distinct ridges that continue downvalley for 200 m until the outer moraine gradually disappears and the inner moraine divides into numerous individual ridges (Figure 8.3). Similar to the north side of the valley, these moraines are obscured in several locations by large avalanche tracks and debris-flow deposits. Approximately 700 m upvalley of the limit of the forefield, the outermost moraine reappears in the forest before being associated with a trimline and cutting across a large avalanche track at the limit of the forefield

On the valley floor, immediately south of the river, there is a well developed 15 m high sharp-crested moraine with several distinct ridges superimposed on it (Figure 8.1; moraine l, Figure 8.3). Although the area south of the river has also been modified by fluvial activity, most of this modification has been confined to distinct channels. Therefore, distinct glacial landforms are still evident on the south side of the valley.

8.2 Previous Investigations

Osborn (1993, 1996) has used tephra, wood fragments and *in-situ* tree trunks buried in the lateral moraines to determine early and pre-LIA glacial events at Stutfield Glacier. He found that the lateral moraines of Stutfield Glacier were not one discrete deposit, but composite features composed of superimposed tills of different ages. Investigations at the north lateral moraine indicate at least three separate advances (Osborn 1993). The presence of a layer of Mazama tephra indicates that the lower portion of the lateral moraine was likely deposited during the Crowfoot Advance of the late Pleistocene. Approximately 1.5 m of till was then deposited during an advance in the early Neoglacial, which overlies a layer of Mazama tephra (6800 yrs. BP) and has a soil and tephra complex containing St. Helens Y (4300 yrs. BP), Yn (3400 yrs. BP) and Bridge River tephras (2350 yrs. BP) above it. This tephra-rich paleosol also provided evidence for the beginning of the LIA; an *in-situ* trunk killed by a glacier advance 940 yrs. BP. These dates suggest the site remained ice-free between the early Neoglacial advance prior to 4300 yrs. BP and the onset of the LIA 940 ^{14}C yrs. BP (Osborn 1993).

Subsequent investigations by Osborn (1996) at the north and south lateral moraine provide more closely limiting dates for the LIA advance. In association with the present study, an *in-situ* trunk and a wood fragment were sampled in the north lateral moraine nearly 2 km upvalley of the LIA maximum position (see section 8.3.1.1). Radiocarbon dating of these samples indicates that Stutfield Glacier was within 8 m of its maximum height during the period 1050-1180 AD. On the southern lateral moraine, approximately 600 m downvalley, the ice did not reach within 20 m of its maximum height until after 1285 AD (Osborn 1996).

8.3 Results of the Present Study

The forefield of Stutfield Glacier is bisected by a stream in the centre of the valley. As the north and south sides of the valley differ morphologically, they will be discussed separately. Following a discussion of each individual area, the evidence will be summarised in the overall site chronology presented in section 8.3.3. Complete results from all samples collected at this site are presented in Appendix V.

8.3.1 Stutfield North

Approximately 350 m upvalley of the present ice front, the lateral moraine is breached by a small creek from a lake on the valley side. Upvalley of this creek, the lateral moraine is a single crested feature whereas immediately downvalley of the creek, a "ledge" of till (Area a) is present between the main lateral moraine and the outermost ridge. Downvalley of the "ledge", the lateral moraine continues for 200 m before grading into till plastered against the valley wall that is dissected by avalanche tracks, minor stream channels and/or slope failures (Figure 8.3). Although the upper limit of till is not always obvious, an indeterminant trimline is evident between the avalanche tracks. About 350 m from the limit of the forefield, the trimline reappears as a definite feature and continues through the forest to the downvalley limit of the forefield, where three moraine ridges can be observed below the trimline. Sampling on the north side of the valley was focused on three areas: 1) the north lateral moraine, upvalley of the creek, 2) the north lateral moraine "ledge" downvalley of the creek and 3) the northern limit of the forefield.

8.3.1.1 North Lateral Moraine

Immediately upvalley of the creek shown on the western edge of Figure 8.3, the northern lateral moraine is a single, sharp crested ridge with proximal slopes 75 m high and distal slopes of 5 m. Steep gullies dissect the inner slope of the north lateral moraine and have exposed several sections containing wood. Sampling with G. Osborn recovered three samples from a gully approximately 600 m upvalley of the current ice front (1996, Table 8.1). A 2 m high, *in-situ* stump (T9601) was rooted in a paleosol which had developed on a thin lacustrine deposit approximately 10-15 m below the moraine crest (Figure 8.4). A second log (T9602) was found protruding horizontally out of the moraine at a similar elevation, approximately 10 m upvalley of the stump. The final sample (T9603) was an *in-situ* stump in a paleosol located 10-15 m below the first two and 20-30 m upvalley. All three trees were successfully crossdated with the regional Athabasca chronology, yielding calendar dates shown in Table 8.1, consistent with the radiocarbon ages obtained by Osborn (1996) for the same samples (Table 8.1).

Figure 8.4 Tree T9601 after sectioning. Located several hundred metres upvalley of the creek on the north side of the valley. The horizontal layer at the level of the roots is a thin bed of lacustrine deposits (Photo BHL. June 21. 1996).

Table 8.1

Radiocarbon Dates for Logs Found in the North Lateral Moraine

Section ID	Calendar Dates	Radiocarbon Age (yrs. BP)	Source
T9601	849-1136	940 ± 60	Osborn 1996
T9602	959-1271	860 ± 70	Osborn 1996
T9603	894-1064	480 ± 60	Osborn, pers. comm., Aug. 1997

Note: Sample ID's include year identification.

The outer surface of the *in-situ* stump (T9601) was weathered and had clearly lost a considerable number of rings. This suggests that the tree was killed later than the outermost ring date implies. The second, smaller log (T9602) was better preserved and had lost fewer rings since it was buried in the moraine. Only half a cross section was obtained, but it yielded a nearly complete radius resulting in a more accurate estimate of its death date. Accommodating for the difference in radial preservation, the dates obtained from these trees suggest that both were likely killed by an advance in the late 1200's that is equivalent to those documented at Robson and Peyto Glaciers (Luckman 1995, 1996c).

8.3.1.2 Northern "Ledge"

Immediately downvalley of the creek on the northern valley wall, two major ridges are evident along the valley side (Figure 8.3). The outer moraine ranges between 1 and 3 m high and delimits the boundary with the forested valley walls. At its upvalley limit it is a broad ridge with several small crests. Downvalley, it grades into a single, sharp crested ridge approximately 3 m high. The inner moraine is the main lateral moraine crest with a distal slope of less than 1 m high. The area between these two ridges can be characterised as a 75 m wide till "ledge" that gently slopes (10-15°) into the centre of the valley before dropping off steeply at its inner edge (the main lateral moraine) to the valley below (Area a in Figure 8.3, Figure 8.5). This "ledge" is composed of a blocky substrate, small (< 2 m high), almost indistinct ridges with a very sparse, stunted vegetation cover. The slope adjacent to the outer moraine is thickly forested with several bedrock outcrops.

Figure 8.5 Downvalley view of the "ledge" along the north lateral moraine at Stutfield Glacier. The small ridge on the right is the main lateral moraine (Photo BJR, July 22, 1996).

8.3.1.2.1 Damaged Trees

Investigations on the "ledge" yielded one scarred and two tilted trees (Table 8.2), providing precisely limiting dates for the presence of ice. T9654 was a recently dead tilted tree found approximately 100 m downvalley of the creek, at the base of the 1 m high outer ridge (Figure 8.6). The tree was tilted at an angle of 30° and was growing out of the base of the moraine (Figure 8.7). Although two cross sections were taken, at 30 and 200 cm above the ground surface, the lower part of the tree and the inner portion of the upper section were rotten, so only the upper section was retained. Most of this section was affected by a reaction wood series that began in 1758, indicating that the original tilting event occurred on this date. Subsequent minor events noted in Table 8.2 attenuated and slightly altered the axis of the existing reaction wood series. However, the last event, in 1834, coincided with traumatic resin canals. These events may have been the result of settling debris on the moraine or subsequent glacial events.

Approximately 50 m downvalley of this location, two more damaged trees were found at the base of the distal face of a 1.5 m high outer moraine, below a bedrock outcrop. Initial observations revealed a tilted tree (T9656) located at the base of the distal slope, rooted below the moraine (Figure 8.6). Further investigation revealed a second tree (T9655) between T9656 and the moraine that had fallen away from the moraine and was being held up by a branch of T9656 (Figure 8.8). Upon excavation, a basal scar was observed on T9655 (Figure 8.9), caused by a large rock embedded in the adjacent moraine surface (the tree had subsequently grown around the rock). This tree appeared to be in growth position, although all portions of the tree below the surface were rotten.

Although two cross sections were attempted from the tilted tree (T9656), the lower portion of the tree was completely rotten so only the section from 0.75 m above the ground surface was retained. As the centre of the upper section was also rotten, the pith was not obtained, although a successful crossdate was obtained. However, the earliest observable ring shows evidence of reaction wood, indicating that the tree was tilted prior to 1779 (Table 8.2). Unfortunately, the precise inception date could not be determined since the inner rings were missing.

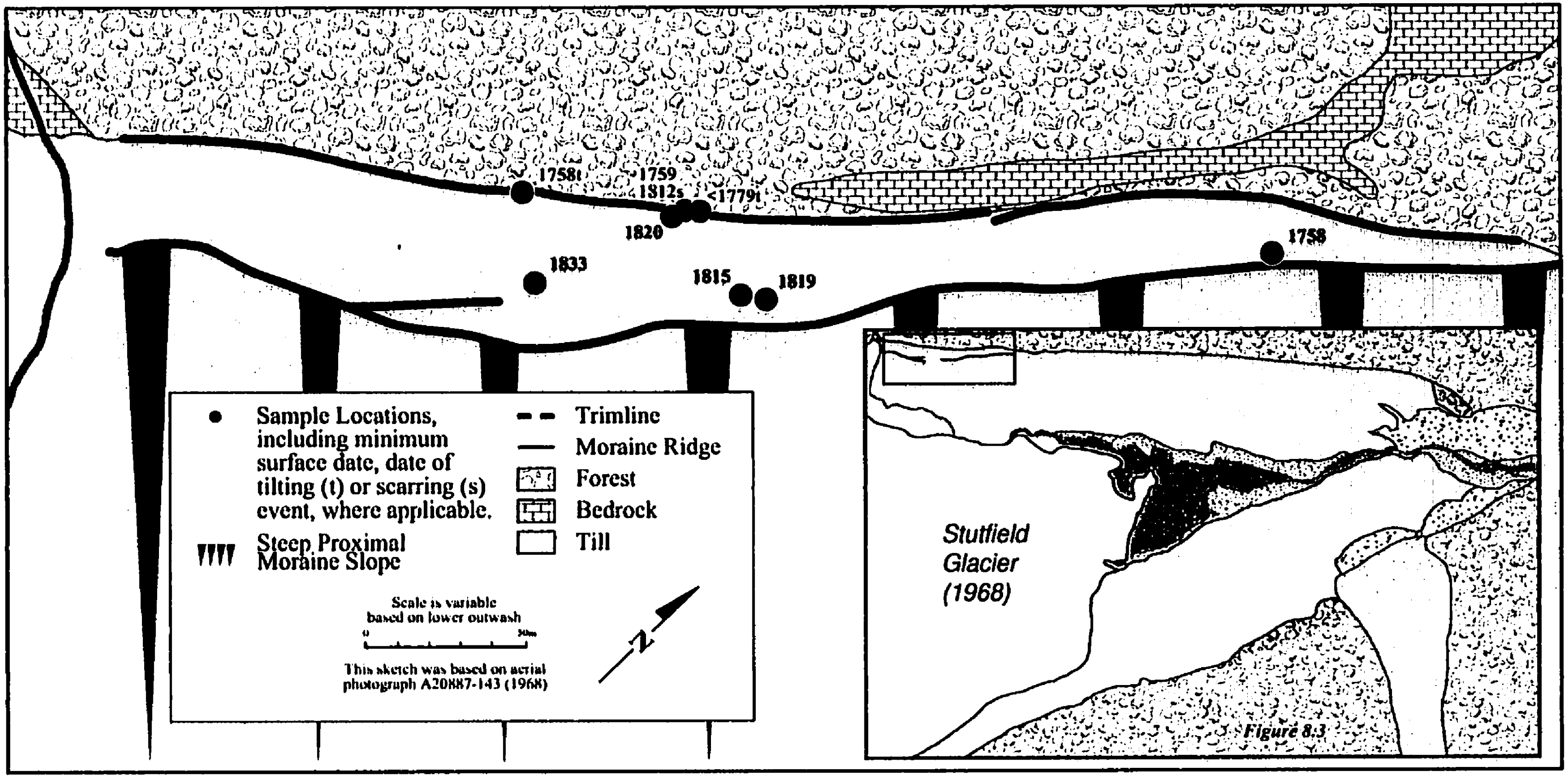


Figure 8.6 Sample locations and event summary for the northern till "ledge."

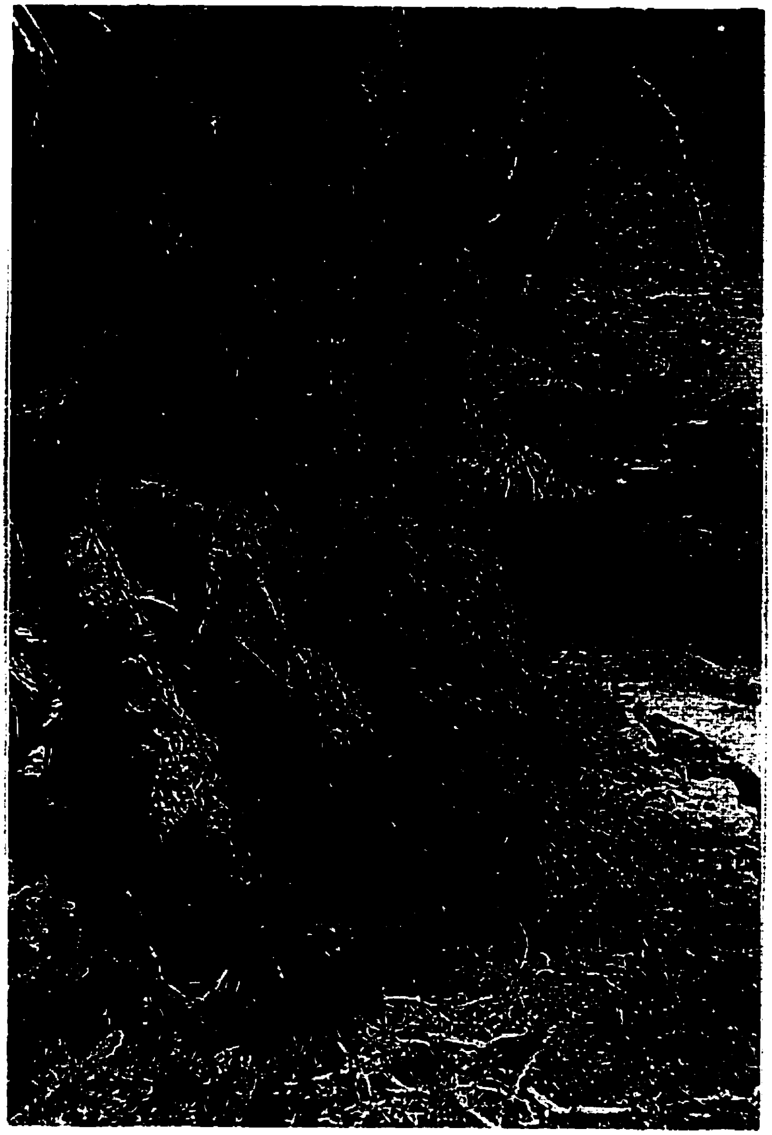


Figure 8.7 Tree T9654 tilted out of the outermost moraine on the north side of Stutfield Valley (Photo BJR, July 22, 1996).

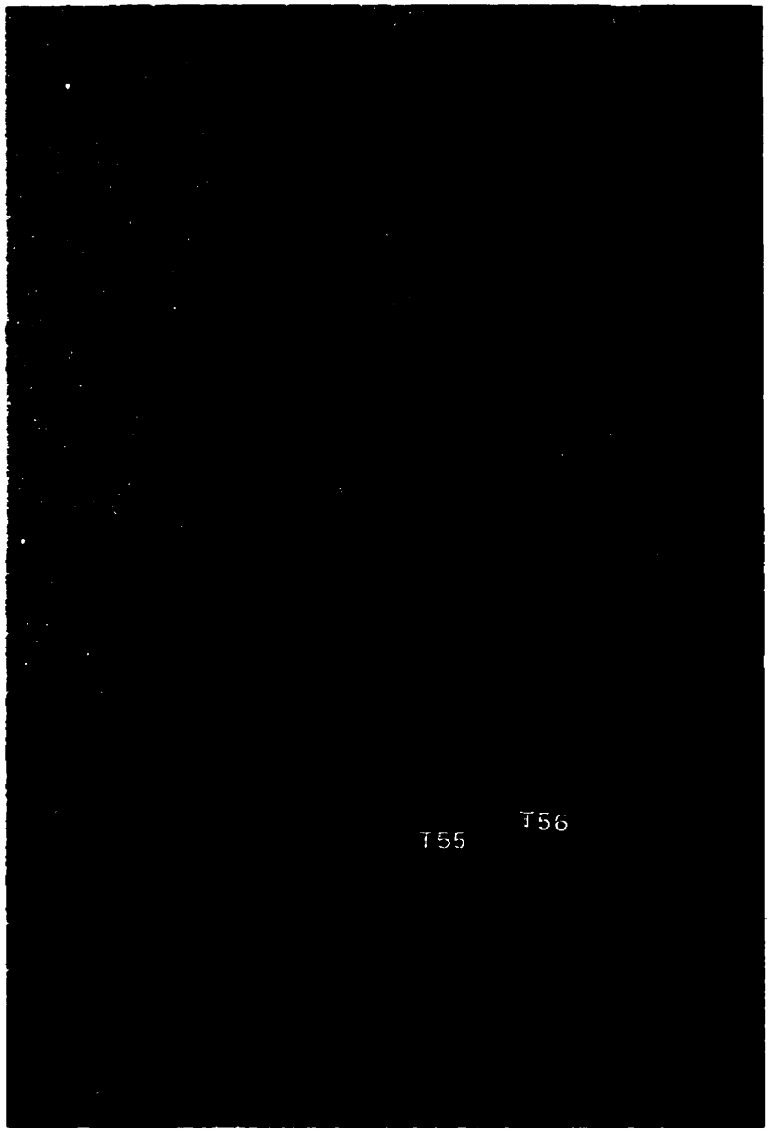


Figure 8.8 Trees T9656 and T9655. T9656 is the large, bark covered tree on the right side of the photo while T9655 the stripped trunk tilted 30° from vertical that is partially obscured by T9656 and a small sapling. Outer moraine is the 1.5 m high ridge located on the left of the photo (Photo BJR, July 22, 1996).

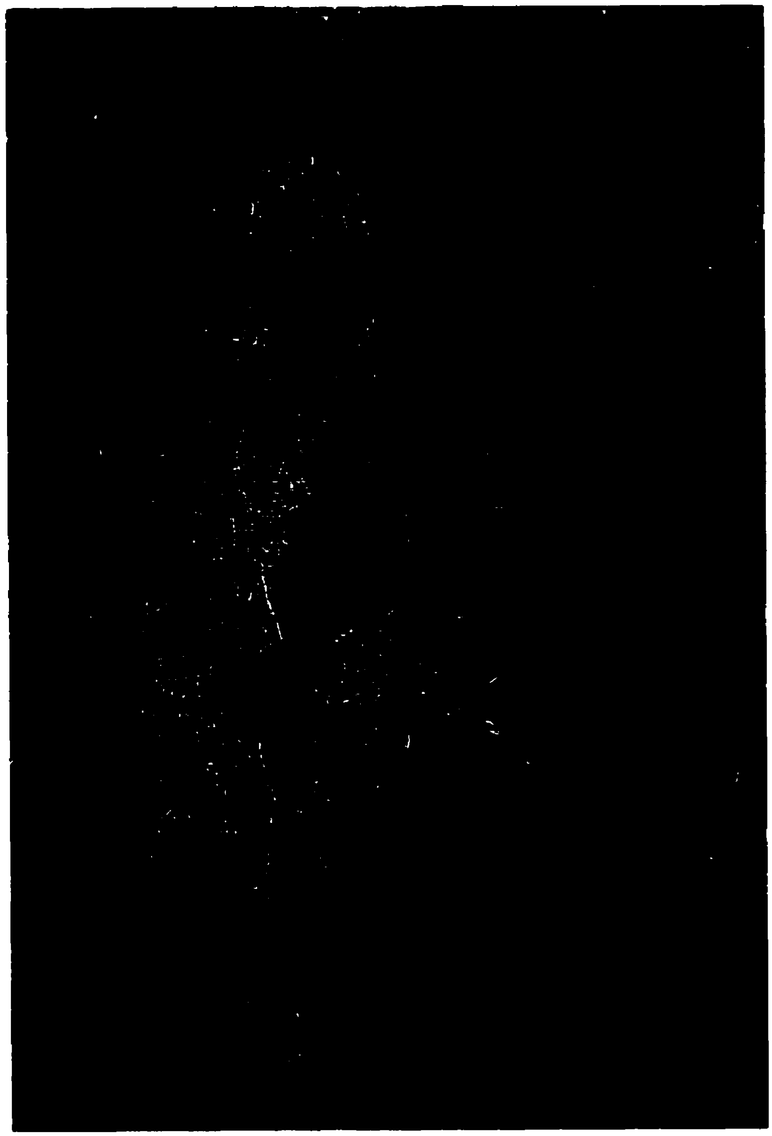


Figure 8.9 The scarred section of tree T9655 after exhumation. The scar roughly matches the shape of the rock around which the tree grew (Photo BJR, July 22, 1996).

Table 8.2

Damaged Trees Along the Northern Lateral Moraine

Section ID	Calendar Dates	Damage Date
T9654	1694-1995	1758, 1781*, 1787*, 1816*, 1821*, 1834*
T9655 Upper	1781-1868	1812
T9655 Lower	1779-1868	1812, 1815s, 1818s, 1831s*, 1834s*
T9656	1779-1963	<1779

Note: Sample ID's include year identification. s - scar dates, all other damage dates refer to initiation of reaction wood series. * - minor event

Germination and damage dates from the scarred tree (T9655) provide evidence for two glacial events, consistent with dates obtained from the other damaged trees. Two cross-sections were taken from this tree, one at the ground surface (Figure 8.10) and another 0.75 m above the ground surface, beyond the scar. Through successful crossdating, a pith date of 1779 was obtained for the lower section (Table 8.2). As a result of this site's proximity to the forested valley sides, an ecesis interval of 20 yrs. was applied, resulting in a surface age of 1759. This establishment date is consistent with the mid 18th century advance documented by the two tilted trees discussed previously. Since this tree appears to have been rooted in the lower part of the moraine, it suggests that T9655 was one of the primary colonisers following the mid 18th century advance.

Evidence of an early 19th century event was obtained from both cross sections of T9655. In the upper section, a major reaction wood series beginning in 1812 was observed. Damage to the lower section included a major tilting event starting in 1812 as well as major scarring event in 1815 (Figure 8.10, Table 8.2). The presence of a large rock in contact with the scar surface suggests that the impact of the rock against the tree during the formation of the moraine resulted in the scar. Although several subsequent minor scars were observed (Table 8.2), all were located along the same radius as the initial scar. In conjunction with the fact that the regrowth near the scar mimics the shape of the rock, this suggests that these scars were the result of the same rock. Therefore, T9655 became established on the surface of the moraine created by a glacial advance in the mid 18th century. The production of a moraine during a subsequent advance in the early 19th century then resulted in the tilting and scarring of this tree.

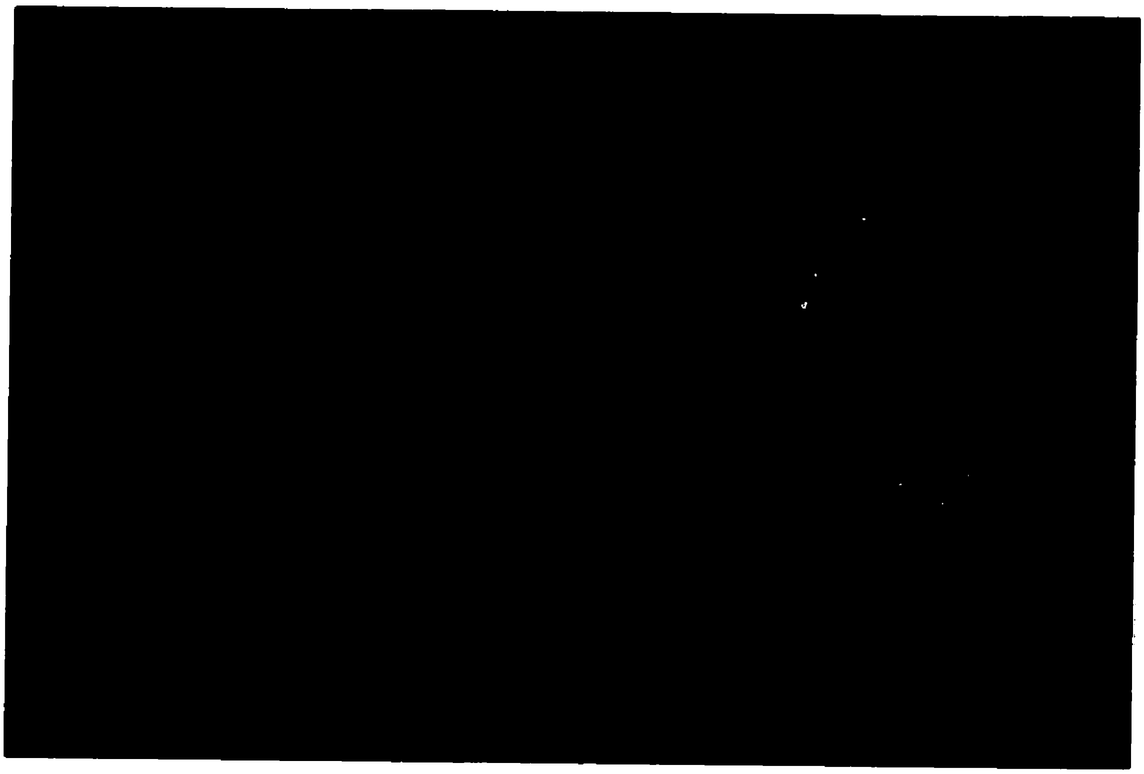


Figure 8.10 Sample T9655-lower. The arrow indicates the direction of glacier movement. a indicates the 1812 tilting event, b indicates the 1815 scar. The scarred portion of the tree was adjacent to the distal slope of the moraine and the reaction wood on the right side of the section indicates that the tree was tilted downvalley (Photo Ian Craig, August 1997).

Another tree (T9720) tilted out of the outer moraine was found approximately 10 m downvalley of T9655 and T9656. As the tree was still alive, only cores were taken. Unfortunately, there was no evidence of a reaction wood event in any of the cores and since it was rooted below the moraine, the surface upon which it was growing was uncertain. Therefore, the age of this sample would not provide definitive evidence for any glacial events. The precisely dated glacial events documented by the three damaged trees indicate at least two distinct periods of glacial activity; the mid to late 18th century and the early 19th century (more precisely between 1812 and 1821). Although there appears to be evidence from two trees for an event in 1834, the damage is located along the same radius as the earlier events. This suggests that the damage was an attenuation of the original tilting and scarring event caused by localised movement of debris adjacent to the trees rather than a discrete event.

8.3.1.2.2 Minimum Surface Ages

Living trees were also sampled along the "ledge" to verify the ages determined from the damaged trees. Only a limited number of trees were available for sampling as a result of the sparse vegetation cover. However, a definite age difference was observed between the trees sampled on the upvalley and downvalley portions of the "ledge". Trees from the upvalley portion indicate surface dates from the early 19th century whereas trees downvalley suggest a mid 18th century age.

A living tree with a pith date of 1840 (T9632) was sampled along the crest of the outermost moraine, several metres from T9655 and T9656 (Figure 8.6). Employing an ecesis interval of 20 yrs., this results in a surface age of 1820, corresponding to the early 19th century event (Table 8.3). Approximately 40 m downslope, three trees near the inner crest (Figure 8.6), suggest the surface had stabilised by the early 19th century. Near and off pith dates of 1850, 1859 and 1873 correspond to surface ages of 1815, 1819, and 1833 (T9711, T9712, T9710). There were no trees located on this feature further upvalley. Near the downvalley limit of the "ledge", a pith dated tree (T9657) indicates a minimum surface age of 1758 (Figure 8.6, Table 8.3).

Table 8.3

Limiting Ages for Northern "Ledge"

Location	Core ID	Inner Ring Date	Surface Date
Outer moraine, adjacent to damaged trees	T9632[a]	1840	1820
Moraine "ledge", downslope of damaged trees	**T9711[b]**	**1850**	**1815**
	T9712[b]	1859	1819
	T9710[b]	1873	1833
Near the downvalley limit of the "ledge"	**T9657[b]**	**1788**	**1758**

Note: Sample ID's include year identification. Surface date includes ecesis and pith corrections; superscripts: a - an ecesis interval of 20 yrs., b - an ecesis interval of 30 yrs. Bold indicates the limiting date.

The lack of a terminal ice position and the dates obtained through minimum surface ages may suggest that the "ledge" was deposited as a contiguous feature during a single event in the mid 18th century. This would mean that the trees growing on the upvalley portion of this feature have a much longer ecesis interval than the trees downvalley. Unfortunately, this is inconsistent with the damaged trees found on the outer moraine. These trees indicate tilting and scarring events in both the mid 18th and early 19th century, supporting the idea that the glacier occupied a similar position at least twice. In conjunction with the minimum surface ages, it appears that only the upvalley portion of this feature was re-occupied by the glacier during the 19th century event. The absence of a distinct moraine ridge between the up and downvalley portions of the "ledge" may reflect a diffuse ice front or a short pulse of activity reworking previous deposits. Alternatively, one of the minor ridges along the "ledge" may represent the early 19th century ice front.

8.3.1.3 Northern Limit of the Forefield

About 350 m upvalley of the limit of the forefield, the trimline reappears as a definite feature and continues through well developed forest to the downvalley limit of the forefield. The area immediately inside the trimline, at the downvalley limit of the forefield is characterised by a mature forest and low, hummocky mounds, completely lacking distinct ridge crests (Area b). Upvalley of this area are three low ridges (ridges c-e in Figure 8.3) that appear to be continuous with the moraine islands located on the uppermost outwash terrace. The outer two moraine ridges (c & d) can be identified on aerial photographs by distinct rows of trees (Figure 8.2). At their downvalley limits, they are subdued, weathered ridges between 1 and 2 m high with well developed vegetational understories. Upvalley, they are sharp, low ridges (<1 m high) that become obscured by material that has eroded from the adjacent steep slopes and the large avalanche track, thereby restricting sampling to the area downvalley. The innermost moraine (ridge e) differs from the outer two in that it is a sharp crested feature approximately 2 m high with no vegetation.

Below the trimline, the oldest tree sampled has a near pith date of 1789 (Y9683), suggesting a surface age of 1767 (Table 8.4). This tree was located just below the trimline near its downvalley limit where it extends to the valley floor. Several other trees located on the lower, hummocky portion suggest similar surface ages of 1771 and 1774 (Y9674 and Y9678). Although several trees were sampled along the steeply sloping portion of this area, the dates obtained were considerably younger, likely a result of the instability of the slope. The oldest tree on ridge c (Y9662) was found halfway up the proximal slope approximately 40 m upvalley from its downvalley limit (Figure 8.3). This pith dated tree indicates that the surface had stabilised by ca. 1778. The close proximity and similarity in age of these two features suggests that they were formed during the same event in the late 18^{th} century.

Table 8.4

Limiting Ages for the Northern Valley Wall

Location	Core ID	Inner Ring Date	Surface Date
Below trimline	**Y9683**	**1789**	**1767**
	Y9674	1801	1771
	Y9678	1799	1774
Moraine c	**Y9662**	**1798**	**1778**
Moraine d	**Y9651**	**1832**	**1802**

Note: Sample ID's include year identification. Surface dates include an ecesis estimate of 20 yrs. and a pith correction. Bold indicates the limiting date.

The oldest tree sampled on the ridge d yielded an off pith date of 1832 (Y9651) and was located on the extreme downvalley limit of the moraine. This suggests that middle ridge was formed prior to 1802 (Table 8.4). As the innermost moraine (ridge e) was completely barren of trees, a limiting date could not be determined. However, its location suggest that this ridge represents a mid 19^{th} century event.

On the valley floor, most of the glacial features have been destroyed by fluvial activity that created several well developed outwash terraces that fill the valley. However, there are three "islands" of till (f-h in Figure 8.3) on the upper outwash terrace near the limit of the forefield that appear to be continuous with the outermost moraines on the valley side. The island furthest downvalley (Island f in Figure 8.3 and Figure 8.11) seems to be an extension of the area b, while the two smaller islands (Islands g & h) appear continuous with ridge d. Both islands f and g are composite features composed of a blocky, till "mound" with surrounding terraces. The terraces are approximately 1 m above the surrounding outwash terrace and are composed of fine grained material. They appear to be surfaces that trimmed the moraines, thus indicating that they postdate the formation of the adjacent till surface.

Island f is composed of two till levels with a small terrace on its northeast side that has undercut part of the till slope (Figure 8.11). The oldest tree on the upper level suggests a surface age of 1803 (T9604), which corresponds with surface stabilisation dates of 1803 and 1808 (T9609 and T9610) on the lower section (Table 8.5). However, older trees were found on the small terrace on the northeast side of the moraine. Living trees with pith and near pith dates of 1825, 1827 and 1828 (T9623, T9615, T9618) and a standing dead snag with a pith date of 1807 (T9653) suggest that the terrace was created prior to 1777 (Table 8.5). Since it appears that the terrace postdates the formation of the moraine, these samples provide a limiting date of 1777 for the terrace and the moraine. While neither island g nor h have trees on their crests, there is a single tree growing on the small terrace surrounding island g. The near pith date of 1838 (Y9645) obtained from this tree indicates that the surface had stabilised prior to 1803 (Table 8.5).

Minimum ages for the two dated islands were determined from trees located on adjacent terraces. Trees sampled on the till portion of island f did not provide a closely

Figure 8.11 View downvalley of moraine island f. The adjacent terrace where T9653 was located is on the left side of the island (Photo BJR, July 29, 1996).

limiting age for the feature, while island g was devoid of vegetation. The older date from the terraces which must slightly postdate the abandonment of the till islands suggest that ecesis is slightly longer on the till than on the outwash surface. This is a plausible explanation considering the variation between the substrate of the till and terrace components. The main till portions of the islands are composed predominantly of large, blocky boulders whereas the adjacent terraces are fine grained alluvium. The proximity of these surfaces ensures that most other factors controlling tree colonisation (e.g. proximity to a seed source and prevailing climatic conditions) are constant. Under these circumstances, it is expected that a fine grained terrace would provide a more suitable environment for seed germination and would therefore have a shorter ecesis interval.

Table 8.5

Limiting Ages for the Northern Valley Floor

Location	Cores			Cross Sections		
	Core ID	Inner Ring Date	Surface Date	Section ID	Section Range	Surface Date
Island f	T9604	1835	1803			
	T9609	1838	1803			
	T9610	1840	1808			
Outwash of Island f	T9623	1825	1795	**T9653**	**1807-1967**	**1777**
	T9618	1828	1796			
	T9615	1827	1797			
Island g	**Y9645**	**1838**	**1803**			

Note: Sample ID's include year identification. Surface dates include an ecesis estimate of 30 yrs. and a pith correction. Bold indicates the limiting date.

The dates obtained from these till islands support the possibility that the moraines along the valley wall extended down onto the valley floor. The similarity in the ages of the trimline (Area b), the outermost moraine (ridge c) and the eastern moraine island (Island f) suggests that all these features were formed during the same event in the late 18th century. Similarly, dates obtained from island g suggest that it was deposited concurrently with ridge d at the beginning of the 19th century. A third event is also recorded along the northern valley wall, although a limiting date could not be obtained for it. However, the position of ridge e in relationship to the outer ridges suggests that it was formed by a subsequent event during the 19th century.

8.3.1.4 Stutfield North Summary

Evidence from the north side of the valley suggests at least four periods of glacial advance. An event in the late 13th century was recorded by several stumps located in the north lateral moraine, about 600 m upvalley of the current ice front. On a till "ledge" on the valley wall, tilted and scarred trees provide precise dating control for at least two separate events, one in the mid 18th century (1758) and another at the beginning of the 19th century (1812-1820). These values are consistent with minimum surface ages on the same "ledge" as well as dates obtained at the limit of the forefield. Ages for the trimline (1767), outer moraine (ridge c, 1778) and outer moraine island (island f, 1777) are roughly consistent with the mid 18th century event. Similarly, the middle moraine (ridge d, 1802) and the upvalley till island (island g, 1803) are consistent with the early 19th century event along the valley wall. The discrepancy between the ages of these events at the north lateral moraine and the limit of the forefield may be attributed to the fact that the maximum position was not achieved concurrently along the entire margin of the glacier. It may also reflect variations in the ecesis interval for the different areas in the forefield. The final event recorded on the north side of the valley must have occurred during the 19th century, although the exact timing is uncertain.

8.3.2 Stutfield South

Upvalley of the present ice front, the southern lateral moraine is a sharp crested ridge, nearly devoid of vegetation, with steep 75 m high inner slopes. Compared to the northern lateral, this moraine has a prominent crest and appears to have been built up between the ice front and the adjacent forest, eliminating the possibility of tilted or killed trees. The distal slope gradually diminishes from approximately 8 m upvalley to several metres high above the current ice front. At this position, a small ridge appears outside the main lateral moraine while the main lateral moraine splits into numerous smaller ridges that descend to the valley floor (Figure 8.3). The central 400 m of the outermost ridge is ill-defined, observed mainly as a transition between till and the substrate in the forest beyond, it is evident as a small ridge in the forest downvalley and is associated with a trimline at the limit of the forefield. Below the trimline, there are no distinct moraine ridges on the valley floor. However, fluvially modified till covers the southern side of the valley. The different landforms found on the valley wall and the valley floor necessitates that the discussion be divided into two sections: 1) the southern valley wall and 2) the southern valley floor.

8.3.2.1 Southern Valley Wall

Above the current ice front, a small ridge about 2 m high (ridge j) appears from beneath the main lateral moraine (ridge i, Figure 8.3). The two moraines merge and diverge for 100 m downvalley until ridge j turns towards the forest. A further 100 m downvalley, where ridge j enters the forest, it has diminished in height to the point that it is no longer evident as a distinct ridge. Ridge i gradually decreases to about 1 m in height and descends the valley wall before subdividing into numerous small (~ 1 m high), discontinuous ridges about 350 m downvalley of the ice front. Nearly 300 m downvalley, a moraine ridge about 1 m high (assumed to be ridge j) is evident in the forest. It continues for 350 m through the forest as a narrow ridge until reaching a major avalanche track (Area k) at the limit of the forefield. It continues through the avalanche track as a ridge with a subdued crest, possibly modified by avalanche activity, and is associated with a trimline (Figure 8.3). Near the limit of the forefield, on the downvalley side of the avalanche track, both the outer moraine and the trimline descend steeply to the valley floor.

Above the current ice front, two trees located in a small depression between ridges i and j revealed pith and near pith dates of 1773 and 1785 (T9646 and T9648), indicating ridge j was created prior to 1743 (Table 8.6). A supporting surface date of 1767 was obtained from two pith dated trees approximately 175 m downvalley on ridge j (Figure 8.3). T9677 was found just below the crest on the distal slope, while T9678 was found in a small depression at the base of its proximal slope. These dates indicate that the ridge j pre-dates 1743 and was likely created in the early 18^{th} century.

Limiting dates from various locations along moraine i indicate that it was formed in the early 19th century (Figure 8.3, Table 8.6). A near pith date of 1842 (T9674) was obtained from a tree halfway up the distal slope of the inner moraine (ridge i) approximately 150 m downvalley of the position where moraine i truncates moraine j (Figure 8.3). This provides a minimum surface age of 1807. Consistent dates were found from trees up and downvalley of this position. Two trees located on the crest of the main lateral moraine, where there is only a single ridge, reveal near pith dates of 1855 and 1859 (T9645 and T9644). Approximately 250 m downvalley of where moraine i subdivides into several ridges, a tree with a near pith date of 1850 (Y9638) was sampled. This tree was located just inside the uppermost discontinuous moraine (Figure 8.3). The similarity in the ages for the trees located along the entire length of the main lateral moraine suggests that the glacier last occupied this position in the early 19th century.

Table 8.6

Limiting Ages for the Southern Valley Wall

Location	Core ID	Inner Ring Date	Surface Date
Outermost moraine (moraine j)	**T9646**	**1773**	**1743**
	T9648	1785	1750
	T9677	1797	1767
	T9678	1797	1767
Main lateral moraine (moraine i)	**T9674**	**1842**	**1807**
	Y9638	1850	1815
	T9645	1855	1820
	T9644	1859	1824

Note: Sample ID's include year identification. Surface dates include an ecesis estimate of 30 yrs. and a pith correction. Bold indicates the limiting date.

8.3.2.2 Southern Valley Floor

Although most of the glacial features on the southern valley floor have been fluvially modified, they have not been completely destroyed as is the case north of the river (Figure 8.3). The dominant feature on the southern valley floor is a 15 m high, sharp crested moraine parallel to the main axis of the valley (ridge l). Superimposed on this moraine are several minor ridges; unfortunately, the almost complete lack of vegetation prevents the determination of the age of any of these features. The northern slope of ridge l has been truncated by the river while the floor of the valley south of the moraine is a complex of distinct meltwater channels, fluvially modified till and numerous patches of unmodified till (Figure 8.12). There is a distinct contrast in vegetation cover between the areas up and downvalley of ridge n (Figure 8.3). Upvalley, the valley floor is devoid of vegetation, whereas the area downvalley has a variety of vegetation types (Figure 8.12). The areas of washed till are covered with mosses, prostrate shrubs and a limited number of trees although the trees are concentrated on the small "mounds" of blocky, unmodified till (many of which are too small to be observed in Figure 8.3.) with little to no moss or shrub cover.

Sampling on the southern valley floor revealed at least two advances: 1) in the late 18^{th} century and 2) in the early 19^{th} century. Samples taken from the unmodified till "mounds" reveal consistent dates throughout the area (Table 8.7). A tree located on the leading edge of the largest individual "mound" (moraine m) suggests that it was created before 1779 (Y9620). Similar dates were obtained from trees on two small, isolated unmodified moraines located 20 and 60 m upvalley from moraine m. Y9604 was located on a 1 m high moraine ridge that was only several metres wide and Y9617 was one of several trees on a small ridge (< 1 m high). Both moraines were completely surrounded by washed till. Living trees growing on the modified till suggest that the surface did not stabilise until the early 19^{th} century. The oldest tree, with a near pith date of 1841 (T9699) was found near the leading edge of the washed till, while a similar aged tree was found near its upvalley limit (Y9628, Table 8.7). These trees provide limiting dates of 1809 and 1819, respectively. These values suggest that the till was deposited during the late 18^{th} century, but portions were being modified by meltwater from the glacier until the early 19^{th} century.

A major glacial advance in the early 19^{th} century created the large moraine in the centre of the valley (area l). A tree growing on the southern slope of area l, near its base, had a near pith date of 1853 (Y9603), indicating a minimum surface age of 1818. This age is roughly consistent with the stabilisation of the modified till area on the valley floor. In conjunction with the fact that the south slope of area l gently grades into the area of

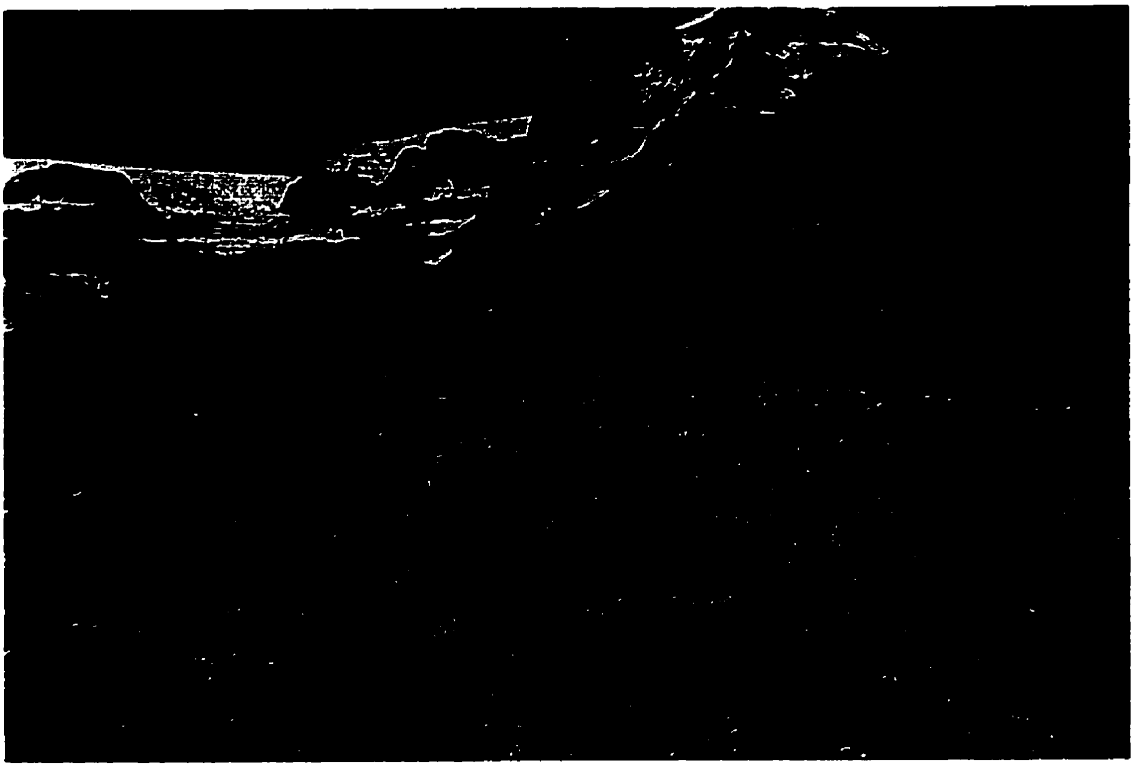


Figure 8.12 The washed till area on the south side of Stutfield Valley (foreground). The ridge in the background is the large moraine in the centre of the valley (Area 1). Note the difference in vegetation between the washed till area and area 1 (Photo BJR, July 28, 1996).

washed till on its southern side, it suggests that the modified till and area l were created simultaneously. As the glacier readvanced to its early 19th century position, the meltwater produced modified the existing till surface.

Table 8.7

Limiting Ages for Landforms on the Southern Valley Floor

Location	Core ID	Inner Ring Date	Surface Date
Unmodified till	**Y9620**	**1809**	**1779**
	Y9604	**1814**	**1779**
	Y9617	**1809**	**1779**
Washed till	**T9699**	**1841**	**1809**
	Y9628	1851	1819
Area l	**Y9603**	**1853**	**1818**
	T9693	1859	1829

Note: Sample ID's include year identification. Surface dates include an ecesis estimate of 30 yrs. and a pith correction. Bold indicates the limiting date.

8.3.2.3 Stutfield South Summary

The dates obtained on the south side of the valley indicate at least three periods of moraine construction: 1) the early to mid 18th century, 2) the late 18th century and 3) the early 19th century. The earliest phase of glacial activity created the trimline and the outermost moraine along the valley wall (ridge j). Minimum surface ages from ridge j indicate that these features were created prior to 1743. A subsequent advance in the late 18th century resulted in the deposition of till on the valley floor. However, the evidence suggests that this area was subsequently modified by meltwater as the glacier readvanced to its final dated position in the early 19th century, creating the main lateral moraine (ridge i) and the large moraine in the centre of the valley (area l).

8.3.3 Evaluation of the Site Chronology

Results from Stutfield Glacier indicate at least five distinct periods of glacial advance: 1) Advance A in the late 13th century, 2) Advance B prior to 1743, 3) Advance C from 1758-1767, 4) Advance D before 1802-1820 and 5) Advance E during the mid 19th century. Advances C and D are supported by several lines of evidence from various locations throughout the valley and all are summarised in Table 8.8.

Table 8.8

Summary of Data at Stutfield Glacier

Location	Surface Date
Advance A	
Tree embedded in north lateral moraine	after 1271
Advance B	
Moraine j	<1743
Advance C	
North lateral moraine	1758*
North trimline	<1767
Island f	<1777
Moraine c	<1778
Till blanket south of the river	<1779
Advance D	
North lateral moraine	1812-1820*
Moraine d	<1802
Island g	<1803
Moraine I	<1807
Area l on south valley floor	<1818
Advance E	
Moraines upvalley of Advance D deposits	undated

* - indicates precisely dated events

Evidence for Advance A was obtained from several pieces of wood buried in the northern lateral moraine. This is consistent with the advance reported by Osborn (1993, 1996) from trees located in the north and south lateral moraine. The next recorded advance, Advance B, created the outermost moraine (ridge j) and trimline along the southern valley wall. Minimum surface ages from a variety of locations along the entire feature support an advance prior to 1743. Unfortunately, there is no evidence of a corresponding event on the north side of the valley.

The furthest downvalley limit of glacial advance preserved in the Stutfield Valley is attributed to Advance C in the mid 18th century. The trimline, outermost moraine and till island on the north side of the valley and the till blanket on the south side of the valley all suggest an advance prior to 1767. This age is supported by damage dates of 1758 from two tilted trees along the northern trimline as well a minimum surface age along the northern "ledge". Deposits for this event are the most extensive on both sides of the valley, consistent with the hypothesis that this was the most extensive advance. Furthermore, the presence of these deposits across almost the entire width of the valley suggests that the glacier was a single lobe during this period.

The similarity in the limiting ages for Advances B and C may suggest that the landforms attributed to these events were formed during the same event prior to 1743. If this were the case, it would mean that the ecesis interval for several locations in the valley was variable and considerably longer, ca. 40-50 yrs., than the estimates calculated at Columbia Glacier in this study. Even though the dates obtained from a few of the sites may be ambiguous and could be attributed to either event, inspection of the type of evidence used clarifies there were at least two events. The two strongest pieces of evidence for these advances are a minimum surface of 1743 and a precise date of 1758. The 1743 date is based on a tree growing on the outermost moraine on the southern valley wall yielded a pith date of 1773. For this tree to be growing here at this date, the ice must have been removed by 1773 at the latest. This assumes that the area was colonised immediately after the ice retreated from this position, which is highly unlikely. Although there is a forest on the slope above, it is not continuously forested. Furthermore, the blocky substrate, lack of moisture, exposure to katabatic winds, proximity to the glacier and high elevation create plausible arguments for an ecesis interval longer than the 20-30 yrs. applied throughout this study. Therefore, the 1743 date obtained from this site provides an absolute minimum age for this location. However, the youngest evidence attributed to Advance C is a precise date of 1758 based on a tilted tree located along the northern trimline. This indicates that the glacier was definitely advancing during this period, and the damage to this tree cannot be attributed to an older advance. Since the oldest tree sampled on the south side of the valley indicates that the surface was clearly created prior to 1743, there must have been at least two events in the 18th century. Although the trimline and outer moraine on the north side of the valley and the till on the southern valley floor may have been created during either event, they were more likely formed during the latter event. The minimum surface age calculated is very similar to the precise date obtained along the valley wall and there is no reason to suspect that the ecesis intervals at these sites would be longer than the 20-30 yr. average. They are all located near large seed sources in the bottom of the valley, with ready access to moisture.

The last dateable period of moraine building was Advance D, with closely limiting dates in the early 19th century. Similar to Advance C, evidence for this advance can be found at a variety of locations throughout the valley. Unlike the evidence for Advance C however, the deposits from this event have not been significantly eroded and a tentative ice front position can be estimated. The most precise dating control for this event was an 1812 scar date obtained from a damaged tree along the north lateral moraine. However, minimum surface ages for the inner moraine on the southern valley wall (ridge i), the large moraine in the centre of the valley (area i) and the middle moraine on the north side of the valley (ridge d), all correspond to an advance prior to 1802. From the position of the relevant features: ridge d, island g, and area i connected to ridge i on the southern valley wall by ridge n (Figure 8.3), it appears that the ice front consisted of two lobes during this early 19th century advance.

Although there are several distinct moraine ridges upvalley of the early 19th century deposits (e.g. ridge e on the north side of the valley), they lack dateable material. It is likely that these deposits were created by an advance in the mid 19th century. The lack of dateable surfaces for a later event may be attributed to the nature of the ice front. The morphology of the glacier and the preponderance of fluvially modified till and abandoned channels suggest that the glacier is a relatively stagnant lobe producing a large amount of meltwater. Therefore, it is possible that many of the landforms created by the most recent advance have been destroyed by fluvial erosion or submerged by the proglacial lake.

8.4 Documented Ice Front Positions

At its LIA maximum in the late 18th century, Stutfield Glacier extended approximately 1.5 km downvalley of its present ice front, creating the trimline and the till island on the north side of the valley. As no other historical records of the position of Stutfield Glacier have been located, the first documented ice front position is from the 1948 aerial photograph. Throughout its documented history, Stutfield Glacier has been retreating continuously (Table 8.9). In the 1948 photograph, the glacier had experienced a net retreat of approximately 750 m from its mid 18th century maximum position (Figure

8.13). Between 1948 and 1955, the glacier retreated at a rate of 37 m/yr., then fell to 11 m/yr. until 1968. Following this, retreat increased close to the earlier rate of 37 m/yr. until 1976. It then dropped again to 12 m/yr. and retreated ca. 220 m between 1974 and 1992. The heavily debris covered nature of the glacier makes it difficult to distinguish the exact position of the commonly irregular ice front. In addition, it causes it to lose a considerable amount of mass through downwasting rather than retreat. These two factors may account for the irregular pattern of recession at Stutfield Glacier.

Table 8.9

Recession of Stutfield Glacier

Period	Distance of Retreat (m)	Rate of Retreat (m/yr.)
ca. 1760 -1948	750	-
1948-1955	257	37
1955-1968	144	11
1968-1974	188	32
1974-1992	220	12

8.5 Summary

Evidence from the Stutfield Glacier forefield indicates at least four periods of glacial advance (Table 8.10), most with corroborating dates throughout the valley. Outer ring dates between 1064 and 1271 AD for three logs found in the northern lateral moraine suggest an advance during the late 13th century, consistent with those reported at other sites in the Canadian Rockies (e.g. Luckman 1995, 1996c). Although the trimline on the southern valley wall was created by an advance in the early 18th century, the maximum downvalley limit of ice was not achieved until the mid 18th century. Damaged trees on the north lateral moraine indicate that part of the glacier was advancing in 1758, consistent with minimum surface ages throughout the forefield that suggest that the glacier began retreating prior to 1767.

The last major advance with dateable deposits occurred at the beginning of the 19th century and is delimited by a nearly continuous series of moraines. The most precise dating for this event placed it between 1812 and 1820 and was obtained from tilted and

0 500m
Scale is variable based on lower outwash

A - 1992 C - 1974 E - 1968 G - 1948
B - 1977 D - 1970 F - 1955

Figure 8.13 Recent Ice Front Positions at Stutfield Glacier (NAPL A31609-48).

scarred trees along the north lateral moraine. However, minimum surface ages from locations downvalley suggest that the ice had been removed by at least 1802. The apparently conflicting dates obtained in the different parts of the valley may indicate that this advance was not entirely synchronous throughout the valley or that the ecesis interval varied through the valley.

The prominent moraines upvalley of the early 19^{th} century deposits are undated but believed to have been formed during a mid 19^{th} century advance that has been widely documented at many other sites throughout the Canadian Rockies (e.g. Luckman 1988, Smith *et al.* 1995). Over the last fifty years and likely throughout its entire history, proglacial lakes, meltwater channels and dead ice topography have been the dominant features at this glacier forefield. Associated processes would likely erode or obscure any moraines produced during subsequent advances, accounting for the lack of post mid 19^{th} century deposits. In addition, the glacier appears to have been a relatively stagnant lobe throughout its documented history, resulting in most of the debris being deposited as melt-out till rather than distinct ridges.

Table 8.10

Major Periods of Glacial Advance at Stutfield Glacier

Date of Advance	Supporting evidence
after 1271	Outer date of overridden tree in north lateral moraine
prior to 1743	Oldest tree on outermost lateral moraine
1758 - 1767	Oldest tree below trimline, outermost moraine on north valley wall, south valley floor, tilting event on northern lateral moraine, oldest tree on north lateral moraine
prior to 1802-1820	Oldest tree on valley floor moraine, north and south lateral moraine, middle moraine on north valley wall, scar and tilting event on northern lateral moraine
ca. mid 1800's	Undated moraines upvalley of the early 19^{th} century event

CHAPTER IX
SASKATCHEWAN GLACIER

9.1 Introduction

The Saskatchewan Glacier is located near the southern end of the Columbia Icefield, between Mount Andromeda to the north and Castleguard Mountain on the south. It is the longest glacier of the Icefield, with a total length of almost 10 km (Robinson 1996). However, less than half of its total area of 28 km^2 (based on 1992 aerial photography) is below the 2500 m.a.s.l. 1992 snowline (Robinson 1996). It is also the only major glacier from the Icefield without a significant icefall; it gradually descends from the Icefield Plateau at 2600 m.a.s.l. to a terminus at approximately 1900 m.a.s.l. The overall surface slope is 5° and it has a thin, gently inclined snout (Figure 9.1). The Saskatchewan Glacier is a geometrically simple ice stream that is approximately 1400 m wide over most of its length but tapers to less than 500 m near the terminus. In addition to its main Icefield source, the glacier is fed by a cirque glacier located on the south face of Mt. Andromeda that joins the main glacier ca. 5.5 km upvalley of the snout. The conspicuous medial moraine on the north side of the Saskatchewan Glacier is formed by the junction of these two glaciers. Meltwater flows from the terminus into a proglacial lake of 0.25 km^2 (based on 1992 aerial photography) that is first seen on the 1966 aerial photographs. Since its appearance, the lake has expanded headwards. The only outlet of this lake is the North Saskatchewan River at its eastern limit.

The south side of the valley is composed of bedrock cliffs and the only evidence of the ice margin against this slope is a trimline across the talus slopes at the base of the cliffs (Figure 9.2). The slopes on the northern valley side are less steep, covered with a thin veneer of till in some areas, but lack significant ice marginal landforms such as moraines or kames. In places, a well defined trimline separates the bare, lower slopes from patches of forest. Approximately 1 km downvalley of the present ice front, the forest cover on the valley side becomes more continuous, with a well developed trimline at its lower edge (Figures 9.3 and 9.4). The glacial trimline extends downvalley for ca. 1 km where it is replaced by a fluvially-trimmed bluff at the limit of the forefield (Figure 9.5). The trimline

Figure 9.1 Saskatchewan Glacier from the northern valley wall (Photo BJR, June 26, 1997).

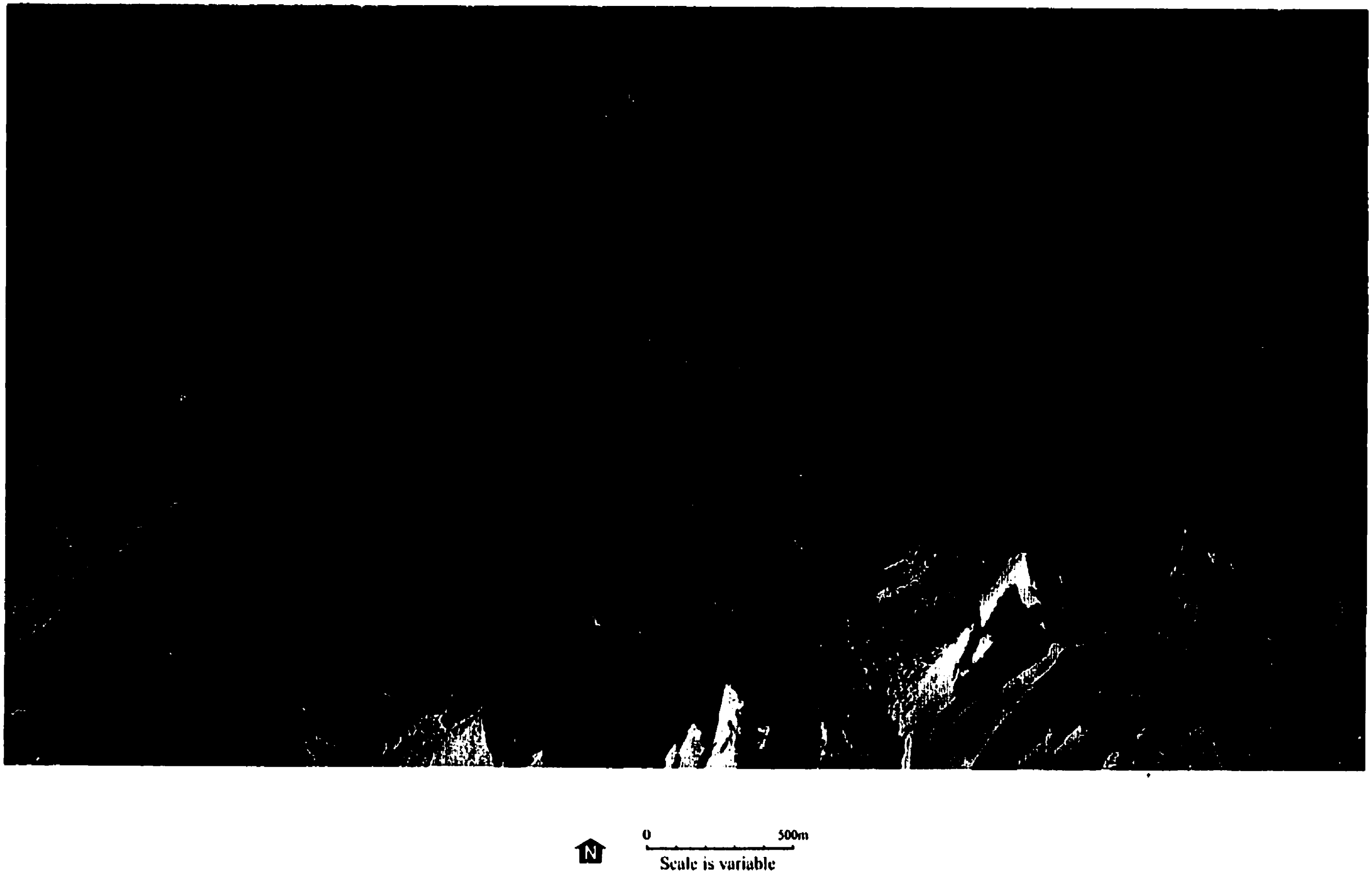


Figure 9.2 Aerial Photo of the Saskatchewan Glacier site (Province of British Columbia, Photo Number BC7876-197, Sept. 1976).

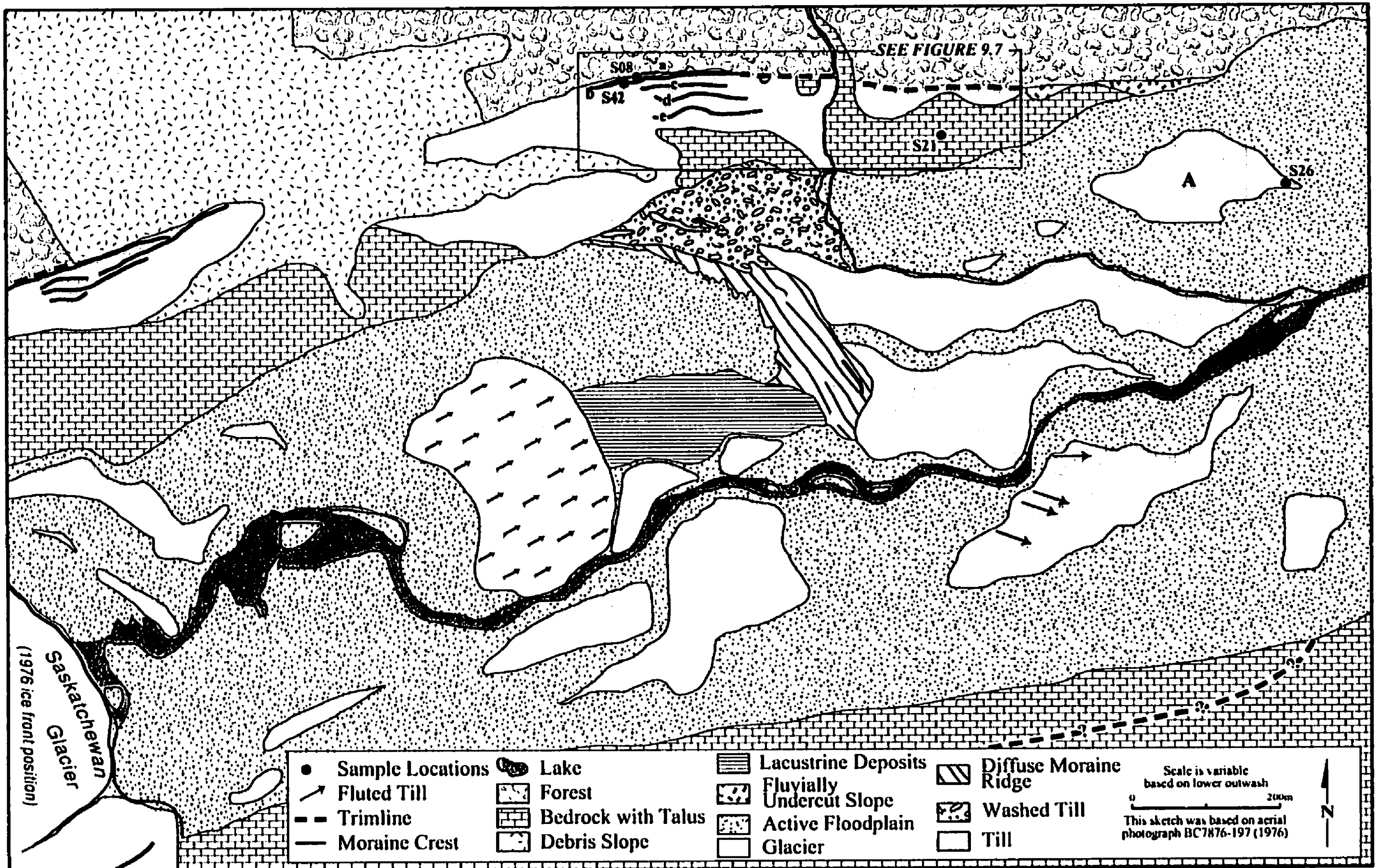


Figure 9.3 Saskatchewan Glacier forefield with samples discussed in text. Year identification has been excluded from sample numbers. (e.g. S08=S9608). The proglacial lake has expanded upvalley between 1976 and 1996.

Figure 9.4 The lower North Saskatchewan Glacier Valley (July 5, 1996). Note the trimline along the northern valley wall in the distance (Photo BHL).

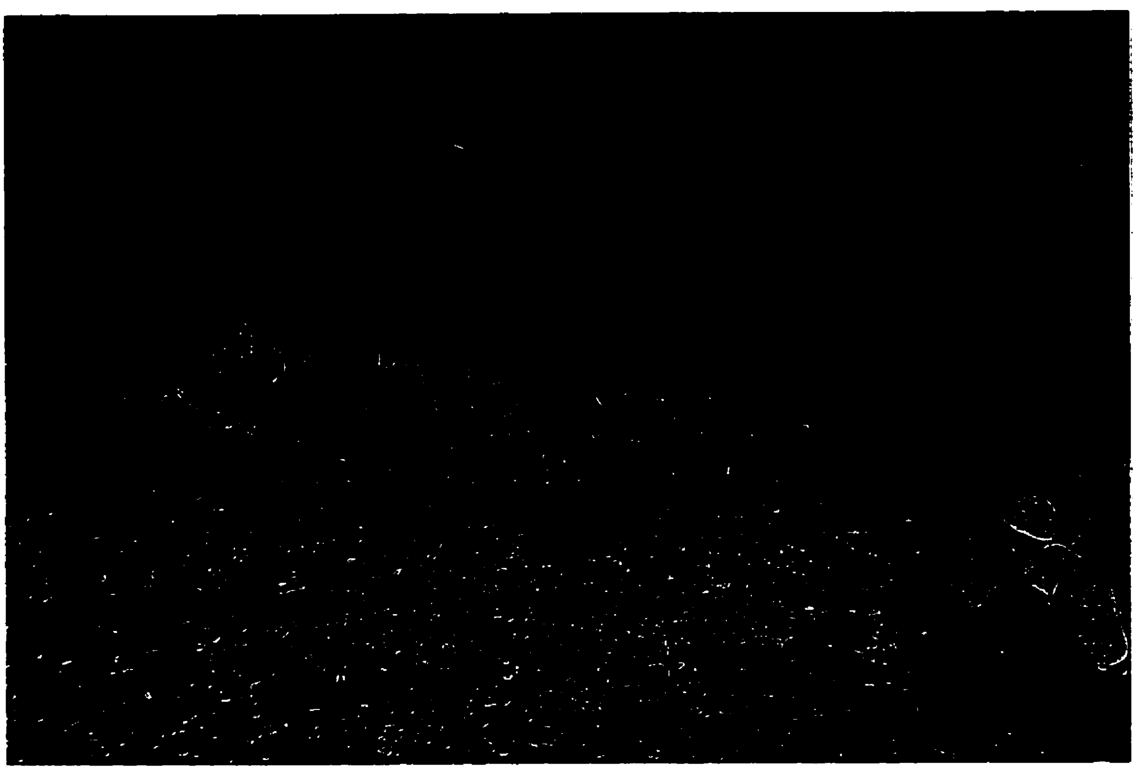


Figure 9.5 Fluvially eroded bluff at the downvalley limit of the trimline. (Photo BJR, July 8, 1996).

can be divided into two major sections on either side of a small tributary creek from the northern valley wall (Figure 9.3). Upvalley of the creek, the trimline consists of a veneer of till, with five small moraine ridges, draped over a bedrock slope (Figure 9.6). Downvalley, the slopes below the trimline are predominantly bedrock, with a small ridge of till near the trimline.

There are no well developed lateral or terminal moraines to indicate the maximum LIA position, but the downvalley limit of the trimline suggests that the glacier extended almost 2 km downvalley from its 1992 position. On the valley floor, several large areas of unmodified, often fluted till have been dissected by active and abandoned outwash channels (Figure 9.3). McPherson & Gardner (1969) concluded that these "topographic highs" were not end moraines but emerged as contiguous surfaces from beneath the ice. Although these surfaces are mainly treeless, they do provide a minimum estimate of the downvalley LIA extent of the Saskatchewan Glacier. Several hundred metres downvalley of the limit of the forefield on the south side of the valley, there is a conspicuous forested terrace about 4 m above the main outwash (Figure 9.2).

9.2 Previous Investigations

As a result of its easy access, the Saskatchewan Glacier has been the subject of several glaciological and geomorphological investigations (e.g. Meier *et al.* 1954, McPherson & Gardner 1969). The nearly continuous coverage afforded by these observations has enabled researchers to reconstruct accurate ice front positions for the 20th century that are discussed in detail in section 9.4. Glacial events prior to these direct observations have also been investigated by several researchers; namely Field (1949), Heusser (1956) and Luckman *et al.* (1993a, 1994).

Most previous work concerning the LIA extent of the Saskatchewan Glacier has been done by W.O. Field and C.J. Heusser (Field 1949, Field & Heusser 1954, Heusser 1956). Heusser (1956) reports two trees near the limit of the trimline on the north side of the valley which record ice thrust events in 1807 and 1813. Cores from trees growing on the slope below the trimline indicate that the glacier had receded from this area by 1854. However, retreat from the northern side of the valley was not uniform. Upvalley and

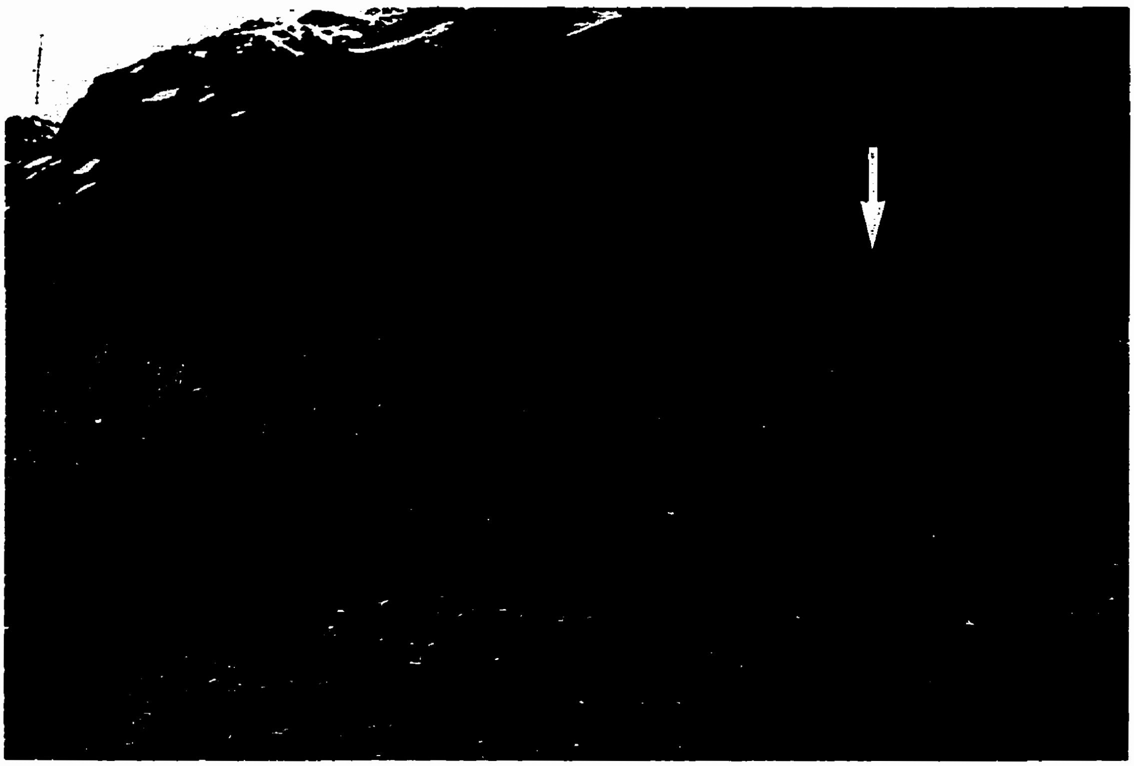


Figure 9.6 Well developed moraine crest and trimline upvalley of the creek on the north side of the valley. Note the moraine ridges and the detrital wood near the top of the slope (Photo BJR, June 26, 1997).

"beyond" the two tilted trees, Field (1949) reported a tree that was pushed over during an ice-thrust event in 1893. As there were living trees that predated this event on slopes downvalley, Heusser (1956) believed that this was a localised advance. However, this advance extended sufficiently far downvalley to create a small moraine just inside the outermost stand of the glacier as defined by the trimline. The ages of trees growing on this moraine indicate that it was formed in approximately 1899 (Heusser 1956).

Luckman *et al.* (1993a, 1994) carried out a reconnaissance study of the downvalley limit of the trimline by sampling snags in this area. However, as these samples were taken from above the fluvially eroded bluff, the dates shown in Table 9.1 may reflect slope instability rather than glacial events.

Table 9.1

Radiocarbon Dates from Material Sampled at Saskatchewan Glacier (Luckman *et al.* 1993a, 1994).

Location	Sample ID	Date
Surface trees at the downvalley limit of the trimline*	SG9311	1647-1732
	SG9310	1701-1835
	SG9313	1836-1905
	SG9312	1885-1970
Buried trees by the glacier snout	BGS 1369	3180 ± 80
	Beta 65384	3170 ± 60
	Beta 29957	2940 ± 90
	Beta 31358	2880 ± 70
	Beta 65383	2540 ± 60
Buried trees excavated from the	Beta 33013	2170 ± 60
southern terrace	Beta 28348	1590 ± 70

* - all dates are calendar dates

Evidence of pre-LIA events have also been reported by Luckman *et al.* (1993a, 1994). Detrital wood recovered near the 1988 ice front revealed outer ring dates between 3200 and 2500 yrs. BP (Table 9.1). Although all of the samples were washed out of the glacier and had been derived from an unknown source upvalley within the ice, they provide strong circumstantial evidence for an advance equivalent to the Peyto Advance (3300-2800 yrs. BP).

Downvalley of the estimated LIA maximum ice extent, the river is actively eroding a 4 m high aggradational fill on the south side of the valley (Figure 9.2). The stratigraphy of this feature suggests that it was built up as part of an active outwash that probably filled the entire valley. Logs recovered from 2.5 m below the surface of this terrace reveal radiocarbon dates of 1590 ± 70 and 2170 ± 60 yrs. BP (Luckman *et al.* 1993a). Twenty four trees growing on its surface were sampled and revealed a minimum surface age of 400 yrs. These dates provide very broad limits for the creation of the terrace, suggesting that the major filling event postdates 1600 yrs. BP and the terrace was completely formed prior to 1600 AD and is therefore not a LIA maximum feature. Unfortunately, the data available are insufficient to determine precisely when the stream began downcutting into the terrace. However, it is believed that the glacier would have probably produced a sufficient amount of meltwater to achieve this as it advanced to its LIA maximum.

9.3 Results of the Present Study

The lack of vegetation and absence of well developed lateral and terminal moraines in the forefield of the Saskatchewan Glacier prohibit the use of living trees to establish useful minimum surface ages. Therefore, trees killed by the glacier along the northern trimline were the primary source of information concerning fluctuations of this glacier. Based on the information they provide, the samples obtained at this site can be divided into three categories; 1) *in-situ* trees killed by the glacier that provide evidence of the time a glacier was in a particular location, 2) detrital wood which indicates the timing of a glacier advance and 3) a limited number of living trees which provide minimum surface ages. Complete results from all samples collected at this site are presented in Appendix VI.

Near the limit of the forefield on the north side of the valley, there is a 1 km long, well defined trimline that is bisected by a small tributary creek (Figure 9.3). The trimline upvalley of the creek is characterised by a veneer of till draped over a bedrock slope. Superimposed on this till surface are five small moraine ridges (2 m high). The three innermost moraines (ridges c-e in Figure 9.3) are nearly imperceptible ridges about 120 m long with marginal vegetation cover at their upvalley ends. The outermost moraine (ridge

a) is located just below the trimline at the top of the slope with ridge b immediately inside it. Upvalley, ridge a is approximately 1-1.5 m higher than ridge b, although at their downvalley limits, the two are equivalent heights and ridge b appears to be "pushed up" into a sharp crest against ridge a, making them nearly indistinguishable as individual ridges in Figures 9.3 and 9.7. While there are several trees growing along this 300 m stretch of moraine, there is also a significant amount of wood on their surfaces (Figure 9.8).

Most of the outer moraines are blanketed with an assemblage of dead wood composed of shattered and smashed wood fragments, entire trees (including the roots), root stocks, detrital logs and *in-situ* stumps. Most of this material displays evidence of glacial shear and is tangled together as a single unit mainly on the outer moraines. Although the assemblage is less dense along the valley wall downvalley of the moraine ridges, numerous sheared, *in-situ* stumps and detrital logs can be found here over a vertical range of about 30 m, from the trimline to ridge c. Within this assemblage, the preservation of the wood ranges from intact trees to others that have completely rotted and are indistinguishable as individual trees. Although the more rotted pieces of wood tended to be located lower on the slope, the most severely rotted wood was located at the bottom of the "pile" along moraine b. So much wood had been deposited on this ridge that its crest was no longer evident. This large volume of wood likely resulted in a moist microenvironment, facilitating decomposition. Furthermore, the weight of the overlying wood likely crushed many of the lower trees.

Downvalley of the moraines, the trimline continues for 250 m over two bedrock outcrops to the creek. Between the two bedrock outcrops, there is a small, sheltered "embayment" (Figure 9.9). In this "embayment" there are two moraine "ledges", less than 0.5 m wide and 1 m apart, parallel to the trimline. The area downvalley of the creek is an almost exclusively bedrock slope capped with a small ledge of till less than 50 m wide immediately below the trimline. This trimline continues downvalley for about 225 m where it grades into a slope controlled by fluvial erosion at its base. On the valley floor, there are several distinct till islands between 10 and 15 m high. These features have subdued surface topographies, consistent with McPherson & Gardner's (1969) hypothesis that they are former glacial landforms that were overridden by the glacier during its most

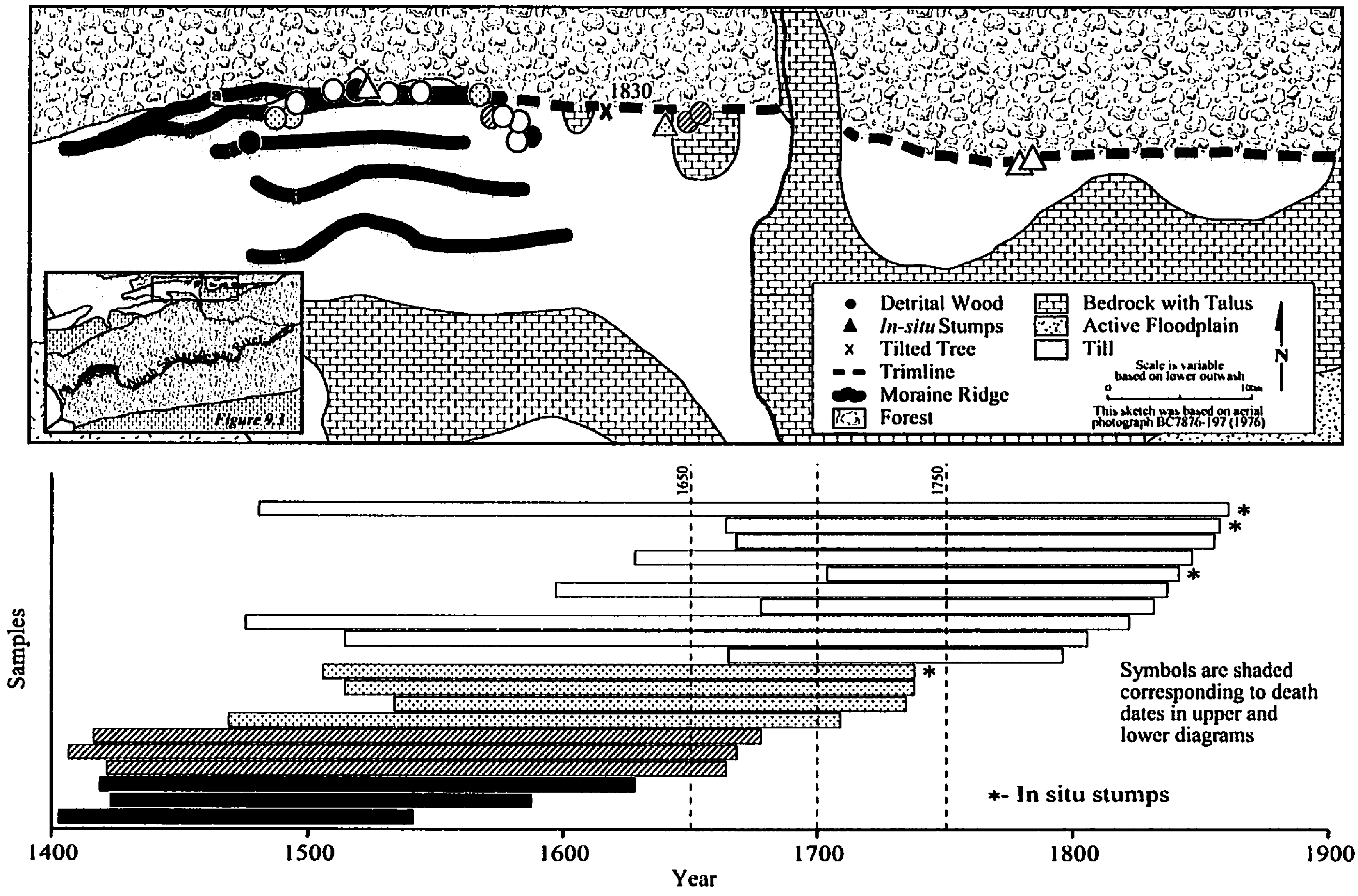


Figure 9.7 Location and dates of Saskatchewan Glacier snags.

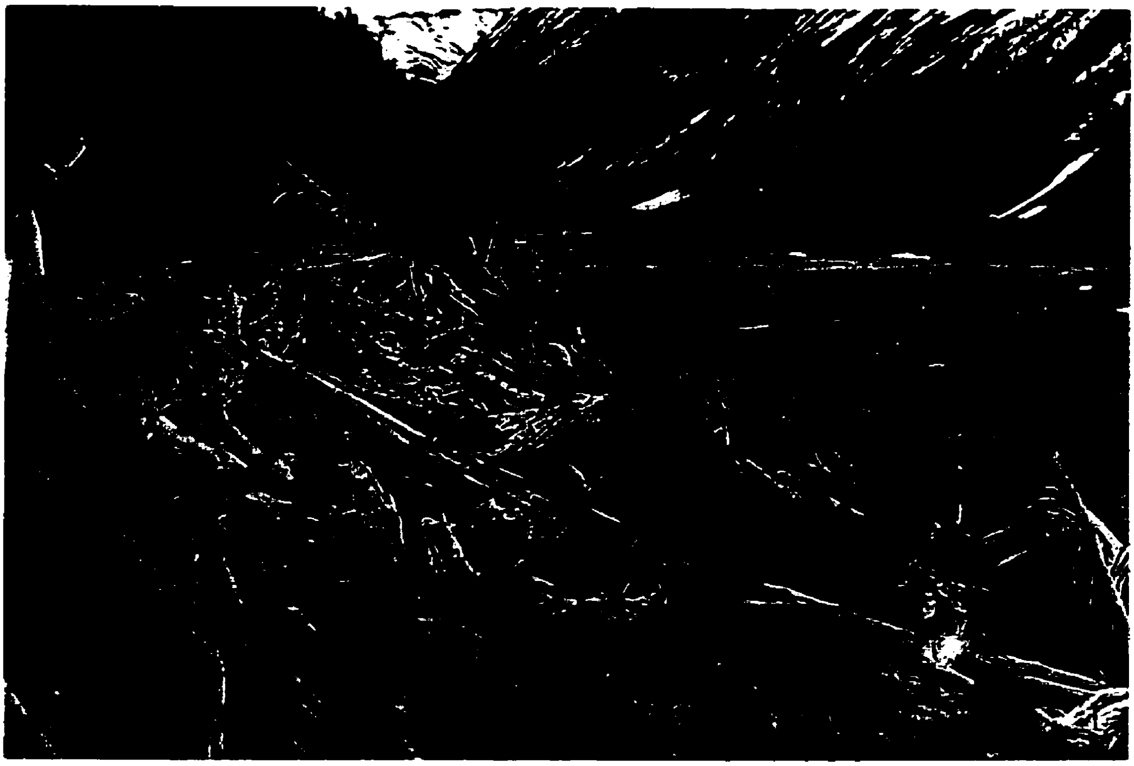


Figure 9.8 Detrital wood assemblage on the outer moraines at Saskatchewan Glacier (Photo BHL, June 19, 1996).

Figure 9.9 Embayment between the two bedrock outcrops at Saskatchewan Glacier, looking downvalley. The tilted tree in the foreground is Tree S9665. The one in the background that is partially obscured was rotten (Photo BJR, June 26, 1997).

recent advance and subsequently exposed as continuous surfaces when the glacier retreated.

9.3.1 *In-situ* Trees

Although there was a considerable amount of dead material along the northern trimline, most of the *in-situ* stumps were located on the slopes below. Unfortunately, all of these stumps were too shattered and rotten to section. However, five *in-situ* trees were found at other locations along the trimline. There were very few rooted stumps in the tangle of wood along the outer moraines and only one was sound enough to yield a recoverable cross section (S9651, Figure 9.7). This 2.5 m tall tree was found tilted about 25° from the vertical at the base of the distal slope of moraine a. The roots on the moraine proximal side had been exposed and sheared, while the distal side of the tree appeared to be in growth position. This configuration suggests that the tree was partially uplifted and tilted from below as the glacier advanced up the valley side. The outer surface of S9651 was not substantially eroded and its outermost preserved ring suggests that the tree was killed shortly after 1862. No reaction wood series were found in either section taken, and it therefore appears that the tree was killed during the uprooting event, providing a maximum limiting date for the presence of the ice.

Between the two bedrock outcrops near the creek, three killed, *in-situ* trees were found (Figure 9.7). These trees were tilted upslope with their trunks nearly parallel to the slope and almost completely severed from their roots which were rotting *in-situ* (Figures 9.9 and 9.10). Unfortunately, only one of the trees could be sampled since the other two were completely rotten. The lower part of S9665 was partially rotten so two wedges were taken; one at the base of the tree and the other approximately 1.5 m up the trunk. Reaction wood was not identified in either wedge, but the extreme angle of this tree suggests that it was killed by tilting that partially severed the tree from its roots. The upper wedge was better preserved than the lower, with an outer ring indicating that the tree was killed shortly after 1740. This provides the first evidence for an ice advance during the 1700's.

Figure 9.10 Tree S9665 (Photo BJR, July 6, 1996).

In addition to these killed trees, a previously cut stump (S9655) was found along the trimline between the bedrock outcrops. The location of this tree suggests that it may have been the tree recording Field's (1949) 1893 event. Although the surface of this stump had been substantially weathered and may have resulted in the loss of a few internal rings, a ring count revealed that the tree was at least 175 yrs. old when cut and contained a reaction wood series near the beginning of the record. Therefore, this tree cannot represent Field's (1949) late 19th century advance. However, it may be one of the ice-thrust trees sampled by Heusser (1956) in 1953. Using an outer ring date of 1953, it was determined that the major tilting event began in 1830, although minor reaction wood series beginning in 1796 and 1807 were also present. Unfortunately, these dates could not be verified through crossdating as the presence of reaction wood and loss of a few internal rings rendered all crossdating attempts unsuccessful. The variation in dates obtained in Heusser's (1956) study and the current investigation for what appears to be the same tree results in uncertainty concerning the correct interpretation. Although there is greater confidence in the ring count from the current investigation, the precise locations of the two trees sampled by Heusser (1956) are not clearly known. Therefore, this tree may not be either one reported by Heusser (1956).

The final two *in-situ* trees were sampled approximately 150 m downvalley of the creek at the trimline (Figure 9.7). Both were tilted into the forest, although only one (S9668) had been broken from its roots, which were rotting *in-situ*. The outer ring dates from both trees indicate that they were killed shortly after 1842 and 1858 (S9667, S9668 respectively), providing a mid-19th century limiting age for the trimline east of the river. Furthermore, these dates are roughly equivalent to the 1862 death date obtained from S9651 upvalley.

9.3.2 Detrital Wood

There was a large amount of detrital wood along the outer moraines, but sampling was limited to the larger specimens that appeared to be the best preserved. Most of the twenty-two cross sections taken were from the outer moraines, although two were located along the trimline downvalley. The locations and corresponding periods of record for the fourteen samples that were successfully crossdated are shown in Figure 9.7. All the sections obtained were well preserved, suggesting that few outer rings had been lost through erosion. Therefore, the outer ring dates of these samples provide closely limiting kill dates for these trees. The age distribution shown in lower section of Figure 9.7 indicates that these snags can be divided into four groups based on their death dates: those killed 1) prior to 1650, 2) between 1650 and 1700, 3) between 1700 and 1750 and 4) post 1750.

Conventionally, one might expect to find samples reflecting the oldest event to be found on the outermost features. However, the only samples that conform to this are the two samples on the bedrock outcrop nearest the creek, indicating a glacial advance in the late 17th century. In conjunction with another detrital log in the tangle of wood along the moraines, these trees provide evidence for the earliest glacial advance at this site, during a period generally lacking dated deposits at other sites in the Canadian Rockies (see Figure 3.1). Within the assemblage on the outer moraines, there is no apparent relationship between the location of the samples and their death dates, only distinct periods of increased mortality. However, if the glacier was advancing continuously into forest at this site, an even distribution of dates over time would be expected, as is observed for the period prior to 1650. The fact that the dates for most of the detrital logs appear to be grouped into three periods with closely spaced death dates suggests at least three distinct pulses of ice advance on this slope: between 1) 1664 and 1678 , 2) 1737 and 1740, and 3) between 1824 to 1856. The absence of a distinct pattern in the death dates prior to 1650 suggests that they were killed during a prolonged advance culminating in the late 17th century. The dates obtained from the *in-situ* stumps are consistent with the age ranges for the latter two events. Three *in-situ* trees located on the trimline had death dates ranging from 1842 to 1862, while the tilted tree found in the embayment between the two bedrock outcrops suggested a tilting event in approximately 1830. This evidence supports a glacier advance during this period. Likewise, a tree killed by the glacier in 1740 provides additional limiting support for an advance in the mid 18th century. An assemblage of mixed age in such a small area is consistent with the notion of a glacier advancing over the same area. Initially, the glacier advanced into the forest and then retreated. As it subsequently readvances, it reworks trees killed during previous advances and kills more trees as it extends further into the forest. When the glacier retreats from its maximum position, the trees killed during its former advances are all deposited in the same area.

9.3.3 Minimum Surface Ages

Several attempts were made to determine the minimum surface age of features in the Saskatchewan forefield (Table 9.2). The four areas with sufficient vegetation to enable this type of sampling were 1) the moraine island A on the valley floor, 2) the bedrock slope east of the creek, 3) moraine a and 4) moraine b (Figure 9.3).

Table 9.2

Limiting Ages for Surfaces on the North Side of the Saskatchewan Valley

Location	Core ID	Inner Ring Date	Surface Date
Moraine island A	**S9626**[b]	**1947**	**1907**
Bedrock outcrop	**S9621**[b]	**1912**	**1877**
Moraine a	**S9608**[b]	**1929**	**1894**
Moraine b	**S9642**[b]	**1935**	**1895**

Note: Sample ID's include year identification. Surface date includes an ecesis of 30 yrs. and a pith correction. Bold indicates the limiting date.

On the furthest downvalley moraine island on the valley floor (island A), the oldest tree sampled revealed an off pith date of 1947 (S9626), corresponding to a surface age of 1907 (Figure 9.3). Earlier ages were obtained for the features along the side of the valley. A near pith date from a 1.5 m spruce located on the predominantly bedrock slope east of the creek suggests that this surface became exposed prior to 1877 (S9621, Figure 9.3). This value is roughly consistent with dates obtained from two trees located on the outermost moraines west of the creek and does not contradict Heusser's (1956) surface age of 1854. The oldest tree sampled on moraine a (S9608), indicates that the moraine was deposited before 1894. An equivalent age of 1895 (S9642) was determined for the adjacent ridge (moraine b) from a tree located on the distal slope of the moraine (Figure 9.3). Compared to other sites investigated in this and other studies, these appear to be young surface ages for the outermost glacial features. However, inspection of photographs from the early 20th century clearly shows that the ice front remained close to its estimated maximum position into the early 1900's. In addition, these photographs also show that the glacier thinned considerably prior to retreating upvalley. Therefore, the lateral positions would be expected to have earlier surface ages as they were ice free earlier and in close proximity to an abundant seed source (the forested valley side).

9.3.4 Event Summary

In evaluating the kill dates obtained from both the *in-situ* trees and detrital wood, there is no apparent relationship between the location of the samples and their death dates. However, the similar death dates indicate three periods of increased mortality, between 1) 1664 and 1678 (3 trees), 2) 1737 to 1740 (3 trees) and 3) 1824 to 1862 (9 trees). These reflect periods when the glacier was advancing into the forest and killing trees. The radial preservation of these samples ensures that few rings were lost through erosion, suggesting that the tight grouping of the death dates reflect distinct glacial events. As the glacier advanced, it killed trees along the valley side, leaving the root stocks and transporting the trunks upslope. Trees killed during each subsequent readvance were mixed with the detrital material reworked from former advances. The resulting assemblage was a mixture of dead wood with a variety of death dates.

The largest grouping of dates are those from the mid 19th century, including three trees killed *in-situ* along the trimline. In addition, the only tilted tree sampled in this study with reaction wood, suggests a tilting event in approximately 1830. This is consistent with the death dates of the trees killed along the margin of the forefield. This suggests that the Saskatchewan Glacier advanced further into the forest during the final event than during previous advances and killed more trees. Furthermore, the *in-situ* trees mark the outer limit of this glacial advance. Therefore, it is highly likely that the Saskatchewan Glacier achieved its maximum position in the late 1850's. This is consistent with a 19th century downvalley maximum suggested by Field (1949) and Heusser (1956) and the minimum surface ages determined in this study. However, Heusser (1956) also reported additional ice thrusting events along the trimline in 1807, 1813 and 1893. The tilted tree sampled in the course of this study casts some doubt concerning Heusser's (1956) 1813 event. Since the position of Heusser's samples are unknown, the discrepancy between his dates and those determined in this study could not be clarified. However, it is possible that these dates represent localised ice-thrusting events at different positions along the glacier margin and not major advances.

9.4 Documented Ice Front Positions

Since its discovery in 1896 by Walter Wilcox (Wilcox 1900), ice front positions of the Saskatchewan Glacier have been well documented as a result of its relative accessibility. The primary source for early 20th century ice front positions has been terrestrial photography, usually from Parker's Ridge (e.g. Field 1949, Field & Heusser 1954, McPherson & Gardner 1969). Ice front positions from 1945 to 1980, were recorded at annual and biennial intervals through glacier surveys completed by the Water Survey of Canada (e.g. WSC 1982). These sources have enabled researchers to calculate the rate of retreat for the Saskatchewan Glacier over the entire 20th century (Figure 9.11). Complete results from these sources are presented in Appendix VII.

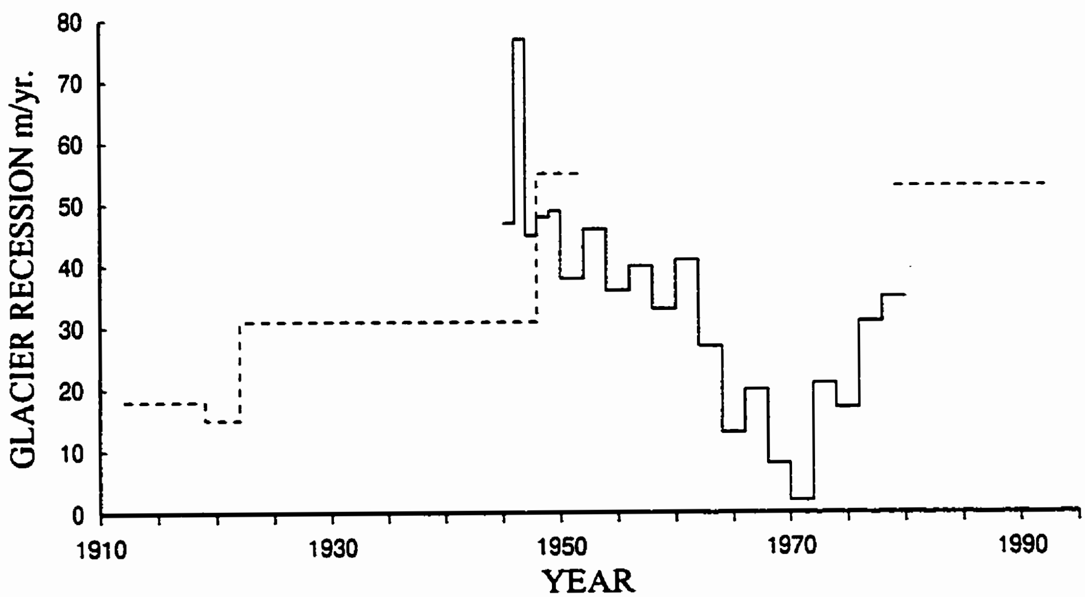


Figure 9.11 Frontal recession of the Saskatchewan Glacier (1912-1992). Solid line based on annual or biennial measurements by the Water Survey of Canada. Dotted line based on ice front positions reconstructed by Field (1949, 1954) and this study based on photographs.

Photographs from 1912 indicate that the Saskatchewan Glacier had receded approximately 170 m from its maximum position. Based on a series of intervening photographs, Field (1949, Field & Heusser 1954) calculated a net retreat of almost 1 km between 1912 and 1948. Measurements by the Water Survey of Canada found that the retreat of Saskatchewan Glacier peaked in the 1946-47 measurement year (Figure 9.11). Compared to the reasonably consistent recession observed between the 1930's and 1950's, these values appear anomalously high. As this represents only the second measurement year, one should consider the possibility that this large value may be a result of an error in measurement. Unfortunately, the values cannot be verified. Retreat then decreased gradually to a low of 2 m/yr. between 1970 and 1972 before rapidly increasing to approximately 53 m/yr. between 1979 and 1992. Overall, the Saskatchewan Glacier has experienced several distinct phases of retreat; minimum rates in the late 1800's and early 1970's with intervening periods of rapid retreat.

9.5 Summary

Evidence from trees killed by the glacier indicate that the Saskatchewan Glacier advanced at least three times in the last 350 yrs. (Table 9.3). Outer ring dates from three detrital logs in different locations on the northern trimline suggest that the earliest advance occurred between 1664 and 1678. Kill dates from one *in-situ* tree and two pieces of detrital wood along the trimline indicate that the glacier was advancing into forest between 1737 and 1740. The final grouping of dates had the largest population and was between 1824 and 1862. Three *in-situ* stumps along the trimline mark the limit of this advance, suggesting that it was the most extensive advance. The large amount of detrital material from this period supports the idea that this was the most extensive event, implying that the glacier was advancing further into the forest than it had during former advances. After reaching its maximum during the mid 19th century, the glacier retreated relatively slowly until the beginning of the 20th century. Retreat then gradually increased to a peak in the mid 1940's before dropping to much lower rates in the early 1970's. Over the last twenty years, the rate of recession has again increased.

Table 9.3

Major Periods of Glacier Advance at Saskatchewan Glacier

Event Date	Supporting Evidence
1664-1678	Outer ring dates of trees killed by the glacier
1737-1740	Outer ring dates of trees killed by the glacier
1824-1862	Outer ring dates of trees killed by the glacier

CHAPTER X
SYNTHESIS AND CONCLUSIONS

10.1 Introduction

The development of millennial length, reference tree-ring chronologies and previous work at Athabasca and Dome Glaciers suggested that precise dendroglaciological dates could be obtained from all major outlet glaciers of the Columbia Icefield. The objective of this thesis was to obtain chronologies for six of these glaciers using these techniques. Once obtained, these histories could be used to assess the synchroneity of the fluctuations of glaciers fed by the Icefield and their response to regional climate forcing during the LIA. Chapters V-IX have presented results from Castleguard, Kitchener, Columbia, Manitoba, Stutfield and Saskatchewan Glaciers. This concluding chapter integrates these data with previously reconstructed histories of Athabasca and Dome Glaciers (Luckman 1988) to provide a comprehensive chronology of glacier fluctuations at the Icefield.

This presentation and discussion will be divided into four parts. The first section will evaluate the limitations of the dating techniques used and problems encountered attempting to produce precise dating control for glacier advances. The second section presents a discussion of the overall synchroneity of events followed by a detailed outline of the chronology. The thesis will then conclude with recommendations for future research.

10.2 Problems with Developing Precise Dating Control

A primary objective of this study was to attempt to obtain precise dates for glacial events at the Columbia Icefield using dendroglaciology and to determine whether the outlet glaciers achieved their maximum positions synchronously. However, it was not possible to determine precise dendroglaciological dates at all outlet glaciers as a result of the absence of appropriate evidence. Prior to the comparison of the results between glaciers, an assessment of the various sources of error will be carried out. Limitations such as the absence of appropriate deposits, error terms associated with individual techniques, uncertainties arising from the comparison of results obtained using different techniques and the fact that the duration of the event being dated is unknown prevented the production of precise, calendar-dated glacial chronologies.

The major obstacle to dating the LIA maximum position at many of the sites was the lack of appropriate deposits. Downvalley, forested terminal moraine complexes provide optimal conditions for precise dating of the LIA maximum events. At many sites, these deposits have been subsequently destroyed or modified by outwash activity and the fragmentary nature of the remaining deposits restricts the ability to reconstruct the maximum position accurately as well as the availability of dateable surfaces. The oldest deposits at many sites were found in lateral positions where the morphology is more complex and sites may have been reoccupied several times by the glacier. In such situations, ice marginal positions were often very close together so there is evidence for multiple events in a small area.

Each dendroglaciological dating technique has a different degree of temporal precision (see Table 4.1). Unfortunately, the predominant technique used in the present study, minimum surface age, has the least well defined temporal precision because of the uncertainty associated with the determination of the ecesis interval. Commonly, the ecesis value is calculated based on contemporary deposits that may not reflect former conditions and are therefore not representative of older surfaces. Furthermore, the true ecesis interval is different for every moraine, at every site, depending on many microenvironmental conditions that cannot be accounted for. In some cases, with the available dating control it is difficult to determine with certainty whether two adjacent areas with minimum surface ages less than 30 yrs. apart are the result of distinct events, the result of local differences in ecesis period or random differences in dating. Therefore, use of an ecesis interval creates an additional source of uncertainty in dating.

One of the main goals of this study was to eliminate some of the dating uncertainty by employing the same, precise technique at all sites. Even though all sites except Kitchener Glacier had well developed trimlines, equivalent evidence was not available at each site. It was therefore necessary to use less precise techniques such as minimum surface ages and trees killed by the glacier to provide minimum and maximum dates for glacial advances. Therefore, while broadly similar dates were obtained for events at many of the sites, differences in dating uncertainty between techniques make it difficult to distinguish between single and multiple events at the same site that are closely spaced in time. To decide whether two similarly-aged deposits within a single forefield were created by the same or different events it is necessary to consider: 1) the morphological continuity of the deposits, 2) the location and relative proximity of the dated sites, 3) differences in age determination, 4) the error terms associated with different techniques and 5) the range of error terms applied with a specific technique (e.g. ecesis).

The availability of a range of estimates for the dating of a feature also poses questions about the expected range of dating for an event. Where only a single date is available, the duration of specific events is unknowable. Where several dates are available, the question is whether they belong to the same or multiple events. Therefore, single precisely dated events (e.g. tilting events) are not necessarily better estimates for dating the period of moraine formation: a range of dates for the same feature may actually be more representative. The ability to provide a precise calendar date for damage does not mean that the advance was restricted to that brief time interval. Glaciers may have remained at or close to their most advanced positions for several years and produce a number of scarring events in adjacent trees resulting in a range of damage dates associated with the same event. Without intermediate dates, it is difficult to determine whether two dates 20 years apart record the same or separate glacial advances. Conversely, the precision afforded by a single calendar date may be somewhat misleading because the duration of the glacial event is unknown. Unfortunately, most contemporary glaciers are presently retreating at increasing rates and therefore observation of present day glaciers does not allow the definition of the duration of a typical glacial advance. In addition, the complex interaction between climate, glaciers and their surrounding topography precludes the assumption that all advances have the same duration or that a value measured at one glacier would be applicable at any other. As a result, there is presently no viable way to estimate this type of error.

10.3 Synchroneity of Advances

Although precise chronologies could not be created for all glaciers investigated in this study, the data provide sufficient detail to assess the overall synchroneity of events. However, the fragmentary nature of the evidence for events prior to 1700 AD precludes a detailed evaluation of earlier events and the assessment of synchroneity is based on the chronology of events post 1700 AD (section 10.4.2). Evidence for multiple events exists at most sites and although there are broad similarities in the timing of advances at several sites, the outlets of the Icefield did not reach their maxima concurrently (Table 10.1). There is evidence of a significant mid 19th century advance at all sites, but only Athabasca, Dome, Saskatchewan and Castleguard Glaciers reached their maximum position during this period (Figure 10.1). The maxima at Columbia/Manitoba, Kitchener and Stutfield occurred during the early and middle parts of the 18th century concurrently with less extensive events at Athabasca and Saskatchewan Glaciers. The large difference in dating between these two periods of maximum extent indicate that these differences cannot simply be attributed to differences in the nature of the dating control. Although there was general synchroneity in the timing of glacial advances, they differed in relative magnitude between glaciers.

Examination of the characteristics of the outlet glaciers at the Columbia Icefield fails to provide a simple explanation for the observed grouping of glacier maximum dates. None of the typical attributes such as size, aspect, elevation, debris cover, morphology, presence of icefalls or terminal environments are consistent discriminators between the glaciers with 18th and 19th century maxima. For example, Stutfield, Kitchener and Dome are heavily debris covered glaciers in adjacent valleys on the east side of the Icefield. The similarities in morphology and setting leads to the expectation that the glaciers should have similar histories. While the elevation and size difference between Stutfield and Kitchener Glaciers may account for the slight difference in maximum dates between these two glaciers, the history of Dome Glacier is quite different. The lack of a single identifiable controlling factor for differences in the timing of the maximum position supports the idea that the glaciers are complex systems subject to a variety of controlling influences.

Table 10.1

Dates for Little Ice Age Maximum Positions at the Columbia Icefield

Site	Maximum
Columbia/Manitoba	1698-1739
Kitchener	<1713
Stutfield	**1758**-1767
Saskatchewan	1824-1862
Athabasca*	**1843-44**
Dome*	**1846**
Castleguard	<1858

Notes: * from Luckman 1988. Boldface indicates a precise date from a damaged tree.

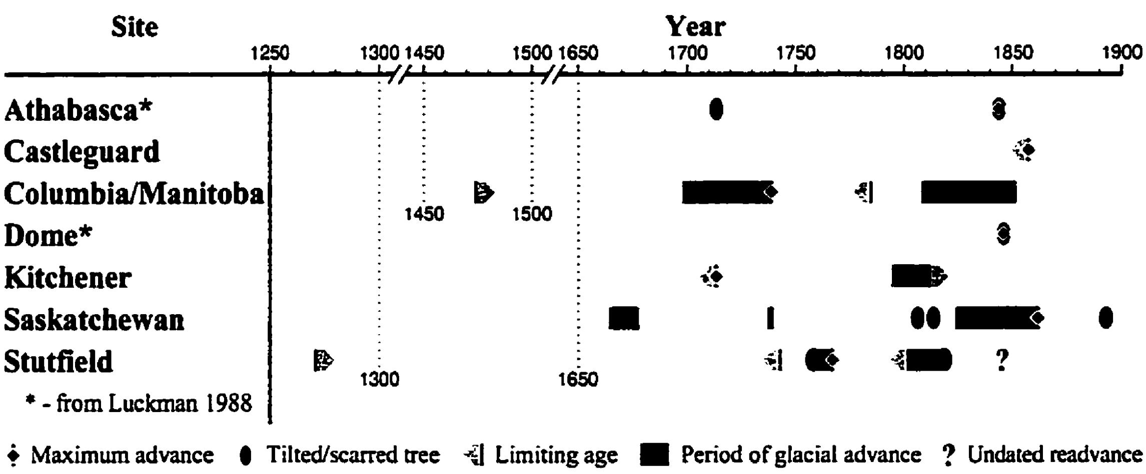


Figure 10.1 Major periods of glacial advance at the Columbia Icefield during the Little Ice Age

10.4 Chronology

The glacial histories for five sites at the Columbia Icefield were reconstructed in this thesis. Three of these glaciers, Castleguard, Kitchener and Stutfield had not been previously investigated. The new investigations at the Saskatchewan and Columbia/Manitoba sites enabled the development of more detailed and precise chronologies than were obtained in previous studies for these sites (e.g. Heusser 1956). Evidence for multiple periods of LIA glacial advance exist at all the sites investigated in this study except Castleguard Glacier (Figure 10.1). While it is common to find that only the evidence for the most extensive and subsequent advances have been preserved, the oldest recorded event was not the maximum downvalley extent at most sites. Conversely, absence of evidence for pre-19th century events at Dome and Castleguard Glaciers does not preclude their occurrence; it simply indicates that no evidence was preserved.

The overall timing of events reported at the Columbia Icefield is consistent with the regional picture that most LIA activity postdates 1700 AD (Figure 10.1) and the record for earlier periods is fragmentary. Therefore, the following discussion of glacial events is divided into two periods; before and after 1700 AD.

10.4.1 Evidence of Glacial Activity Prior to 1700 AD

As the late Little Ice Age advance was the most extensive at most glaciers, morphological and stratigraphic evidence documenting earlier glacier margins is incomplete and discontinuous. Despite this, this study reported evidence for four glacial events at the Columbia Icefield that predate 1700; one pre-LIA in age, two from the early LIA and one immediately prior to the late LIA advances.

10.4.1.1 Evidence of Pre Little Ice Age Glacial Activity

At Stutfield Glacier, Osborn (1993) has reported evidence of a Crowfoot and an early Neoglacial advance based on stratigraphic sections in the lateral moraines. In addition, detrital wood recovered by Luckman *et al.* (1993a, 1994) near the front of Saskatchewan Glacier suggests an advance equivalent to the Peyto Advance between 3300 and 2800 yrs. BP. However, the only morphological evidence for a possible pre-

LIA glacial event reported in this study was at Kitchener Glacier. The exact genesis of the outermost features at this site is uncertain, but their morphology and similarity to other glacial features suggest a glacial origin. Minimum surface ages suggest that they were formed prior to the early 16th century. However, the earliest ring dates of snags buried in the valley fill inset into the "kettled area" suggest that the glacier could not have occupied this position after the beginning of the 12th century. If these are glacial deposits, this evidence suggests that the maximum Holocene extent of Kitchener Glacier was achieved prior to the LIA. Pre-LIA tills also possibly occur in the lateral moraines of Columbia Glacier but radiocarbon dating of the buried logs recovered for that site have not been completed.

10.4.1.2 Evidence of Early Little Ice Age Glacial Activity

Evidence for the earliest dateable LIA advance comes from three logs buried in the lateral moraine of Stutfield Glacier. These logs indicate that the glacier advanced over them after 1271 AD and reached a position 10-15 m below the present moraine crest at that time. The till underlying these trees was deposited by one of the Neoglacial advances described by Osborn (1993). This evidence is consistent with radiocarbon dates obtained by Osborn (1993, 1996) from the same site and with calendar dated early LIA advances at Robson (1142-1150 AD, Luckman 1995) and Peyto Glaciers (1246-1375 AD, Luckman 1996c). Radiocarbon dates from glaciers located in the Premier Range, Purcell and Coast Mountains all document an advance between 900 and 600 ^{14}C yrs. BP (Luckman 1986, Osborn 1986, Ryder & Thomson 1986, Ryder 1987, Osborn & Karlstrom 1989, Desloges & Ryder 1990).

Evidence from two other sites in this study record glacial advances in the 15th and 17th centuries. Elsewhere, this period has a sparse and fragmentary record of glacial activity usually dated by minimum surface ages. As a result, dates are often poorly constrained and may significantly underestimate the age of these features. For example, at several sites it is quite possible that the timing of colonisation may reflect periods of more favourable conditions (e.g. upslope migration of trees) rather that immediate postglacial colonisation (Luckman 1995, Smith *et al.* 1995).

In this study, dating control for events between the 14th and 18th century was obtained from outer ring dates of trees killed during glacial advances. A log protruding from the inner face of the lateral moraine at Manitoba Glacier yielded an outer ring date of 1474. This log was deposited on the distal slope of a pre-existing moraine and buried by accreting debris from ice overtopping the former moraine crest. It indicates an advance in the late 15th century. These dates also indicate that the moraine was built up to within 1 m of its present height by the 1400's and stabilised relatively soon thereafter. The well developed forest cover on the distal moraine slope upvalley suggests that although the moraine was re-occupied in the 18th and 19th centuries, the later advances did not overtop the moraine crest over its entire length.

Kill dates from detrital wood at Saskatchewan Glacier range between 1543-1678 AD with three trees dating from 1664-1678. It is possible that these death dates represent a single advance that culminated in the early 18th century. However, in the absence of trees killed between 1678 and 1737, this evidence is interpreted as a separate event that culminated in the late 17th century. However, as is the situation with many of the other dates obtained in this study, it is impossible to distinguish with certainty whether they were two discrete events based on the available evidence.

10.4.2 Evidence of Glacial Activity After 1700 AD

In his review of the glacial history of the Canadian Rockies, Luckman (1996a) found that the majority of glaciers reached their maximum Holocene extent after 1700 AD, with the greatest concentration of moraines dated to the beginning of the 18th and the mid 19th century. The evidence obtained at the Columbia Icefield is consistent with this view, although conspicuous evidence for glacial activity was also identified in the mid 18th and early 19th centuries (Figure 10.1). Ironically, neither of these periods are represented in the histories of Athabasca and Dome Glaciers; two glaciers believed to depict the regional glacial history (e.g. Luckman 1996a). Therefore, while the glacial chronologies of Athabasca and Dome Glaciers may parallel the regional pattern of the Rocky Mountains, they do not reflect the complete, complex history of the Columbia Icefield.

As a result of the volume of evidence postdating 1700 AD, the evidence will be discussed on a century by century basis. The final section will focus on recession of the outlet glaciers of the Columbia Icefield in the 20th century.

10.4.2.1 Evidence of Eighteenth Century Glacial Activity

Evidence from the Columbia Icefield suggests two periods of glacial advance during the eighteenth century; in the early 1700's and mid to late 1700's. The first event is consistent with the regional chronology for the Canadian Rockies and was observed at all glaciers except Castleguard and Dome Glaciers where no evidence for pre-19th century events has been preserved. Both Columbia/Manitoba and Kitchener Glaciers reached their maximum LIA extent during these advances. A death date of 1698 for a tree deposited on the surface of the Manitoba lateral moraine and a minimum surface age of 1739 provide limiting dates for the most extensive advance at the Columbia/Manitoba site. A minimum surface age of 1713 at Kitchener Glacier indicates that it reached its maximum LIA extent during the same approximate period.

Evidence for similarly dated advances was also reported from Saskatchewan, Stutfield and Athabasca Glaciers. At Saskatchewan Glacier, three trees with tightly clustered death dates between 1737 and 1740 were found along the northern trimline, indicating that the glacier was advancing into the forest at that time. At Stutfield Glacier, a minimum surface age on the south side of the valley above the ice front indicates that the outermost lateral moraine had stabilised by 1743. However, unfavourable conditions for tree growth at this site suggest that it probably had a longer than normal ecesis interval and the moraine was likely deposited at the beginning of the 18th century, earlier than suggested by the minimum date. Downvalley, this moraine was overridden by a subsequent, more extensive event. Furthermore, previous studies by Heusser (1956) and Luckman (1988) found a tilted tree that indicates that the Athabasca Glacier advanced close to its maximum position in 1714.

Evidence for a second eighteenth century advance occurs at Columbia/Manitoba and Stutfield Glaciers where it is the LIA maximum position. A minimum surface age at the downvalley limit of the Stutfield Glacier forefield suggests an advance prior to 1767, consistent with a minimum surface age of 1758 along the northern "ledge". This is corroborated by a precise date of 1758 from a tilted tree at this site. The nature of the evidence used to date this event distinguishes it from the slightly older event reported earlier. The minimum surface age indicates that the ice had already retreated from the south side of Stutfield Valley prior to 1743, while the tilted tree on the north side of the valley indicates that the glacier was clearly advancing in 1758. A similarly dated event was also found at the Columbia/Manitoba site where a minimum surface age of 1785 was obtained for several moraines, mainly in the zone of coalescence of the Columbia and Manitoba Glaciers. The location in the forefield and the tightly clustered grouping of dates for features attributed to this event distinguish them from landforms created in other advances.

10.4.2.2 Evidence of Nineteenth Century Glacial Activity

In addition to the well documented mid 19th century glacier advance at most sites in the Canadian Rockies, there is evidence for an event at the beginning of the 19th century at Kitchener, Stutfield and Saskatchewan Glaciers. Minimum surface ages at Kitchener Glacier indicate that the "middle moraine" was formed between 1713 and 1812 AD. However, trees buried in the outwash attributed to that advance suggest burial continued after 1794 while trees growing on the moraine indicate that the surface was colonised by 1812. At Stutfield Glacier, there are several moraines with minimum surface ages dating to the early 1800's. Furthermore, tilted and scarred trees provide precise dates between 1812 and 1820 corresponding to a period of glacial advance. Two tilted trees found by Heusser (1956) along the northern trimline of Saskatchewan Glacier document at least one early 19th century ice thrusting event. However, neither tree was found during this investigation so the dates obtained from these trees could not be verified. Furthermore, the lack of corroborating evidence in the current study suggests that this event was likely a localised ice thrusting event and not a major advance.

The mid 19th century event documented at numerous sites throughout the Canadian Cordillera (e.g. Heusser 1956, Leonard 1981, Smith *et al.* 1995, Luckman 1996a) is also evident at all of the glaciers investigated at the Columbia Icefield. Tilted and scarred trees indicate that the Athabasca Glacier reached its maximum position in

1843/44 while the maximum of the adjacent Dome Glacier was reached in 1846 (Luckman 1988). Of the six glaciers investigated in this study, only the Castleguard and Saskatchewan Glaciers reached their maximum positions during this period. At Castleguard Glacier, the mid 19th century advance is the only event documented. This advance overrode any evidence of earlier events at this site. Abundant evidence from a tilted tree, *in-situ* and detrital wood at the trimline of Saskatchewan Glacier suggest that it was advancing to its most extensive position between 1824 and 1862. However, the preservation of *in-situ* trees damaged by former events suggests that the trimline was reoccupied during several events and has a complex history. At the Columbia/Manitoba site, there was considerable evidence from throughout both valleys for an advance between 1808 and 1852 to a position close to the maximum of the coalesced glaciers. At Kitchener and Stutfield Glaciers, there were no closely limiting ages for the innermost moraines. However, based on their location relative to other dated ice front positions, these features were likely formed during the mid 19th century.

Heusser (1956) reported a tilted tree along the northern trimline of Saskatchewan Glacier that indicates that the ice was at that position in 1893. Similar to the two trees with tilting events reported at the beginning of the 19th century, the exact location of this tree is unknown and corroborative evidence was lacking. Therefore, it too is believed to represent a localised ice thrusting event along a small section of the glacier rather than a major advance.

10.4.2.3 Twentieth Century Glacier History

There are photographic and historical observations of Columbia/ Manitoba, Athabasca, Dome, Castleguard and Saskatchewan Glaciers for the early 20th century, but no data are available for Kitchener and Stutfield Glaciers prior to the first aerial photographs (1948). The ice fronts at Athabasca and Saskatchewan Glaciers were surveyed from 1945 to 1980 by the Water Survey of Canada (e.g. WSC 1982), allowing more precise determination of recession than the estimates derived from aerial photographs or maps at the other sites.

The irregular spacing and limited number of ice front observations prior to 1948 limit detailed comparison of these records (Figure 10.2). However, several general comments can be made. Many of the earliest observations indicate that most glaciers remained close to their LIA maximum until the early 20th century (e.g. Field 1949). Relatively high rates of recession were observed during the 1940's and 50's at all glaciers except Columbia Glacier. Rates of recession were much lower during the 1960's and 70's and Kitchener and Columbia Glaciers readvanced. This is consistent with advances reported by Luckman *et al.* (1987) in the Premier Ranges, at the small glacier on the north face of Mt. Athabasca (Luckman *et al.* 1993b) and other sites in the Canadian Rockies.

Over the last twenty years, the debris covered glaciers - Manitoba, Stutfield and Kitchener - have shown little frontal recession while the Columbia and Saskatchewan Glaciers have experienced a steep increase in the rate of retreat. At Columbia Glacier this is likely a result of its floating snout. In this environment, the greatest loss of mass may be attributed to calving, which is strongly influenced by the water depth. At Saskatchewan Glacier, the increase in the rate of retreat may be related to the morphology of the glacier. Since the glacier is very clean and lacks an icefall, any changes in its overall mass balance will be readily observable through changes in length. This relatively unfiltered signal reflects the fact that the 1980's were the warmest decade in the instrumental climate record (Luckman *et al.* 1997).

10.5 Conclusions

Based on previous work at Athabasca and Dome Glaciers, the Columbia Icefield appeared to be an optimal location to develop precise chronologies of glacier fluctuations and to compare the history of its outlet glaciers. Unfortunately, precise chronologies could not be obtained at all sites as a result of differences in the types of evidence found. However, even if calendar dates had been obtained, there would still be a degree of uncertainty in correlations between sites because glacier advances are not instantaneous events. While there is no doubt that a single, precise date from a damaged tree represents an advance, there is no method to determine the period over which that advance occurred. Similar dates found on a contiguous feature may reflect different phases of the same

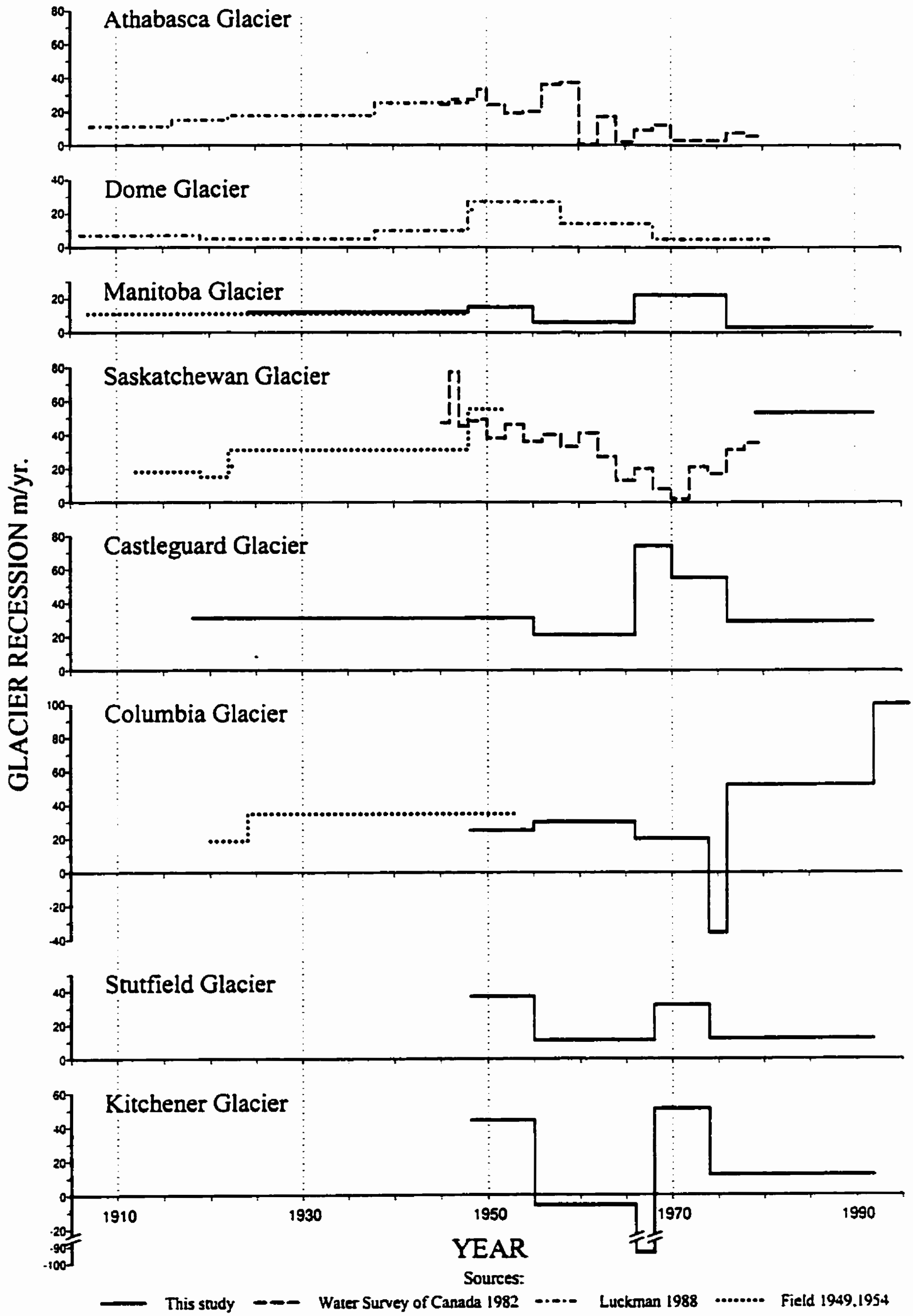


Figure 10.2 Frontal recession of the outlet glaciers of the Columbia Icefield. Note that negative values indicate a glacial advance.

advance or closely spaced, discrete events.

Although tightly constrained chronologies could not be developed at all sites, there is sufficient evidence to determine whether the responses of the outlet glaciers were synchronous. The evidence suggests that there were broad similarities in the timing of advances, although the maxima of the glaciers at the Columbia Icefield were not achieved concurrently. A significant mid 19th century advance was documented at all sites and Athabasca, Dome, Saskatchewan and Castleguard Glaciers reached their maximum position during this period. Columbia/Manitoba and Kitchener Glaciers achieved their maximum positions at the start of the 18th century, concurrent with less extensive events at Athabasca and Saskatchewan Glaciers, while the LIA maximum at Stutfield Glacier occurred in the mid 18th century. In addition, a considerable amount of evidence from Columbia/Manitoba, Kitchener, Saskatchewan and Stutfield Glaciers suggest an early 19th century advance. These two periods, the late 1700's and early 1800's represent two periods of glacial activity that had not been previously reported at the Columbia Icefield. Fragmentary evidence for early LIA events also suggests events that were previously unreported in this area. Trees buried in the lateral moraines at Stutfield and Manitoba Glaciers provide minimum ages of 1271 and 1474 AD, respectively, for two early LIA advances.

The lack of evidence from these two periods at Athabasca and Dome Glaciers suggests that they may not be as representative as first believed. Investigation of the other outlet glaciers of the Columbia Icefield have shown that these two glaciers have deceptively simple chronologies that do not accurately represent the complex history of the entire Icefield. In addition, the results from these and several other glaciers indicate that a simple morphology (of the trimline or moraines) does not necessarily result from a simple glacier history, e.g. Stutfield and Saskatchewan Glaciers. The local variability in both dating control and glacier history suggest that a large sample size is necessary to evaluate a regional glacial history.

10.6 Recommendations for Future Research

Although the production of precise chronologies for several glaciers from the same source area was not achieved in the present study, this objective should not be dismissed for future studies. Despite these limitations, it does provide valuable information about glacier history in this region. As is the case with exploratory studies such as this one, it is nearly impossible to predict success beforehand as it is dependant on the nature of the evidence found. However, a lack of precisely dated evidence for a glacial advance does not deny its existence, it only indicates that no evidence was found in the course of this study. It may be revealed in future studies at these sites.

While sufficient evidence was obtained in the present study to determine the overall synchroneity of glacier advances at the Icefield, additional studies may be useful for completing the chronology. The data obtained for Castleguard Glacier is basically a reconnaissance study because sampling was limited to the north side of the valley as a result of time constraints. Investigation of the southern trimline may reveal damaged or killed trees that may provide evidence of other glacial events. At Kitchener Glacier, more extensive investigation, including an analysis of the deposits and surface material is required to determine the genesis and age of the outermost features. Evidence from both the Stutfield and Columbia/Manitoba sites indicate complex histories. Although reasonably comprehensive chronologies were produced for both sites, the multitude of deposits suggests that considerably more detail could be obtained with more intensive sampling, e.g. using proper climbing equipment along the steep lateral moraines close to the ice fronts. At Saskatchewan Glacier, there is no shortage of detrital wood to be sampled in the assemblage along the upper moraines. A larger sample population may enable the identification of periods of glacial advance more precisely. A final suggestion to produce a complete and accurate picture of the Columbia Icefield would be to investigate other outlet glaciers, most notably the glaciers on the south end of the Icefield flowing into the Bush River and Bryce Creek valleys. Further investigations will inevitably result in the production of a more complete glacial history of the Columbia Icefield. A more complete glacial history enables comparisons with other paleoenvironmental records, e.g. dendrochronological temperature reconstructions.

APPENDIX I

APPENDIX I

Height-Age Relationship for the Ecesis Samples at Columbia Glacier

Height	CE0	CE1	CE2	CE3	CE4	CE5	CE6	CE7	CE8	CE9
10	1963	1963	1963	1969	1961	1954	1966	1955	1954	1960
9	1963	1961	1963	1967	1961	1954	1966	1955	1954	1959
8	1962	1960	1963	1965	1960	1953	1965	1954	1953	1958
7	1962	1959	1962	1964	1960	1953	1965	1954	1953	1958
6	1961	1958	1962	1963	1960	1952	1965	1953	1952	1957
5	1961	1956	1962	1961	1960	1952	1964	1951	1951	1956
4	1960	1955	1961	1959	1959	1951	1964	1950	1951	1956
3	1958	1954	1961	1957	1959	1950	1964	1949	1950	1955
2		1953			1956		1963	1947	1949	1955
1		1952					1961		1949	1954
0									1947	

APPENDIX II

APPENDIX II

Sample Data from the Castleguard Site

Location	Sample Identification	Earliest Ring	Proximity to Pith	Pith Date	Surface Date (30 yr. ecesis)
		Cores			
Outermost	G9604	1888	P	1888	1858
moraine	G9603	1894	NP	1889	1859
	G9615	1890	P	1890	1860
	G9624	1901	NP	1896	1866
	G9606	1903	NP	1898	1868
	G9612	1905	C	1903	1873
	G9621	1908	P	1908	1878
	G9616	1915	NP	1910	1880
	G9608	1913	C	1911	1881
	G9605	1922	OP	1912	1882
	G9614	1917	NP	1912	1882
	G9625	1919	C	1917	1887
	G9623	1928	OP	1918	1888
	G9609	1921	C	1919	1889
	G9618	1924	NP	1919	1899
	G9613	1921	P	1921	1891
	G9610	1924	C	1922	1892
	G9602	1928	P	1926	1896
	G9619	1937	OP	1927	1897
	G9611	1936	NP	1931	1901
	G9607	1935	C	1932	1902
	G9617	1937	NP	1932	1902
	G9622	1944	OP	1934	1904
	G9601	1948	OP	1938	1908
	G9620	1947	NP	1942	1912

Note: P - contains pith, C - < 2 yrs. from pith, NP - < 5 yrs. from pith, OP - 5-10 yrs. from pith and >OP - > 10 yrs. from pith

APPENDIX III

APPENDIX III

Sample Data from the Kitchener Site

Location	Sample Identification	Cores Earliest Ring	Proximity to Pith	Pith Date	Surface Date
Ridge a[a]	X9619	1722	OP	1712	1692
	X9621	1773	>OP	1758	1738
	X9620	1811	NP	1806	1786
	X9622	1830	OP	1820	1780
Ridge b	X9623[a]	1642	NP	1637	1617
	X9625[a]	1761	C	1759	1739
	X9614[b]	1814	NP	1809	1779
	X9613[b]	1890	OP	1880	1850
	X9624[a]	1927	OP	1917	1897
	X9617[b]	1947	OP	1937	1907
	X9615[b]	1942	C	1940	1910
	X9616[b]	1954	OP	1944	1914
Ridge c[b]	K9605	1745	C	1743	1713
	K9606	1806	C	1804	1774
	K9601	1846	P	1846	1816
	K9602	1865	NP	1860	1830
	K9603	1944	OP	1934	1904
	X9601	1945	C	1943	1913
Ridge d[b]	X9602	1967	NP	1962	1932
Ridge e[b]	K9604	1847	NP	1842	1812
"Kettled" Area[a]	K9616	1715	OP	1705	1685
	K9613	1733	OP	1723	1703
	K9612	1734	OP	1724	1704
	K9615	1746	NP	1741	1721
	K9611	1847	P	1847	1827
	K9614	1897	OP	1887	1867
Terrace on the east side of the valley[b]	X9604	1850	P	1850	1820
	X9605	1871	OP	1861	1831
	X9603	1905	NP	1900	1870
	X9607	1917	OP	1907	1877

Appendix III (cont'd)

Location	Sample Identification	Earliest Ring	Proximity to Pith	Pith Date	Surface Date
Slope above the terrace[b]	X9610	1823	>OP	1807	1777
	X9611	1912	OP	1902	1872
	X9612	1910	C	1908	1878
	X9609	1938	NP	1933	1903
	X9608	1941	NP	1937	1907

Note: P - contains pith, C - < 2 yrs. from pith, NP - < 5 yrs. from pith, OP - 5-10 yrs. from pith and >OP - > 10 yrs. from pith. Superscripts: a - 20 yrs. ecesis, b - 30 yrs. ecesis.

Appendix III (cont'd)

Location	Sample Identification	Calendar Dates	Comments
	Cross Sections		
Valley Fill	K9651	NM	
	K9652	NM	
	K9653	1539-1768	Two piths in the
		1539-1781	section
	K9654	1423-1782	
		1414-1782	
		1405-1783	
		1405-1783	
	K9655	NM	
	K9656	1447-1778	Not to pith
		1417-1772	
		1564-1775	
	K9657	1672-1794	
		1672-1794	
	K9660	X	225 rings
	X9651	1139-1311	
		1139-1317	
	X9652	X	
	X9653	1334-1686	
		1334-1704	
	X9654	NM	
	X9655	X	
	X9656	1628-1781	
		1625-1790	
	X9657	NM	
"Kettled" Area	K9650	1543-1803	Centre's rotten
		1545-1758	
	K9658	1595-1890	
		1595-1898	
		1595-1896	
	K9659	1591-1893	
		1591-1884	
Terrace	X9650	X	216 rings

Appendix III (cont'd)

Location	Sample Identification	Calendar Dates	Comments
Ridge b	X9658	1637-1995	Asymmetric growth pattern
	X9659	1712-1970	
		1769-1969	
	X9660	NM	
	X9661	1582-1906	
		1614-1863	
		1576-1901	
		1578-1901	
	X9662	X	214 rings
	X9663	1591-1885	
		1591-1883	

Note: X - unsuccessful crossdate, NM - not measured

APPENDIX IV

APPENDIX IV

Sample Data from the Columbia/Manitoba Site

Location	Sample Identification	Cores Earliest Ring	Proximity to Pith	Pith Date	Surface Date (20 yr. ecesis)
		Manitoba Valley			
Eastern lateral moraine	M9613	1806	OP	1796	1776
	M9611	1827	NP	1822	1802
	M9601	1867	OP	1857	1837
	M9612	1852	OP	1842	1842
Eastern junction Moraine parallel to Atha	F9610	1901	OP	1891	1871
Below trimline	M9640	1777	C	1775	1755
	C9602	1824	NP	1819	1799
	M9641	1843	OP	1833	1813
	M9642	1903	C	1901	1881
Ridge a	M9644	1840	C	1838	1818
	M9643	1871	NP	1866	1846
Ridge b	M9645	1862	NP	1857	1837
	M9648	1877	P	1877	1857
	M9647	1900	NP	1895	1875
	M9646	1912	NP	1907	1887
Western main lateral moraine	M9635	1806	C	1804	1784
	M9634	1822	NP	1817	1797
	M9637	1824	P	1824	1806
	M9636	1844	NP	1839	1819
Below trimline	M9633	1771	OP	1761	1741
	M9632	1802	NP	1797	1777
	M9630	1818	P	1818	1798
	M9631	1854	OP	1844	1824

Appendix IV (cont'd)

Location	Sample Identification	Earliest Ring	Proximity to Pith	Pith Date	Surface Date (20 yr. ecesis)
Between trimline and moraine c	C9630	1790	P	1790	1770
	M9682	1821	OP	1811	1791
	M9680	1828	NP	1823	1803
	C9631	1849	OP	1839	1819
	M9681	1916	OP	1906	1886
		Zone of Coalescence			
Above moraine e	M9674	1818	NP	1813	1793
	M9677	1832	NP	1827	1807
	M9671	1834	C	1832	1812
	M9676	1838	NP	1833	1813
	M9678	1841	NP	1836	1816
	M9673	1857	NP	1852	1832
	M9679	1863	NP	1858	1838
	M9672	1875	>OP	1860	1840
	M9670	1870	P	1870	1850
	M9675	1826	C	1824	1904
Between moraines c and d	M9690	1884	P	1884	1864
Moraine in forest	C9625	1770	OP	1760	1740
	C9623	1796	OP	1786	1766
	C9622	1789	P	1789	1769
	C9621	1816	>OP	1801	1781
	C9624	1834	OP	1824	1804
	C9620	1845	OP	1835	1815
Outwash	C9612	1814	P	1814	1794
	C9610	1845	OP	1835	1815
	C9606	1851	NP	1846	1826
	C9607	1852	NP	1847	1827
	C9609	1920	NP	1915	1895
	C9608	1928	NP	1923	1903

Appendix IV (cont'd)

Location	Sample Identification	Earliest Ring	Proximity to Pith	Pith Date	Surface Date (20 yr. ecesis)
Moraine II	MS200	1811	C	1809	1789
	MS203	1815	NP	1810	1790
	MS201	1817	NP	1812	1792
	MS202	1843	NP	1838	1818
	MS205	1951	P	1951	1931
Moraine I	MS210	1814	NP	1809	1789
Outwash between valley wall and moraines	MS211	1813	P	1813	1793
	MS212	1855	OP	1845	1825
	MS213	1856	OP	1846	1826
	MS215	1884	NP	1879	1859
	MS214	1895	OP	1885	1865
Moraine IV	C96108	1902	NP	1897	1877
	C96109	1898	P	1898	1878
	C96110	1906	NP	1901	1881
	C96105	1913	NP	1908	1888
	C96101	1917	NP	1912	1892
	C96107	1917	NP	1912	1892
	C96112	1917	P	1917	1897
	C96106	1923	NP	1918	1898
	C96104	1938	NP	1933	1913
	C96103	1944	P	1944	1924
	C96111	1959	NP	1954	1934
Moraine north of moraine IV	C96151	1910	C	1908	1888
	C96150	1925	NP	1920	1900
	C96152	1936	OP	1926	1906
Moraine III	C96129	1838	P	1838	1818
	C96133	1847	P	1847	1827
	C96128	1866	C	1864	1844
	C96120	1869	C	1867	1847
	C96132	1875	NP	1870	1850
	C96122	1876	NP	1871	1851
	C96134	1879	C	1877	1857
	C96124	1882	C	1880	1860
	C96130	1893	NP	1888	1868

Appendix IV (cont'd)

Location	Sample Identification	Earliest Ring	Proximity to Pith	Pith Date	Surface Date (20 yr. ecesis)
Moraine III	C96131	1894	C	1892	1872
	C96135	1897	NP	1892	1872
	C96127	1896	C	1894	1874
Moraine island NW of island III	C96139	1808	P	1808	1788
	C96138	1854	C	1852	1832
	C96136	1876	C	1874	1854
	C96137	1948	NP	1943	1923
Athabasca Valley					
Contiguous till in centre of the valley	C96162	1907	P	1907	1887
	C96161	1930	OP	1920	1900
	C96160	1942	NP	1937	1917
Area B	CC9604	1886	P	1886	1866
	CC9605	1902	P	1902	1882
	MS218	1916	P	1916	1896
	CC9606	1937	OP	1927	1907
	CC9603	1934	C	1932	1912
Island f	CC9602	1882	OP	1872	1852
	CC9608	1890	P	1890	1870
	CC9607	1922	NP	1917	1897
Area C	CN9682	1842	P	1842	1822
	CN9683	1848	P	1848	1828
	CN9684	1869	C	1867	1847
	CN9681	1873	NP	1868	1848
	CN9680	1921	OP	1911	1891

Appendix IV (cont'd)

Location	Sample Identification	Earliest Ring	Proximity to Pith	Pith Date	Surface Date (20 yr. ecesis)
		North valley wall			
Above terrace	CN9603	1761	C	1759	1739
	CN9601	1887	NP	1882	1862
	CN9602	1898	NP	1893	1873
Above Moraine	CN9629	1796	C	1794	1774
	CN9635	1918	>OP	1903	1883
	CN9636	1920	NP	1915	1895
Below trimline	CN9648	1892	P	1892	1872
	CN9649	1893	P	1893	1873
	CN9645	1894	P	1894	1874
	CN9646	1894	P	1894	1874
	CN9643	1898	C	1896	1876
	CN9638	1897	P	1897	1877
	CN9637	1927	OP	1917	1897
	CN9640	1936	NP	1931	1911
	CN9641	1941	NP	1936	1916
Valley side terrace	CN9615	1877	C	1875	1855
	CN9604	1886	NP	1881	1861
	CN9611	1894	NP	1889	1869
	CN9609	1895	NP	1890	1870
	CN9606	1898	P	1898	1878
	CN9616	1900	P	1900	1880
	CN9614	1901	P	1901	1881
	CN9612	1912	OP	1902	1882
	CN9613	1915	NP	1910	1890
Valley side moraine	CN9623	1887	NP	1882	1862
	CN9621	1892	C	1890	1870
	CN9631	1902	NP	1897	1877
	CN9626	1905	NP	1900	1880
	CN9634	1905	NP	1900	1880
	CN9622	1902	P	1902	1882
	CN9632	1904	C	1902	1882
	CN9633	1908	NP	1903	1883
	CN9630	1911	NP	1906	1886

Appendix IV (cont'd)

Location	Sample Identification	Earliest Ring	Proximity to Pith	Pith Date	Surface Date (20 yr. ecesis)
Valley side moraine	CN9624	1920	OP	1910	1890
	CN9620	1915	C	1913	1893
	CN9627	1928	NP	1923	1903
	CN9628	1942	OP	1932	1912
Valley floor moraine	CN96BHLXX	1939	NP	1934	1914

Note: P - contains pith, C - < 2 yrs. from pith, NP - < 5 yrs. from pith, OP - 5-10 yrs. from pith and >OP - > 10 yrs. from pith

Appendix IV (cont'd)

Sections

Location	Sample Identification	Calendar Dates	Comments
On east Manitoba lateral moraine	M9660	1421-1656	
		1420-1690	
	M9661	1658-1787	
		1658-1796	
	M9662	1661-1798	Scar 1694
		1661-1796	
	M9663	1426-1698	
		1426-1697	
	M9664	1593-1808	
		1593-1773	
		1593-1774	
In eastern Manitoba lateral moraine	M9654	1425-1475	Lot of rot
		1425-1477	
	M9655	X	128 rings
Below trimline at eastern valley junction	M9642	1777-1986	
		1777-1981	
	M9650	1795-1989	
		1795-1988	
	M9651	NM	
Distal slope of eastern Manitoba lateral moraine	M9652	NM	
Western trimline in Manitoba Valley	C9656	1791-1989	
		1791-1990	
Forested moraine on south wall of the Athabasca Valley	C9650	1642-1788	
		1642-1789	
	C9651	1772-1877	
		1773-1879	
	C9652	X	158 rings
	C9653	1658-1881	Small scar 1837, 1847
		1658-1880	
	C9654	1609-1831	
		1609-1834	
		1609-1809	

Appendix IV (cont'd)

Location	Sample Identification	Calendar Dates	Comments
Trimline on south wall of Athabasca Valley	C9655	X	97 rings
Outwash north of island I	MS204	X	161 rings
	MS206	X	129 rings
	MS207	X	127 rings
	MS216	X	179 rings
	MS217	X	171 rings
Terrace on north wall of Athabasca Valley	CN9607	1740-1990	
		1726-1990	
Moraine on north wall	CN9650	1860-1995	
of Athabasca Valley	CN9651	1758-1841	
		1758-1842	

Note: X - unsuccessful crossdate, NM - not measured

APPENDIX V

APPENDIX V

Sample Data from the Stutfield Site

Location	Sample Identification	Cores Earliest Ring	Proximity to Pith	Pith Date	Germination Date (30 yr. ecesis)
Outer north lateral moraine	T9632	1840	P	1840	1810
Between outer moraine and bedrock on north "ledge"	T9714	1750	>OP	1735	1705
	T9701	1756	OP	1746	1716
	T9722	1792	OP	1782	1752
	T9702	1788	NP	1783	1753
	T9720	1789	NP	1784	1754
	T9715	1896	OP	1886	1856
	T9713	1914	>OP	1899	1869
North till "ledge"	T9711	1850	NP	1845	1815
	T9712	1859	OP	1849	1819
	T9710	1873	OP	1863	1833
	T9716	1878	NP	1873	1843
	T9717	1881	NP	1876	1746
	T9636	1892	OP	1882	1852
	T9640	1893	P	1893	1863
	T9638	1900	NP	1895	1865
	T9635	1900	C	1898	1868
	T9641	1898	NP	1893	1863
	T9721	1932	>OP	1917	1887
	T9718	1933	>OP	1918	1888
Area b	Y9683	1789	C	1787	1757
	Y9674	1801	OP	1791	1761
	Y9678	1799	NP	1794	1764
	Y9677	1809	OP	1799	1769
	Y9675	1811	OP	1801	1771
	Y9680	1809	NP	1804	1774
	Y9673	1822	OP	1812	1782
	Y9682	1845	OP	1835	1805
	Y9681	1847	NP	1842	1814
	Y9676	1839	OP	1929	1899

Appendix V (cont'd)

Location	Sample Identification	Earliest Ring	Proximity to Pith	Pith Date	Germination Date (30 yr. ecesis)
Ridge c	Y9662	1798	P	1798	1768
	Y9666	1814	C	1812	1782
	Y9657	1876	NP	1871	1841
	Y9670	1822	OP	1812	1782
	Y9668	1835	OP	1825	1795
	Y9671	1834	P	1834	1804
	Y9672	1834	P	1834	1804
	Y9658	1842	NP	1837	1807
	Y9663	1864	>OP	1849	1819
	Y9667	1857	P	1857	1827
	Y9665	1875	OP	1865	1835
	Y9669	1868	P	1868	1838
	Y9661	1887	P	1887	1857
	Y9664	1901	NP	1896	1866
	Y9659	1912	OP	1902	1872
	Y9660	1955	>OP	1940	1910
Below trimline and above ridge c	T9629	1847	NP	1842	1812
	T9628	1853	NP	1848	1818
	T9630	1892	NP	1887	1857
	T9631	1895	NP	1890	1860
Ridge d	Y9651	1832	OP	1822	1792
	Y9653	1900	OP	1890	1860
	Y9656	1910	>OP	1895	1865
	Y9655	1897	P	1897	1867
	Y9654	1912	OP	1902	1872
	Y9652	1918	OP	1908	1878
Island g	T9604	1835	C	1833	1803
	T9601	1845	P	1845	1815
	T9607	1857	NP	1852	1822
	T9603	1871	C	1869	1839
	T9602	1903	OP	1893	1863
	T9606	1897	C	1895	1865
	T9605	1910	NP	1905	1875
	T9608	1940	C	1938	1908
	T9609	1838	NP	1833	1803
	T9610	1840	C	1838	1808

Appendix V (cont'd)

Location	Sample Identification	Earliest Ring	Proximity to Pith	Pith Date	Germination Date (30 yr. ecesis)
Island g	T9611	1850	NP	1845	1815
	T9612	1889	NP	1884	1854
	Y9645	1838	NP	1833	1803
Island f terrace	T9623	1825	P	1825	1795
	T9618	1828	C	1826	1796
	T9615	1827	P	1827	1797
	T9621	1845	OP	1835	1805
	T9619	1838	P	1838	1808
	T9617	1849	NP	1844	1814
	T9620	1850	NP	1845	1815
	T9616	1854	NP	1849	1819
	T9613	1860	NP	1855	1825
	T9622	1865	NP	1860	1830
	T9624	1899	NP	1894	1864
Island f	T9625	1917	P	1917	1887
	T9627	1945	OP	1935	1905
	T9626	1941	P	1941	1911
Outwash by island g	Y9646	1845	C	1843	1813
	Y9649	1878	C	1876	1846
	Y9650	1882	P	1882	1852
	Y9648	1902	P	1902	1872
	Y9647	1942	NP	1937	1907
North outwash	Y9689	1863	P	1863	1833
	Y9686	1872	P	1872	1842
	Y9684	1902	NP	1897	1867
	Y9687	1918	OP	1908	1878
	Y9685	1928	NP	1923	1893
	T9689	1918	C	1916	1886
Ridge i	T9674	1842	NP	1837	1807
	Y9638	1850	NP	1845	1815
	T9645	1855	NP	1850	1820
	T9644	1859	NP	1854	1824
	Y9637	1868	NP	1863	1833
	T9679	1869	P	1869	1839

Appendix V (cont'd)

Location	Sample Identification	Earliest Ring	Proximity to Pith	Pith Date	Germination Date (30 yr. ecesis)
Ridge i	T9671	1884	NP	1879	1849
	Y9643	1885	C	1883	1853
	T9642	1887	P	1887	1857
	Y9636	1892	P	1892	1862
	Y9634	1914	NP	1909	1879
	Y9635	1935	OP	1925	1895
	T9643	1961	OP	1951	1921
Between ridges i and j	T9646	1773	P	1773	1743
	T9648	1785	NP	1780	1750
	T9677	1797	P	1797	1767
	T9678	1797	P	1797	1767
	T9680	1798	P	1798	1768
	T9673	1813	C	1811	1781
	Y9641	1831	NP	1826	1796
	T9686	1833	C	1831	1801
	Y9639	1854	NP	1849	1819
	T9675	1852	P	1852	1822
	T9676	1854	P	1854	1824
	T9670	1863	C	1861	1831
	Y9640	1869	C	1867	1837
	Y9644	1876	NP	1871	1841
	T9649	1885	OP	1875	1845
	T9682	1890	OP	1880	1850
	Y9642	1891	OP	1881	1851
	T9681	1883	P	1883	1853
	T9684	1890	NP	1885	1855
	T9683	1886	P	1886	1856
	T9672	1932	NP	1927	1897
	T9647	1958	OP	1948	1918
Washed till south of river	T9699	1841	C	1839	1809
	Y9628	1851	C	1849	1819
	T9694	1852	C	1850	1820
	Y9606	1883	NP	1878	1848
	T9698	1880	P	1880	1850
	Y9627	1880	P	1880	1850
	T9687	1898	NP	1893	1863
	Y9629	1903	OP	1893	1863

Appendix V (cont'd)

Location	Sample Identification	Earliest Ring	Proximity to Pith	Pith Date	Germination Date (30 yr. ecesis)
Washed till south of river	T9692	1904	P	1904	1874
	Y9630	1925	NP	1920	1890
	Y9605	1932	NP	1927	1897
	T9697	1935	P	1935	1905
	T9691	1955	OP	1945	1915
Ridge l	Y9603	1853	NP	1848	1818
	T9693	1859	P	1859	1829
	T9696	1866	NP	1861	1831
	Y9602	1880	NP	1875	1845
	T9688	1888	OP	1878	1848
	T9690	1887	P	1887	1857
	Y9601	1892	P	1892	1862
Unwashed till south of the river	Y9604	1814	NP	1809	1779
	Y9617	1809	P	1809	1779
	Y9611	1817	P	1817	1787
	Y9607	1855	NP	1850	1820
	Y9612	1862	NP	1857	1827
	Y9618	1857	P	1857	1827
	Y9608	1860	P	1860	1830
	Y9610	1871	C	1869	1839
	Y9616	1875	NP	1870	1840
	Y9615	1872	P	1872	1842
	Y9609	1887	NP	1882	1852
	Y9619	1906	OP	1896	1866
	Y9613	1902	NP	1897	1867
	Y9614	1957	NP	1952	1922
	Y9623	1957	P	1957	1927
Moraine m	Y9620	1809	P	1809	1779
	Y9626	1820	NP	1815	1785
	Y9625	1836	OP	1826	1796
	Y9621	1834	C	1832	1802
	Y9624	1845	P	1845	1815
	Y9631	1859	OP	1849	1819
	Y9632	1889	C	1887	1857
	Y9633	1924	NP	1919	1889

Note: P - contains pith, C - < 2 yrs. from pith, NP - < 5 yrs. from pith, OP - 5-10 yrs. from pith, >OP - > 10 yrs. from pith

Appendix V (cont'd)

Cross Sections

Location	Sample Identification	Calendar Dates	Comments
In north lateral moraine	T9601	855-1136	
		855-1116	
		849-1110	
	T9602	959-1197	
		959-1271	
		959-1265	
	T9603	894-1059	
		894-1064	
	T9604	889-1107	resample of T9601
		891-1133	
Island f	T9650	X	128 rings, reaction wood
	T9651	X	101 rings, reaction wood
Island f terrace	T9652	X	136 rings
	T9653	1807-1959	
		1807-1964	
		1807-1962	
	T9664W	1819-1993	Scars in 1957, 1973 and 1986
		1819-1983	
Outer moraine on northern till "ledge"	T9654	1694-1995	Reaction wood series 1758 and 1834 - coincides with traumatic resin canals
		1737-1995	
	T9655-upper	1791-1864	Reaction wood series 1812
		1781-1868	
	T9655-lower	1779-1868	Reaction wood series 1812. Scars 1815, 1818, 1835
		1779-1866	
	T9656	1779-1969	Reaction wood to the innermost ring, not to pith, centre's rotted
		1779-1963	
Northern till "ledge"	T9657	1788-1995	V. tight
	T9658	1791-1995	V. tight

Appendix V (cont'd)

Location	Sample Identification	Calendar Dates	Comments
Ridge i	T9660	1893-1995	V. tight
Ridge j	T9661	1868-1995	V. tight
Between ridge i and j	T9659	1784-1995	V. tight
	T9662	X	140 rings
	T9663	1823-1995	V. tight
	T9664	1849-1995	V. tight
Ridge c	T9665	1894-1983	
		1907-1983	
	T9666	1833-1984	
		1834-1982	
		1827-1984	
		1827-1988	
	T9668	X	151 rings
	T9669	NM	
Northern trimline	T9667	X	166 rings
	W9651	X	302 rings

Note: X - unsuccessful crossdate, NM - not measured

APPENDIX VI

APPENDIX VI

Sample Data from the Saskatchewan Site

	Cores				
Location	Sample Identification	Earliest Ring	Proximity to Pith	Pith Date	Surface Date
Forest above the northern trimline	S9611[a]	1814	OP	1804	1784
	S9610[a]	1888	OP	1878	1858
	S9601[a]	1946	OP	1936	1916
	S9620[a]	1964	P	1964	1944
East bedrock knoll	S9621[b]	1912	NP	1907	1877
	S9622[b]	1931	OP	1921	1891
Moraine island	S9626[b]	1947	OP	1937	1907
	S9625[b]	1953	NP	1948	1918
	S9623[b]	1955	P	1955	1925
	S9624[b]	1971	OP	1961	1931
Moraine a	S9608[b]	1929	NP	1924	1894
	S9632[b]	1926	P	1926	1896
	S9634[b]	1933	NP	1928	1898
	S9606[b]	1931	C	1929	1899
	S9607[b]	1930	P	1930	1900
	S9633[b]	1940	NP	1935	1905
	S9631[b]	1948	OP	1938	1908
	S9630[b]	1956	OP	1946	1916
Moraine b	S9642[b]	1935	OP	1925	1895
	S9640[b]	1948	NP	1943	1913
	S9641[b]	1962	OP	1952	1922
Southern terrace	S9684[a]	1671	P	1671	1651
	S9693[a]	1684	OP	1674	1654
	S9686[a]	1694	P	1694	1674
	S9697[a]	1714	P	1714	1694
	S9698[a]	1749	OP	1739	1719
	S9687[a]	1758	OP	1748	1728
	S9694[a]	1759	OP	1749	1729
	S9695[a]	1791	NP	1786	1766
	S9688[a]	1809	NP	1804	1784
	S9690[a]	1832	OP	1822	1802
	S9691[a]	1842	OP	1832	1812
	S9692[a]	1859	>OP	1844	1824

Appendix VI (cont'd)

Location	Sample Identification	Earliest Ring	Proximity to Pith	Pith Date	Surface Date
Southern terrace	S9681[a]	1892	OP	1882	1862
	S9683[a]	1895	NP	1890	1870
	S9685[a]	1904	OP	1894	1874
	S9689[a]	1912	OP	1902	1882
	S9680[a]	1918	OP	1908	1888
	S9682[a]	1926	OP	1916	1896
	S96100	1906	OP	1896	1876

Note: P - contains pith, C - < 2 yrs. from pith, NP - < 5 yrs. from pith, OP - 5-10 yrs. from pith, >OP - > 10 yrs. from pith. Superscripts: a - 20 yrs. ecesis, b - 30 yrs. ecesis.

Appendix VI (cont'd)

Cross Sections

Location	Sample Identification	Calendar Dates	Comments
Embayment between the bedrock knolls	S9655	1778-1953 NM	Heusser's tree, Major reaction wood series 1830. Reaction wood 1796, 1807
	S9665-lower	1536-1733	A lot of rot, not to pith
		1556-1726	
	S9665-upper	1507-1740	A lot of rot, not to pith
		1506-1726	
Eastern bedrock knoll	S9671	1408-1668	Not to pith
		1407-1669	
	S9672	1417-1672	
		1417-1678	
East of the creek	S9667	1707-1814	
		1704-1838	
		1704-1842	
	S9668	1664-1858	Not to pith
		1705-1831	
Tangle of wood along outer moraines	S9650	1598-1810	
		1598-1838	
	S9651	1482-1862	Suppression in mid 1700's
		1482-1849	
	S9652	1515-1806	
		1515-1802	
		1515-1799	
	S9653	1414-1564	
		1424-1588	
	S9656	1403-1540	
		1436-1543	
	S9657	X	191 rings
	S9658	1419-1629	
		1422-1613	
	S9659	X	85 rings
	S9660	X	284 rings
	S9661	1423-1598	
		1423-1573	
		1423-1664	

Appendix VI (cont'd)

Location	Sample Identification	Calendar Dates	Comments
Tangle of wood along outer moraines	S9662	1669-1856	
		1683-1851	
	S9663	X	178 rings
	S9669	X	199 rings, scarred
	S9670	X	164 rings, very wide rings in centre of section
	S9752	1667-1798	
		1667-1776	
	S9753	1477-1824	
		1477-1816	
	S9754	1520-1739	Not to pith
		1515-1739	
	S9755	1629-1847	Not to pith
		1631-1846	
	S9756	1534-1737	
		1534-1723	
	S9757	1483-1710	
		1470-1694	
	S9758	X	189 rings, not to pith

Note: X - unsuccessful crossdate, NM - not measured

APPENDIX VII

APPENDIX VII

Recession of Saskatchewan Glacier

Period	Distance of Retreat (m)	Rate of Retreat (m/yr.)	Source
mid 19th century - 1912	170	-	Field 1949
1912-1919	125	18	Field 1949
1919-1922	45	15	Field 1949
1922-1948	800	31	Field 1949
1948-1953	275	55	Field & Heusser 1954
1854-1911	122	-	McPherson & Gardner 1969
1912-1924	168	14	McPherson & Gardner 1969
1925-1944	632	33	McPherson & Gardner 1969
1945-1950	265	64	McPherson & Gardner 1969
1951-1956	240	48	McPherson & Gardner 1969
1957-1962	225	45	McPherson & Gardner 1969
1963-1966	80	27	McPherson & Gardner 1969
1945-1946	N/A	47	WSC 1982
1946-1947	N/A	77	WSC 1982
1947-1948	N/A	45	WSC 1982
1948-1949	N/A	48	WSC 1982
1949-1950	N/A	49	WSC 1982
1950-1952	N/A	38	WSC 1982
1952-1954	N/A	46	WSC 1982
1954-1956	N/A	36	WSC 1982
1956-1958	N/A	40	WSC 1982
1958-1960	N/A	33	WSC 1982
1960-1962	N/A	41	WSC 1982
1962-1964	N/A	27	WSC 1982
1964-1966	N/A	13	WSC 1982
1966-1968	N/A	20	WSC 1982
1968-1970	N/A	8	WSC 1982
1970-1972	N/A	2	WSC 1982
1972-1974	N/A	21	WSC 1982
1974-1976	N/A	17	WSC 1982
1976-1978	N/A	31	WSC 1982
1978-1980	N/A	35	WSC 1982
1979-1992	690	53	this study

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TEST TARGET (QA-3)

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