

**FLUVIAL RESPONSE TO LATE HOLOCENE GLACIER
FLUCTUATIONS IN THE NOSTETUKO RIVER VALLEY,
SOUTHERN COAST MOUNTAINS, BRITISH COLUMBIA**

by

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ABSTRACT

Late Holocene glacier fluctuations and changes in bed elevation are recorded in the alluvial fill of the west fork Nostetuko River valley, southern Coast Mountains, British Columbia. Valley-wide aggradation coincides with periods of local and regional glacier advance on centennial timescales. Peat layers containing *in situ* tree roots and stumps formed during periods of floodplain stability that coincide with intervals when glaciers were restricted. Radiocarbon ages on roots, tree stems, and woody plant detritus in the peat layers record five major phases of Holocene glacier advance, the most recent at the culmination of the Little Ice Age. A high-resolution record of Little Ice Age glacier fluctuations was derived by cross-dating ring-width series of fossil trees in the peat layers with a previously established master ring-width chronology and by constraining floating ring-width chronologies with radiocarbon ages.

For Grandpa

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TABLE OF CONTENTS

Approval	ii
Abstract.....	iii
Dedication	iv
Acknowledgements	v
Table of Contents	vi
List of Figures.....	viii
List of Tables	x
Chapter 1 Introduction.....	1
1.1 Introduction.....	1
1.2 Objectives.....	6
1.3 Thesis format.....	6
1.4 Figures.....	7
Chapter 2 Fluvial response to Holocene glacier fluctuations along the west fork of Nostetuko River	8
2.1 Abstract	8
2.2 Introduction.....	9
2.3 Study area.....	11
2.4 Methods.....	12
2.5 Results	13
2.5.1 Geomorphology	13
2.5.2 Sedimentology	16
2.5.3 Stratigraphy.....	18
2.5.4 Chronology	20
2.6 Discussion	21
2.6.1 Peats	21
2.6.2 Relation between stratigraphy and glacier fluctuations	22
2.6.3 Completeness and sensitivity of record	25
2.7 Conclusion	27
2.8 Acknowledgements	27
2.9 Figures.....	28
2.10 Tables	39
Chapter 3 Constraining times of Little Ice Age glacier fluctuations in the west fork of Nostetuko River valley using dendrochronology	40
3.1 Abstract	40

3.2	Introduction.....	41
3.3	Study area.....	43
3.4	Methods.....	44
3.4.1	Sampling	44
3.4.2	Laboratory analysis.....	44
3.5	Results	46
3.5.1	Internal tree-ring correlations	46
3.5.2	Stratigraphic correlations and ring-width chronologies.....	47
3.5.3	Ages of peat layers and aggradation	48
3.6	Discussion	50
3.6.1	Little Ice Age chronology	50
3.6.2	Possible sources of error	54
3.7	Conclusion	55
3.8	Acknowledgements	56
3.9	Figures.....	57
3.10	Tables	63
Chapter 4	Conclusions.....	64
4.1	Overview	64
4.2	Future work	65
	References.....	67
	Appendices.....	77
	Appendix 1. Locations of studied sections.....	77
	Appendix 2. Lithostratigraphy of studied sections.....	78

LIST OF FIGURES

Figure 1.1	The paraglacial sedimentation cycle (from Church and Ryder, 1972). Sediment yields are highest during and immediately after deglaciation. Geological Society of America Bulletin, 1972, by permission.	7
Figure 2.1	Location of the study area in the southern Coast Mountains of British Columbia (modified from BMGS data; reproduced with permission of the Province of British Columbia).	28
Figure 2.2	Aerial photomosaic of the valley of the west fork of Nostetuko River, showing locations of studied sections. The aerial photographs were flown in 1998, one year after the outburst flood.	29
Figure 2.3	Photomosaic of the eroded distal face of a large moraine fan complex located 1.5 km north of Queen Bess Lake. The moraine was built during late Holocene time.	30
Figure 2.4	Four terraces (T-1, T-2, T-3, and T-4) inset into a gravel fan below the upper bedrock gorge. An <i>in situ</i> root in the scarp below T-2 yielded a radiocarbon age of 370 ± 50 ¹⁴ C yr BP (Kershaw, 2002).	31
Figure 2.5	Scarp between terraces T-2 and T-3, showing a soil overlain and underlain by cobble-boulder gravel. Dashed lines bracket the dated soil.	32
Figure 2.6	Lithostratigraphy of late Holocene sediments exposed at seven sites in the study area. Peat layers are shown in black. See Figure 2-2 and Appendix 1 for site locations and Table 2-1 for details on radiocarbon ages.	33
Figure 2.7	The lower part of the section at site 4. A basal gravel unit is sharply overlain by silt, sand, and peat beds. The two arrowed stumps are rooted in two peat beds separated by about 10 cm. Dashed lines delineate the peat beds. Radiocarbon ages of 520 ± 50 and 940 ± 50 ¹⁴ C yr BP (Table 2-1) were obtained on the outer rings of the two stumps.	34
Figure 2.8	Upper 2 m of the section at site 5, showing a prominent peat layer, an <i>in situ</i> stump, and an oxidized horizon (directly above the trowel). The heavy dashed lines delineate the peat layer, and the dotted lines mark contacts between sand beds. A radiocarbon age of 600 ± 60 ¹⁴ C yr BP was obtained on a root near the base of this section (not shown in the photograph).	35
Figure 2.9	Upper 2 m of the section at site 9, showing peat layers, laminated fine-grained sediments, and a stump in growth position. Dashed	

	lines delineate peat layers. A radiocarbon age of 470 ± 60 ^{14}C yr BP was obtained on an <i>in situ</i> root (not shown in the photograph).	36
Figure 2.10	Examples of sections documented in this study.	37
Figure 2.11	Plot of radiocarbon ages obtained for this study and their relation to independently dated Holocene glacier advances in western Canada. Cluster analysis placed the radiocarbon ages into five groups. The youngest group consists of two sub-groups (5A and 5B).	38
Figure 3.1	Location of the study area (modified from BMGS data; reproduced with permission of the Province of British Columbia).	57
Figure 3.2	Aerial photomosaic of the valley of the west fork of Nostetuko River, showing locations of sampled trees and radiocarbon ages.	58
Figure 3.3	Upper 1.5 m of the section from which sample KW39 was collected. Dashed lines delineate peat layers. Sample KW39 is from a tree rooted in peat 2 (see text).	59
Figure 3.4	Photographic mosaic of the eroded distal face of a large moraine fan complex, 1.5 km north of Queen Bess Lake. The moraine was built during middle and late Holocene time. Inset photographs show logs associated with a paleosol within the fan.	60
Figure 3.5	Floating tree-ring chronologies for peat layers in the west fork of Nostetuko River valley. Lengths of correlated tree-ring series, inter-series correlation values, and radiocarbon ages are shown for each peat layer.	61
Figure 3.6	Schematic diagram showing chronological framework of late Holocene peat and clastic units in the west fork of Nostetuko River valley. Unit thicknesses are representative of those at measured sections.	62

LIST OF TABLES

Table 2.1	Radiocarbon ages from the west fork of Nostetuko River valley.	39
Table 3.1	Summary of tree-ring statistics for samples from the study area.	63
Table 3.2	Radiocarbon ages bearing on floodplain stability.	63

CHAPTER 1

INTRODUCTION

1.1 Introduction

Glaciofluvial systems are highly dynamic and complicated, with characteristics that interact with one another and vary both in time and space (Fenn and Gurnell, 1987). They are controlled by internal (autocyclic) and external (allocyclic) factors that operate on timescales ranging from minutes to millions of years and that differ in relative and absolute importance depending on spatial scale (Beerbower, 1964; Vandenberghe, 1995, 2002). Autocyclic controls typically operate on shorter timescales and include channel migration and avulsion, bed scour and fill, and bank erosion and accretion (Beerbower, 1964; Schumm et al., 1987). Allocyclic controls occur on longer timescales and include sea level and climate change, and tectonism (Beerbower, 1964; Schumm et al., 1987). Autocyclic processes are commonly triggered or influenced by allocyclic factors (Holbrook et al., 2003, 2006). Autocyclic and allocyclic controls are thus not independent of one another and rarely, if ever, operate in mutual exclusion (Holbrook et al., 2003). For example, a river channel may change in response to changes in the calibre and amount of transported sediment, bank stability, or channel geometry, but these factors are also influenced by the precipitation regime and other aspects of climate (Schumm et al., 1987; Holbrook et al., 2003).

Rivers will aggrade or incise as they adjust to a new longitudinal profile resulting from tectonic uplift or subsidence, sea-level change, or other external forces (Gilbert,

1877; Mackin, 1948; Bull, 1991). Among the many factors that can alter local river bed elevation are landslides, a change in sediment delivery to streams, and a change in precipitation regime. The influence of sea level on river bed elevation diminishes upstream, so that base level in areas distant from the sea is dominated by allocyclic, climatic, and tectonic forcings (Shanley and McCabe, 1994; Törnqvist, 1998; Blum and Törnqvist, 2000; Holbrook et al., 2006).

Climate influences fluvial and glaciofluvial systems on different timescales. On timescales of thousands to hundreds of thousands of years, alpine areas have experienced glacial, periglacial, and nonglacial conditions, with attendant changes in base level, vegetation, sediment availability, and water and sediment discharge (Church and Ryder, 1972; Fenn and Gurnell, 1987; Church and Slaymaker, 1989; Knox, 1995; Marren, 2005). On decadal and centennial timescales, alpine glaciers have advanced and retreated, affecting the flux of sediment and water through the glaciofluvial system (Karlén, 1976; Maizels, 1979, 1983; Leonard, 1986, 1997; Karlén and Matthews, 1992; Ballantyne and Benn, 1994, 1996; Ashmore and Church, 2001; Ballantyne, 2002; Church, 2002). On annual and shorter timescales, glaciofluvial streams are characterized by large fluctuations in the discharge of water and sediment, which are derived primarily from the glacierized part of the basin (Fenn and Gurnell, 1987). Meltwater discharge varies diurnally and seasonally due to changes in air temperature, insolation, and other factors. In this context, the size and location of glaciers in a watershed are important, as they affect the magnitude of the response and determine whether or not there are significant lags between climatic forcings and water and sediment fluxes (Fenn and Gurnell, 1987).

Water and sediment discharge in alpine glaciofluvial systems is thus strongly influenced by changes in the extent of glaciers on decadal to millennial timescales

(Gregory, 1987; Orwin and Smart, 2004). Primary glacial drift in glacier forefields, colluvial deposits on slopes, and channel and overbank deposits on floodplains ensure that most glaciofluvial systems are not supply-limited and that climate change can drive large changes in sediment flux to alpine streams. These streams may aggrade or incise in response to changes in sediment and water discharge linked to glacier advance or retreat (Fenn and Gurnell, 1987).

Changes in sediment flux to streams is greatest during and immediately after continental deglaciation (Fig. 1.1; Church and Ryder, 1972; Church and Slaymaker, 1989; Ashmore and Church, 2001; Ballantyne, 2002; Church, 2002). The term “paraglaciatiion” was first defined by Church and Ryder (1972, p. 3059) to describe:

“...nonglacial processes that are directly conditioned by glaciation. It [paraglaciatiion] refers both to proglacial processes, and to those occurring around and within the margins of a former glacier that are the direct result of the earlier presence of the ice. It is specifically contrasted with the term “periglacial”, which does not imply the necessity of glacial events occurring.”

In its original formulation, paraglacial sedimentation is controlled largely by sediment availability. Continental deglaciation exposed vast areas of unstable, poorly vegetated sediment, facilitating the transfer of large amounts of sediment to streams and leading to aggradation in valleys and complex changes in channel planform (Church 1983; Brooks, 1994; Ashmore and Church, 2001). Sediment exhaustion may take up to many thousands of years in large basins before the system returns to a more normal regime of sediment yield dominated by debris produced by primary weathering (Church and Slaymaker, 1989; Ashmore, 1993; Church et al., 1999). Fluctuations of alpine glaciers on shorter timescales can prolong the primary paraglacial sediment pulse by mobilizing previously sequestered sediment (Brooks, 1994). Glaciers erode and entrain the sediments they

override and ultimately deliver much of this material to ice margins where it becomes available to glaciofluvial streams (Fenn and Gurnell, 1987). Equally important are variations in meltwater discharge during intervals of glacier advance and retreat, as these strongly influence the temporal pattern of sediment transport and deposition (Maizels, 1979, 1983, Fenn and Gurnell, 1987).

Maizels (1979) introduced a conceptual model linking proglacial aggradation and incision to glacier mass balance and water and sediment supply. According to the model, meltwater discharge is relatively low at times of both restricted and extensive ice cover and is relatively high at times of intermediate ice cover. Thus, in valleys where sediment supply is not a limiting factor, aggradation is most likely during periods when glaciers advance or retreat from maximum positions. Sediment delivery to proglacial lakes commonly increases during and immediately following glacier advances (Karlén, 1976; Souch, 1994; Leonard, 1997; Leonard and Reasoner, 1999; Nesje et al., 2000). During glacier advance, initial fluvial incision due to the increased competence of proglacial meltwater streams is quickly followed by aggradation as sediment supply increases (Maizels, 1979). Sediment stored within and beneath glaciers is delivered at an increasing rate to streams as glaciers advance (Karlén, 1976; Maizels, 1979, 1983; Leonard, 1986, 1997; Karlén and Matthews, 1992; Souch, 1994). The area of subglacial erosion also expands during a glacier advance, introducing more sediment into river valleys than at times when glaciers are more retracted (Clague, 1986, 2000).

Sediment yield in the Canadian Cordillera remains high more than 10,000 years after the disappearance of the Cordilleran ice sheet, suggesting that landscape adjustment to Pleistocene glaciation may span tens of thousands of years in large basins (Church and Ryder, 1972; Church and Slaymaker, 1989; Ashmore, 1993; Church et al., 1999). The

timescale of paraglacial sedimentation, however, is proportional to basin size and sediment availability (Church and Slaymaker, 1989; Ballantyne, 2002; Church, 2002), and the response of fluvial systems to the relatively small climate changes of the Holocene remains unclear.

Dendrochronology, lichenometry, and radiocarbon dating of *in situ* and detrital wood from glacier forefields in the Canadian Cordillera have provided a centennial-scale chronology of middle to late Holocene glacier fluctuations in the region. Four major phases of glacier expansion are currently recognized from this evidence: (1) the Little Ice Age (late Neoglacial), which began as early as about AD 1200 and culminated between the late AD 1600s and late AD 1800s (Osborn, 1986; Ryder and Thomson, 1986; Osborn and Luckman, 1988; Desloges and Ryder, 1990; Luckman, 1993, 2000; Smith et al., 1995; Smith and Desloges, 2000; Koch et al., 2004a); (2) the “first millennium A.D. advance,” which culminated about 1500 years ago and has recently been documented throughout western North America (Reyes et al., 2006); (3) the Tiedemann or Peyto Advance (middle Neoglacial), which occurred between 3500 and 1800 years ago (Ryder and Thomson, 1986; Clague and Mathews, 1992; Luckman et al., 1993; Larocque and Smith, 2003; Clague et al., 2004; Koch et al., 2004a; Reyes and Clague, 2004); and (4) the Garibaldi Advance between 7300 and 5800 years ago (Ryder and Thomson, 1986; Koch et al., 2003, 2004a, 2004b; Smith, 2003). The term “glacier advance” has been used in much of the previous literature on Holocene glaciation in western Canada to describe periods of glacier expansion. However, “advance” is a poor word to apply to a generally cooler period with highly complex glacier fluctuations. The Tiedemann and Garibaldi events are not monotonic advances, but rather periods, perhaps 1000 years long, characterized by repeated advance and retreat of glaciers.

1.2 Objectives

This thesis examines the relation between downstream river response and glacier fluctuations during the middle and late Holocene in the west fork of the Nostetuko River valley, about 220 km north of Vancouver in the southern Coast Mountains of British Columbia. Evidence of Holocene glacier fluctuations is preserved in fluvial sediments and landforms of other rivers draining the Coast Mountains (Desloges and Church, 1987; Gottesfeld and Johnson-Gottesfeld, 1990). Nostetuko valley was chosen for this study because reconnaissance studies suggested that Holocene glacier activity influenced sediment supply to Nostetuko River, at times triggering valley-wide aggradation (Kershaw, 2002; Kershaw et al., 2004). The objectives of the study are to document the fluvial response of the west fork of Nostetuko River to upvalley glacier fluctuations during the middle and late Holocene and to date periods of floodplain stability and aggradation.

1.3 Thesis format

The thesis includes two journal-style papers. Both papers address the downvalley response of the fluvial system to glacier fluctuations, but they differ in methodology and timescale. The first paper (Chapter Two) reconstructs the aggradation history of the west fork of Nostetuko River during the last half of the Holocene. Stratigraphic and geochronological data provide evidence for five overbank aggradation phases that coincide with independently dated local and regional glacier advances. The second paper (Chapter Three) focuses on the aggradation history of the west fork during the Little Ice Age, the most recent of the phases described in Chapter Two. Stratigraphic and dendrochronological data and radiocarbon ages provide evidence for periods of Little Ice Age floodplain stability and forest growth coinciding with intervals of warmer climate.

These intervals were preceded and followed by periods of aggradation. Chapter Four summarizes the main conclusions of the research in the context of the proglacial fluvial sediment archive.

1.4 Figures

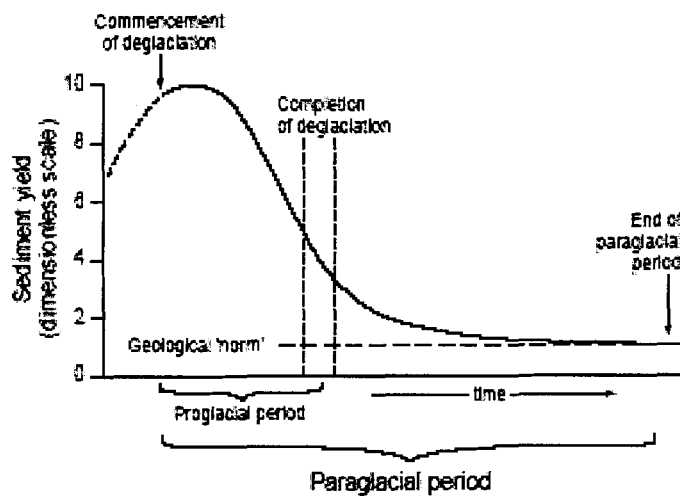


Figure 1.1 The paraglacial sedimentation cycle (from Church and Ryder, 1972). Sediment yields are highest during and immediately after deglaciation. Geological Society of America Bulletin, 1972, by permission.

CHAPTER 2

FLUVIAL RESPONSE TO HOLOCENE GLACIER FLUCTUATIONS ALONG THE WEST FORK OF NOSTETUKO RIVER

2.1 Abstract

Late Holocene glacier fluctuations are recorded in the upper part of the alluvial fill in the Nostetuko River valley, southern Coast Mountains, British Columbia. Glacier advances on decadal to centennial timescales triggered valley-wide aggradation and changes in river planform. Periods when glaciers were restricted in extent coincide with periods of incision of the valley fill and floodplain stability. As many as ten overbank aggradation units are separated by peat layers containing tree roots and stems in growth position. Twenty five radiocarbon ages on roots, tree stems, and woody plant detritus in the peat layers constrain intervals of aggradation. The oldest phase of aggradation occurred about 5810 ± 70 ^{14}C yr ago and coincides with the Garibaldi Advance documented elsewhere in the southern Coast Mountains. A second phase of aggradation, recorded by several units of clastic sediment, dates to 2450 ± 70 ^{14}C yr BP, near the peak of the middle Neoglacial Tiedemann Advance. The third phase occurred about 1280 ± 70 ^{14}C yr BP, during or shortly after a glacier advance that recently has been documented in British Columbia and Alaska. The most recent and greatest phase of aggradation began about 800 years ago and continued until the late 1800s. It coincides with the Little Ice Age, the time when glaciers in the Nostetuko River basin and elsewhere in the southern Coast Mountains achieved their maximum Holocene extent. The stratigraphic and geochronologic evidence presented in this paper suggests that relatively minor changes in

climate (1-2°C) can have significant impacts on local base level and channel planform in glacierized mountains. Glaciofluvial sediments, like glacial landforms and deposits, are sensitive indicators of climate change.

2.2 Introduction

Neoglaciation is the period of generally cool climate that followed the warmest part of the Holocene. It was marked by recurrent expansion of alpine glaciers around the world (Porter and Denton, 1967). The Little Ice Age, which spans much of the last millennium, is the most recent phase of Neoglaciation (Grove, 1988; Luckman, 2000), and most glaciers in the Northern Hemisphere reached their maximum Holocene extent at this time. Much of the evidence of earlier Holocene advances was destroyed by glacier overriding during the Little Ice Age.

Evidence for Neoglacial advances and climate variability in western Canada has come largely from studies of glacial deposits and landforms (Osborn, 1986; Ryder and Thomson, 1986; Osborn and Luckman, 1988; Reyes and Clague, 2004) and from paleoecological studies (Hebda, 1995, Clague and Mathewes, 1996; Pellatt and Mathewes, 1997; Clague et al., 2004). Moraine sequences have been indirectly dated using tephrochronology (Luckman and Osborn, 1979), lichenometry (Larocque and Smith, 2003), lateral moraine stratigraphy (Desloges and Ryder, 1990; Osborn et al., 2001; Reyes and Clague, 2004), and dating of *in situ* glacially overridden and sheared tree stumps (Ryder and Thomson, 1986; Luckman, 1993, 2000; Luckman et al., 1993; Wiles et al., 1999; Osborn et al., 2001; Reyes et al., 2006). However, the fragmentary nature of the moraine record and poor dating control have precluded complete reconstruction of the Neoglacial history of glaciers in western Canada.

An alternative method of reconstructing glacier fluctuations is to study downvalley lake sediments. Lake sediment cores may provide records of glacier activity that are more continuous and more easily dated than the moraine record (Karlén, 1976; Leonard, 1986, 1997; Karlén and Matthews, 1992; Souch, 1994; Leonard and Reasoner, 1999; Menounos, 2002; Nesje et al., 2000; Menounos et al., 2004). Studies of sediment cores from proglacial lakes in the Coast and Rocky Mountains suggest that, at decadal and centennial timescales, sedimentation increases during and following glacier advances, and is low during periods of reduced ice extent (Leonard, 1986, 1997; Menounos, 2002; Menounos et al., 2004). Subglacial erosion and production of clastic sediment increase during glacier advance (Karlén, 1976; Clague 1986, 2000), and large volumes of silt and sand may be delivered to streams by advancing glaciers. Glaciers may also override and erode older sediments, further increasing the sediment flux to streams. Sediment is delivered to mountain rivers, not only by glaciers in headwater areas, but also by streams and landslides from steep valley walls throughout the watershed (Fenn and Gurnell, 1987). Considerable doubt, however, still exists about the relation between sediment production and glacier extent (Hicks et al., 1990; Harbor and Warburton, 1993; Hallet et al., 1996).

Glacier fluctuations are also recorded in the sediments and paleo-floodplains of rivers draining glacierized mountains. Rivers in British Columbia have responded to changes in climate over timescales ranging from several years to millennia by aggrading and incising their channels and by altering channel planforms (Church and Ryder, 1972; Church, 1983; Ryder and Church, 1986; Desloges and Church, 1987; Gottesfeld and Johnson-Gottesfeld, 1990; Ashmore and Church, 2001; Clague et al., 2003). For example, Morice River in the central Coast Mountains had a different planform during

the Little Ice Age than today, with the active floodplain occupying up to twice its present area (Gottesfeld and Johnson-Gottesfeld, 1990). Dendrochronological dating of log jams and flood scars along Morice River has provided a chronology of flood events and channel changes, indicating that the active part of the valley floor expanded during the Little Ice Age due to increased sediment and water discharge (Gottesfeld and Johnson-Gottesfeld, 1990). Church (1983) and Desloges and Church (1987) documented an increase in sediment delivery and deposition in the Bella Coola River valley in the central Coast Mountains in the 19th century, coincident with glacier expansion late during the Little Ice Age.

These important studies aside, the proglacial fluvial archive is a largely unexploited source of information on upvalley glacier fluctuations. Some alluvial stratigraphies provide records of landscape response to environmental change. The impact of changing climate on proglacial fluvial systems can be recorded in floodplain sediments (Brackenridge, 1988), and these sediments can be used in conjunction with other proxy records, such as pollen and lake sediment data, to better interpret paleoenvironmental change (Rumsby and Macklin, 1996; Brown, 1998; Straffin and Blum, 2002) This paper documents the response of a river in the southern Coast Mountains to changes in sediment supply during the last half of the Holocene. I show that sediment supply is intimately linked to upvalley glacier fluctuations.

2.3 Study area

My study area is the valley of the west fork of Nostetuko River in the southern Coast Mountains of British Columbia, 220 km north of Vancouver, British Columbia (Fig. 2.1). The west fork flows 11 km north and east from valley glaciers at the edge of

Homathko Ice Field to the main stem of Nostetuko River (Fig. 2.2). A major tributary of the west fork flows from Queen Bess Lake, a moraine-dammed lake that partially drained during an outburst flood in August 1997 (Kershaw, 2002; Kershaw et al., 2004).

Queen Bess Lake is impounded by a large composite moraine produced by at least two Neoglacial advances of Diadem Glacier (Kershaw, 2002). The lake formed behind the composite moraine during glacier retreat in the late 1800s and early 1900s. In 1997, a large ice avalanche fell from the toe of Diadem Glacier into Queen Bess Lake, generating displacement waves that overtopped and incised the moraine. The resulting flood eroded sediments in the valley below the dam, causing aggradation upstream and downstream of channel constrictions (Fig. 2.2). It also created exposures of the upper part of the valley fill, which allowed me to conduct this study.

2.4 Methods

The aggradation history of the west fork of Nostetuko River was determined through stratigraphic, sedimentological, and geochronological analyses of key sections. Fieldwork was conducted during the summer of 2004. Detailed topographic maps (1:5000 scale), constructed from aerial photographs flown in 1998, one year after the outburst flood, were used to map deposits and landforms. Locations of sections, terraces, trimmed colluvial fans, and tree stumps exhumed by river incision were located (± 10 m) using a hand-held GPS unit and cross-referenced to the topographic maps. These data were subsequently entered into a Geographic Information System (GIS).

Detailed sedimentologic and stratigraphic logs were made of exposed valley-fill sediments at seven sites (Fig. 2.2; Appendices 1 and 2). Sections were logged using a metric tape and a barometric altimeter (elevation accuracy = ± 1 m). The altimeter was

used to measure relative elevations of the tops and bases of sections. The altimeter readings at each section were converted to absolute elevations using the site elevation on the topographic map as a datum; the inferred accuracy of the absolute elevations is ± 2 m. The level of the west fork of Nostetuko River did not vary during the period of field work and served as a local datum for inferring the amount of valley-floor aggradation in the past. Recorded field data include grain size (field estimates), sorting, sedimentary structures, Munsell colour, unit thickness, the nature of unit contacts, and fossils.

Samples of *in situ* fossil tree stems and roots, and detrital plant fossils were collected from peat layers and rooting horizons and submitted to Beta Analytic for conventional (radiometric) ^{14}C analysis. Radiocarbon ages were calibrated using the software OxCal v.3.9 (Bronk Ramsey, 1995, 2001), which is based on the decadal data of Stuiver et al. (1998). The radiocarbon ages provide chronological control on periods of stability and aggradation in the valley.

2.5 Results

2.5.1 Geomorphology

The west fork of Nostetuko River flows through rugged terrain with local relief of up to 2000 m. Alluvial reaches are separated by short rock canyons located approximately 1 km and 6 km north of Queen Bess Lake (Fig. 2.2). The canyons control valley gradients and local base level. The average gradient of the west fork of the river is 3.8° , but it ranges from a maximum of 14° in the upper rock canyon to a minimum of about 0.5° along broad alluvial reaches (Kershaw, 2002).

The modern floodplain is covered by deposits of the 1997 outburst flood. The sediments are poorly sorted, cobble-boulder gravel with a matrix of fine gravel and sand.

Kershaw (2002) and Kershaw et al. (2004) noted two units in some sections – a lower, finer gravel deposited during the early, overtopping phase of the outburst flood, and an upper boulder gravel deposited during the breach phase of the flood. Sand and silty sand were deposited locally along the margins of the flood path and in areas upstream of constrictions, where hydraulic ponding occurred during the outburst flood.

The flood altered the planform of the west fork of Nostetuko River. Prior to the flood, the west fork had a single dominant channel with local low-gradient, multi-channel reaches. The flood changed the position of the main channel and temporarily imposed a braided planform on the active floodplain. Since the flood, the west fork has incised its flood deposits and partly reestablished its pre-flood form. The modern channel is dominantly straight, but it is multi-threaded along broad, low-gradient reaches below canyons. The modern floodplain is sparsely vegetated, but trees are beginning to recolonize areas flooded in 1997.

Lateral migration of the west fork is constrained by the steep valley slopes and colluvial fans and aprons draping the valley walls. The fans and aprons are eroded during major floods and thus are an important source of sediment to Nostetuko River. A large moraine fan complex is located on the west side of the valley 1.5 km downvalley of Queen Bess Lake (Fig. 2.3). The fan complex comprises bouldery colluvium deposited on the distal side of a terminal moraine constructed during middle and late Neoglacial time. A branch recovered from the base of this fan, just above bedrock, yielded an age of 2790 ± 60 ^{14}C yr BP (Table 2.1).

Four terraces are inset into a large gravel fan directly below the upper bedrock canyon. The upper two terraces (T-1 and T-2; Fig. 2.4) support sharp-crested boulder levees composed of crudely bedded boulder and cobble gravel. The levees are bordered

by better sorted cobble gravel associated with relict channels. The uppermost terrace (T-1) is partly vegetated. T-1 and T-2 support lichens (*Rhizocarpon* spp.) up to 3.1 cm in diameter. The lower two terraces (T-3 and T-4) were swept by the 1997 outburst flood and thus lack lichens.

Aggradation of the outwash fan to the T-1 level likely occurred during the Little Ice Age when the outer, sharp-crested terminal moraine at the east end of Queen Bess Lake was constructed by Diadem Glacier (Kershaw, 2002). T-2 is inset into, and therefore younger than, T-1. It probably dates to the late 1800s or early 1900s (Kershaw, 2002). T-3 records the upper limit of aggradation of the 1997 flood, and T-4 formed during the waning stage of the flood or soon thereafter.

A 30-cm-thick, buried organic soil exposed in the scarp of T-2 contains *in situ* stumps and abundant detrital plant material (Fig. 2.5). Kershaw (2002) reported a radiocarbon age of 370 ± 50 ^{14}C yr BP (AD 1390-1590; TO-8923) on a root in the soil. The soil is underlain and overlain by cobble-boulder gravel. Kershaw (2002) proposed that T-2 formed sometime after AD1500 on the basis of the stratigraphy at this site, upvalley radiocarbon ages, and lichen measurements on T-2 and Diadem Glacier moraines. The buried soil rests on a floodplain that records a bed elevation 8 m higher than present, probably shortly before AD 1500.

Terraces are uncommon along the west fork, but notable exceptions occur 3.5 km and 7 km north of Queen Bess Lake. The terraces at these two sites are 1-2 m above present river level, support forest at least 100 years old, and were not inundated by the 1997 flood. They record a higher bed elevation prior to the 20th century and may correlate with terrace T-2 farther upvalley.

2.5.2 Sedimentology

The upper, exposed part of the valley fill varies both laterally and vertically. Lithostratigraphic logging of sections revealed four dominant lithofacies, which are described and interpreted below. The description excludes the capping 1997 outburst flood deposits, because they have been described elsewhere (Kershaw, 2002; Kershaw et al., 2004).

2.5.2.1 Gravel facies

Gravel units are present at six of the seven logged sections (Fig. 2.6; Appendix 2). They consist mainly of pebbles and cobbles, and are clast-supported, horizontally bedded, and locally imbricated. Clasts are subangular to well rounded and locally iron-stained. The matrix comprises sand and granules. Gravel occurs at the base of four sections (Fig. 2.7) and as discontinuous lenses up to 15 cm thick within finer grained sediments at six sections (Fig.2.6).

The gravel facies are interpreted to be channel deposits of a braided or wandering river (Williams and Rust, 1969; Rust, 1972, 1984; Miall, 1977, 1978; Rust and Koster, 1984). Horizontal stratification and clast imbrication suggest deposition on near-horizontal surfaces such as braid bars, medial and lateral bar complexes, and channel floors. These environments are common in braided and wandering gravel-bed rivers (Bluck, 1979; Church, 1983; Desloges and Church, 1987; Brierley, 1996; Sambrook Smith, 2000; Lewin et al., 2005).

2.5.2.2 Sand facies

Massive and stratified sand is the dominant sediment of the uppermost part of the valley fill (Figs. 2.6 and 2.8). Tabular and lenticular beds of mottled and oxidized,

massive, well sorted, very fine to medium sand occur at all sites. Beds of planar cross-stratified and ripple-stratified, medium to coarse sand are also common. Horizontally laminated, fine to very fine sand is interstratified with the coarser sand. Laminae are typically flat to undulating. Small-scale, trough cross-stratified, fine to medium sand occurs in the uppermost 0.5 m of the sequence at three sites (Fig. 2.6). Sand beds commonly have sharp lower contacts and sharp to gradational upper contacts (Fig. 2.8).

I interpret the sand facies to be channel, bar, and levee deposits. Rippled and horizontally bedded sand may have been deposited under a range of flow conditions, from lower flow regime in back channels to upper flow regime in the main active channels (Desloges and Church, 1987). Discontinuous, lenticular beds of structureless to cross-stratified sand were deposited in channels immediately before they were abandoned. Trough cross-stratified sand records in-channel migration of large lunate ripples and bars. These bedforms are typically transitory, as variable flow depths and velocities impede preservation (Desloges and Church, 1987). The abundance of this facies in the upper part of the valley fill is consistent with deposition by an aggrading braided river (Rust, 1984).

2.5.2.3 Fine facies

The fine facies consists of massive, bedded, and laminated very fine sand, silt, and minor clay (Fig. 2.9). Strata range from laminae a few millimetres thick to beds up to 17 cm thick. Fine sediments dominate sections of valley fill up to 1 m thick, but isolated, lenticular strata are also common. Sediments are mainly olive grey, but locally are oxidized and mottled. Interbeds of coarser sand, fibrous peat, and plant detritus occur within the fine facies.

The laminated and bedded fine sediments indicate an overbank depositional environment (Wolman and Leopold, 1957; Brierley, 1991, 1996; Nanson and Croke, 1992). Sediments commonly fine upwards from an erosive base, reflecting scour during the rising stage of a flood, followed by deposition during the waning stage (Brierley, 1996). Massive, very fine sand and silt record either rapid deposition during floods or bioturbation (Collinson, 1996). Fine sediment occurs locally within sandur and valley train deposits, particularly in infilled abandoned channels (Miall, 1992).

2.5.2.4 Organic facies

Beds and laminae of brown peat and silty peat are abundant at all sites (Figs. 2.6, 2.8, and 2.9). The strata range from a few millimetres to 18 cm thick. Thicker layers comprise woody and herbaceous peat mats with tree stumps in growth position (Figs. 2.8 and 2.9). Roots of herbaceous plants and trees extend downward from the organic horizons into underlying silt and sand. Contacts with overlying sediment are typically sharp, whereas basal contacts are either sharp or gradational.

The organic facies records peat accumulation on poorly drained, stable floodplain surfaces (Wright, 1986; Brierley, 1996). River bed elevation was either stable or dropping at times of peat deposition and soil development. Some peat layers are unconformably overlain by coarse sand or gravel, indicating that they were eroded and buried during aggradation following stable floodplain phases (Desloges and Church, 1987; Gottesfeld and Johnson-Gottesfeld, 1990).

2.5.3 Stratigraphy

Correlation of strata in high-energy proglacial fluvial systems based solely on lithofacies is difficult. Thicknesses and lithologies of units differ markedly over short

distances, and peaty soils present at one site may be missing at others due to erosion. Nevertheless, stratigraphic relations and age constraints provided by radiocarbon dating (see section 2.5.4) permit correlation of some peat layers and documentation of periods of floodplain stability.

Sequences of gravel and sand facies alternate with sequences dominated by fine sediment at all of the studied sections in the west fork Nostetuko valley (Fig. 2.6). Coarse gravel is exposed at the base of four sections in sharp contact with overlying fine sand. Gravel also overlies massive and laminated silt higher in the sequence at sites 7 and 9. Contacts between fine-grained sediments and overlying coarser sand and gravel are sharp and erosive.

Coarse sand and granule-rich sand dominate the uppermost 1.5 m of the valley-fill sequence at all sites. These sediments overlie massive and laminated silt with layers of peat and plant detritus. A 58-cm-thick unit of clast-supported diamicton at site 5 was deposited by a debris flow from the nearby valley slope.

All sections have multiple peat layers that record episodic floodplain stability. Units of sand and silt with abundant plant matter alternate with the peat layers. Fine-grained beds commonly have gradational contacts with overlying peat layers, whereas the latter are typically sharply overlain by silt or sand. Dark brown, herbaceous peat mats at sites 6 and 9 contain discontinuous lenses of massive fine sand and silt, and may correlate (Fig. 2.10). The herbaceous peat mat at site 6 correlates with a peat bed 2.9 m below the surface at site 7 (Fig. 2.10). Two thin, dark brown peat layers at sites 3 and 4 occur at depths of 2.3 and 2.1 m, respectively. At both sites, they are separated by about 2 cm of fine sand and silt. Peat layers just above the basal gravel units at these sites and also at site 7 are correlated on the basis of radiocarbon ages obtained on *in situ* roots.

Radiocarbon ages on tree stumps allow correlation of a thick, herbaceous peat at site 10 with the submerged peat bed at site 9.

The vertical succession of sediments, with numerous erosion surfaces and abrupt changes in facies, indicates frequent changes in river stage and channel position (Miall, 1977, 1978). Laterally discontinuous units of sand and gravel within fine-grained sediment are typical of near-channel floodplain environments (Smith, 1983; Marren, 2005). Sequences of silt suggest periods of overbank deposition and floodplain accretion, whereas coarser sediment likely records deposition in channels.

2.5.4 Chronology

Twenty-five samples of roots, tree stems, and detrital wood were radiocarbon dated (Table 1.1). Samples were chosen to date peat beds that were correlated on the basis of lithostratigraphic relations. Nineteen of the 25 samples are outer rings of fossil trees in growth position in peat layers. Their ages are interpreted to be the time of death and, presumably, burial of the trees. The other six samples are fragile branches and twigs in peat beds that are unlikely to have been reworked from older sediments. The ages of these six samples, nevertheless, must be considered maxima for the age of the sediments from which they were collected.

The oldest radiocarbon age (5810 ± 70 ^{14}C yr BP) is from the outer 10 rings of a stump rooted in a peat layer below modern river level at the downvalley end of the study area. A branch at the base of the moraine fan complex below the upper bedrock canyon gave an age of 2790 ± 60 ^{14}C yr BP. The two youngest radiocarbon ages, 110 ± 60 ^{14}C yr BP and 130 ± 50 ^{14}C yr BP, are from silty and sandy peat layers within the uppermost 1.5 m of sediment at site 7.

Cluster analysis was performed on the suite of 25 radiocarbon ages to determine if the ages are randomly distributed or grouped. The analysis identified five groups of ages (Fig. 2.11). The youngest group was analyzed independently and yielded two sub-groups (5A and 5B in Fig. 2.11). The age groupings corroborate the provisional correlations of peat layers based on field observations and stratigraphic relations at measured sections.

The oldest 'group' is actually a single radiocarbon age of 5810 ± 70 ^{14}C yr BP, obtained from the outermost rings of a rooted stump at site 11. The second group comprises five radiocarbon ages at sites 1 and 10, centered on 2300 ^{14}C yr BP. The third group consists of three radiocarbon ages ranging from 1300 ± 70 to 1160 ± 50 ^{14}C yr BP at sites 9 and 10. The three samples that yielded these ages are growth-position fossils in a thick peat bed. The fourth group includes three radiocarbon ages ranging from 1030 ± 50 to 940 ± 50 ^{14}C yr BP. The fifth group comprises 11 radiocarbon ages ranging from 710 ± 60 to 110 ± 60 ^{14}C yr BP. The greater abundance of ages in this last group probably reflects the better preservation of the youngest sediments and thus some bias in sampling.

2.6 Discussion

2.6.1 Peats

The oldest stable floodplain recorded in this study is the surface at the north end of the study area, dated at 5810 ± 70 ^{14}C yr BP (4490-4800 BC). Bed elevation at this time was lower than today because the dated stump and others in the same area are rooted at least 1 m below present river level.

Radiocarbon ages from three thick peats record a substantial period of floodplain stability, followed by aggradation beginning about 2500 ^{14}C yr BP. The oldest radiocarbon age in this group (2790 ± 60 ^{14}C yr BP; 820-1100 BC) is from the base of the

moraine fan complex on the west side of the valley, 1.5 km north of Queen Bess Lake (Figs. 2.3 and 2.4). It is a maximum age for the beginning of moraine construction and correlates with the beginning of a significant period of aggradation downvalley (2490 ± 70 ^{14}C yr BP; 410-790 BC; site 10).

The third and fourth groups of radiocarbon ages record floodplain stability followed by rapid aggradation, respectively, about 1300 ^{14}C yr BP and 1000 ^{14}C yr BP. The ages are youngest at upvalley sites, which are nearer sediment sources, and oldest at downvalley, more distal sites.

The youngest group of radiocarbon ages is divided into two subgroups on stratigraphic and statistical grounds. The older subgroup (5A) represents at least two peat beds that range in age from 710 ± 60 to 470 ± 50 ^{14}C yr BP (AD 1220-1400 to AD 1310-1640) and occur primarily in fine-grained sediments. The younger subgroup of radiocarbon ages (5B) is derived from peat layers in coarse sandy sediments. The ages range from 370 ± 50 to 110 ± 60 ^{14}C yr BP (AD 1390-1590 to AD 1670-1960).

2.6.2 Relation between Stratigraphy and Glacier Fluctuations

The foundation of the paradigm of paraglacial sedimentation is that sediment yield increases during glacier retreat (Church and Ryder, 1972). Sediment flux is highest during retreat and decreases as sediment sources are exhausted or stabilized (Church and Ryder, 1972; Church and Slaymaker, 1989). High rates of sedimentation during glacier retreat are the result of greater availability of glacial sediment in newly deglaciated terrain. Thus, according to the paraglacial paradigm, aggradation should occur during periods of substantial glacier retreat. The paraglacial concept, however, was originally developed for regional-scale, ice-sheet deglaciation, and its applicability to small alpine

catchments with much smaller glaciers is uncertain (Fenn and Gurnell, 1987). Forefields of alpine glaciers, for example, may become stabilized within decades of exposure (Ballantyne and Benn, 1994, 1996; Harrison and Winchester, 1997; Curry, 1999; Orwin and Smart, 2004).

Many researchers have linked increased sediment yield and aggradation to periods of more extensive ice cover and glacier advances (Karlén, 1976; Leonard, 1986, 1997; Karlén and Matthews, 1992; Nesje et al., 2000; Menounos, 2002; Davies et al., 2003; Menounos et al., 2004). Increased glacial erosion and sediment production during glacier advance, coupled with climatically induced changes in discharge and sediment yield can cause rivers to aggrade their beds (Knighton, 1998). Sediment delivery to streams in the Coast Mountains, for example, increased during the Little Ice Age (Church, 1983; Gottesfeld and Johnson-Gottesfeld, 1990).

As discussed below, intervals of floodplain aggradation in the west fork of Nostetuko River valley coincide with previously documented Holocene glacier advances in the Coast and Rocky Mountains and parts of Alaska. My results thus imply that sediment delivery to the fluvial system increased at times of glacier expansion.

Evidence for middle Holocene advances in western Canada is sparse because landforms and associated sediments were overridden, eroded, and buried during later glacier expansion (Mathews, 1951; Luckman, 1986; Osborn, 1986; Ryder and Thomson, 1986; Ryder, 1987; Desloges and Ryder, 1990). However, evidence from a few glacier forefields and lakes shows that glaciers advanced 6000-5000 ^{14}C yr BP (the Garibaldi Advance; Ryder and Thomson, 1986; Koch et al., 2003, 2004a; Smith, 2003). Advances of this age are also recognized in interior and coastal Alaska (Calkin, 1988). The oldest

phase of fluvial aggradation in the west fork of Nostetuko valley coincides with this event.

The second period of aggradation in the west fork Nostetuko valley coincides with a middle Neoglacial expansion of glaciers, which Ryder and Thomson (1986) termed the Tiedemann Advance. At its type locality 30 km northwest of my study area, the Tiedemann Advance has been dated to 3300-1900 ^{14}C yr BP, culminating around 2300 ^{14}C yr BP (Ryder and Thomson, 1986; Arsenault, 2004). The radiocarbon age of 2790 ± 60 ^{14}C yr BP (820-1100 BC) from sediments just above bedrock at the base of the moraine fan complex north of Queen Bess Lake dates an advance of one of the glaciers in the Nostetuko River watershed to near its Neoglacial limit. The oldest fluvial aggradation age, several kilometres downvalley, is slightly younger, 2490 ± 70 ^{14}C yr BP (410-790 BC), suggesting a minor lag in response at that site.

Ryder and Thomson (1986) suggest that recession from the Tiedemann limit at the type section was slow and pulsatory until at least 1300 ^{14}C yr BP. Evidence is mounting, however, for a distinct glacier advance in the Coast Mountains, beginning as early as 1700 ^{14}C yr BP (AD 210-440) and culminating after 1400 ^{14}C yr BP (AD 540-770) (Koch et al., 2004a; Reyes and Clague, 2004; Reyes et al., 2006). A correlative advance has been documented in the Wrangell Mountains (Wiles et al., 2002), southern Kenai Mountains (Wiles and Calkin, 1994), and in the mountains of coastal Alaska (Calkin, 1988; Calkin et al., 2001; *see* also review in Reyes et al., 2006). Lichenometric studies record at least two pre-Little Ice Age advances of Tiedemann Glacier, one about AD 620 and the other around AD 925-933 (Larocque and Smith, 2003). The glacially overridden inner moraine at Queen Bess Lake is probably older than 1480 ± 50 ^{14}C yr BP (TO-8928; AD 430-660; Kershaw, 2002), and may have been constructed during an

advance of Diadem Glacier in the first millennium AD. If so, downvalley aggradation recorded by group 3 radiocarbon ages is the fluvial response to this advance.

Aggradation between 1030 ± 50 and 940 ± 50 ^{14}C yr BP (AD 900-1140 and AD 1000-1210) coincides with the onset of earliest Little Ice Age activity in the Coast and Rocky Mountains. Lichenometric analysis places the earliest Little Ice Age advance of Tiedemann Glacier at about AD 1118 (Larocque and Smith, 2003). One or more glacier advances in the southern Rocky Mountains at the same time have been documented by Osborn and Luckman (1988), Leonard and Reasoner (1999), and Luckman (2000). Koch et al. (2003) have identified a Little Ice Age advance of the same age in Garibaldi Park. The phase of floodplain aggradation between 710 ± 60 and 470 ± 60 ^{14}C yr BP (AD 1220-1400 and AD 1310-1640) coincides with this earliest Little Ice Age glacial activity.

Radiocarbon ages of 490 ± 60 ^{14}C yr BP (AD 1300-1620) from a peat clast within till of the outer moraine at Queen Bess Lake and 370 ± 50 ^{14}C yr BP (AD 1390-1590) from a stump in growth position below the T-2 terrace (Kershaw, 2002) are maxima for the time that Diadem Glacier achieved its greatest Holocene extent. An age of 180 ± 50 ^{14}C yr BP (AD 1640-1960) from fluvial sediments overlying ice-proximal gravel outwash (Kershaw, 2002) is a minimum for this event. Lichen measurements suggest that the outer Diadem Glacier moraine was abandoned in the middle or late 19th century (Kershaw, 2002). It was constructed, therefore, between the 15th and 19th centuries, which corresponds closely to the most recent aggradation phase in the valley.

2.6.3 Completeness and sensitivity of record

Moraine records have long been used to reconstruct glacial histories. However, most moraine systems record only recent glacier fluctuations because the major advances

of the Little Ice Age destroyed or buried much of the evidence of earlier glacier activity. Researchers have partially overcome this problem by examining more complete, although indirect, proxy records, including lacustrine varves and glacially sheared stumps and detrital logs in glacier forefields (Karlén, 1976; Leonard, 1986, 1997; Karlén and Matthews, 1992; Souch, 1994; Leonard and Reasoner, 1999; Luckman, 2000; Nesje et al., 2000; Menounos, 2002; Koch et al., 2003, 2004a; Smith, 2003; Menounos et al., 2004).

Lakes trap much of the inflowing sediment eroded from surrounding catchments, and sediment cores recovered from lakes can provide a comprehensive, integrated signal of basin-scale geomorphic activity (Oldfield, 1977; Souch, 1994). Intervals of coarse clastic sedimentation in proglacial lakes may record advances of alpine glaciers, but they may also represent large floods or hillslope processes within the basin (Leonard, 1986, 1997; Menounos et al., 2004). The location and size of a lake relative to its contributing watershed and the hydraulics of the fluvial system dictate the sensitivity and resolution of the record (Souch, 1994). High sedimentation rates in proglacial lakes offer high-resolution, proxy paleoenvironmental records, but the practical limit on the lengths of cores that can be retrieved in such situations results in a limited temporal record (Souch, 1994).

Another, largely unexploited archive of proxy information is fluvial deposits within glacierized basins. The sensitivity of the fluvial system to climate change has long been acknowledged, but fluvial deposits have not been widely used in reconstructing glacial histories. Proglacial areas, including laterally constricted valley trains, are areas of significant aggradation (Fenn and Gurnell, 1987), and the archived sediments can provide a record of glacier and climate change. Preservation of this archive will be greatest in

situations where aggradational events are not separated by erosion and incision (Maizels, 1990). In particular, frequent overbank deposition promotes floodplain and channel aggradation, with little erosion of former floodplain deposits.

My research has shown that, in favourable settings, the fluvial system is sensitive to low-magnitude climate change on centennial timescales. Increases in sediment supply, and attendant aggradation, in the west fork of Nostetuko River valley coincide with independently dated, middle and late Holocene glacier advances. Fluvial sequences may also provide a more complete record of Holocene glacier and climate change than deposits and landforms in glacier forefields, which are strongly biased toward the late Little Ice Age.

2.7 Conclusion

Evidence of late Holocene glacier fluctuations at the head of the west fork of Nostetuko River is preserved in the downvalley fluvial sedimentary sequence. Periods of floodplain stability, recorded by peat beds and forest growth, coincide with times when glaciers were restricted. Five major periods of base-level rise and aggradation coincide with times when glaciers were more extensive than today. The results of this study demonstrate that the valley fills of some mountain rivers archive information on glacier fluctuations on at timescales of centuries or less.

2.8 Acknowledgements

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provided helicopter transport to and from the study area. NSERC (Natural Sciences and Engineering Research Council of Canada) and Simon Fraser University funded the project.

2.9 Figures

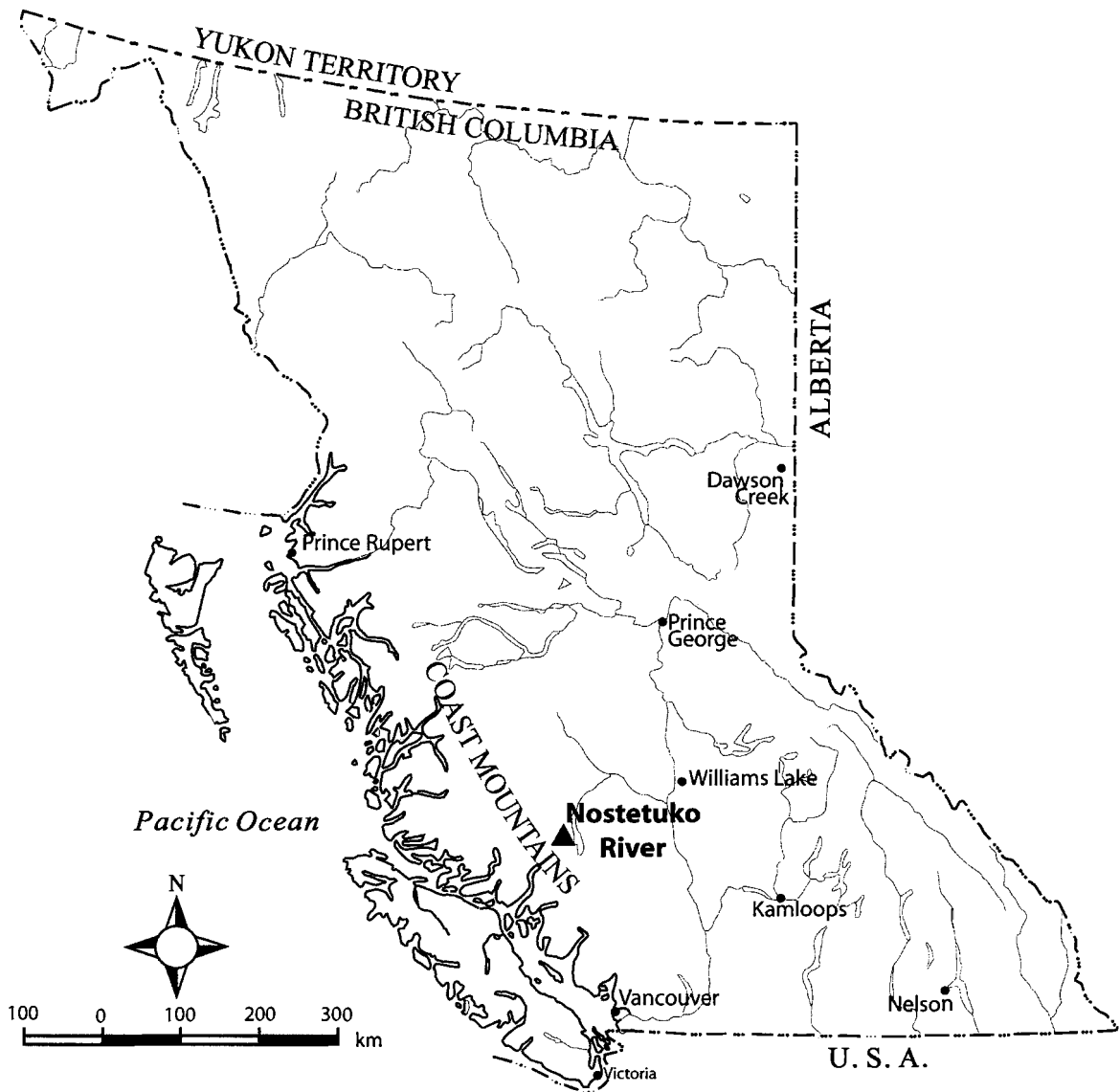


Figure 2.1 Location of the study area in the southern Coast Mountains of British Columbia (modified from BMGS data; reproduced with permission of the Province of British Columbia).

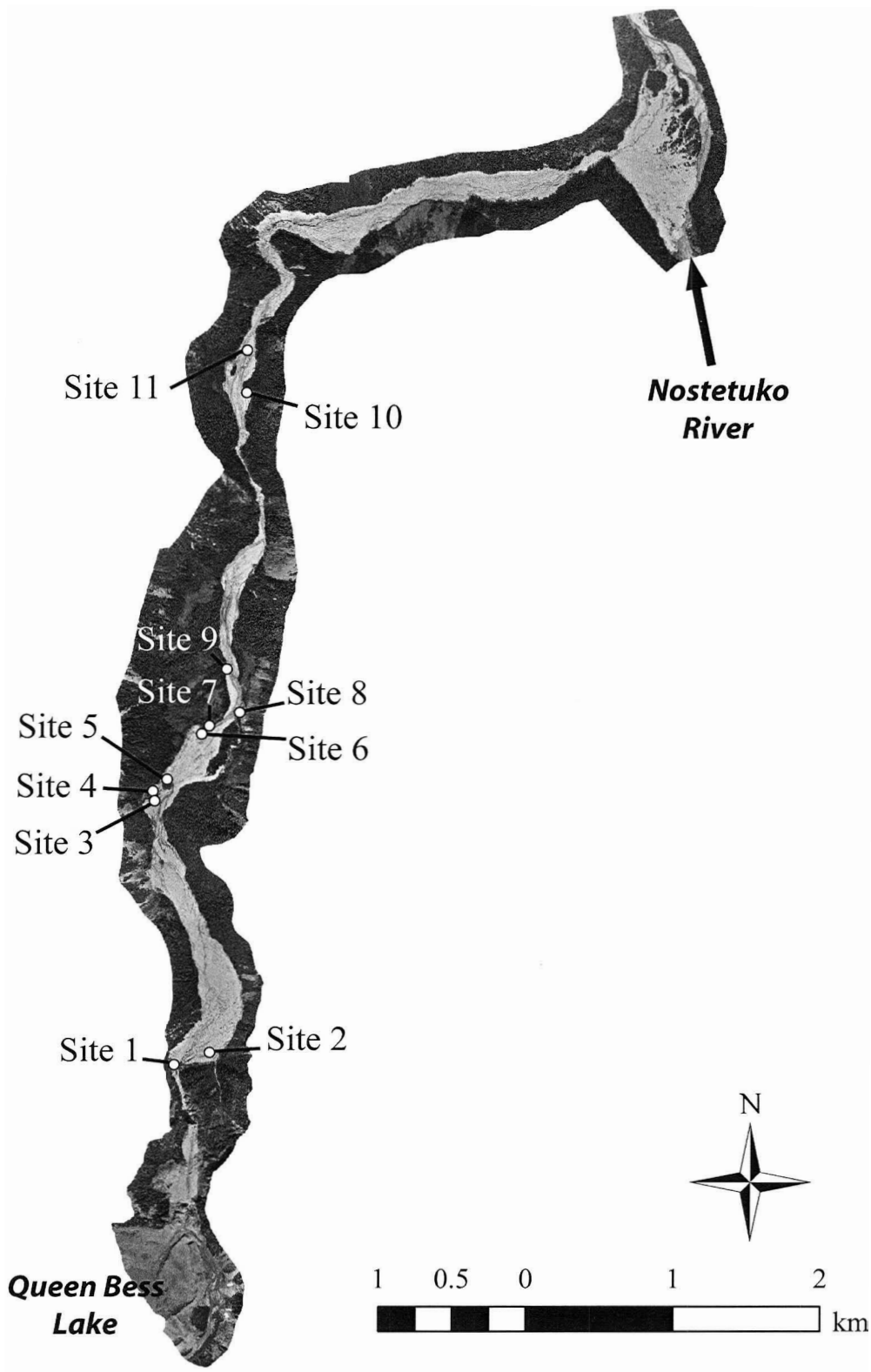


Figure 2.2 Aerial photomosaic of the valley of the west fork of Nostetuko River, showing locations of studied sections. The aerial photographs were flown in 1998, one year after the outburst flood.

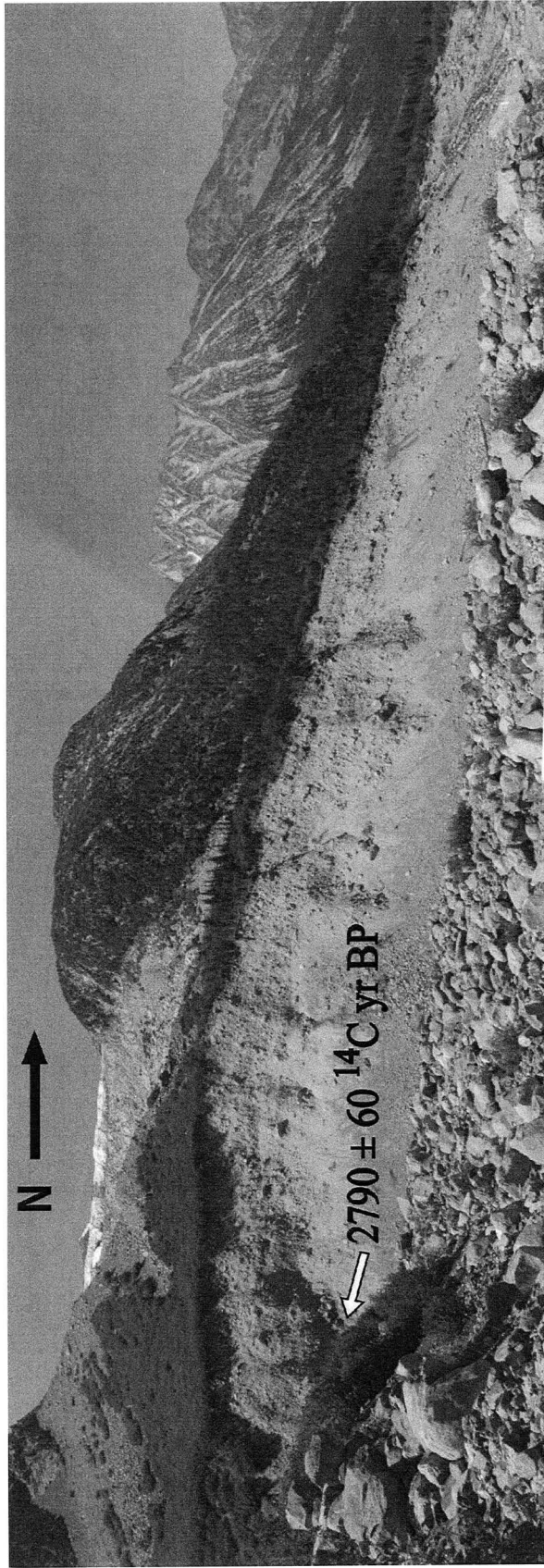


Figure 2.3 Photomosaic of the eroded distal face of a large moraine fan complex located 1 north of Queen Bess Lake. The moraine was built during late Holocene time.

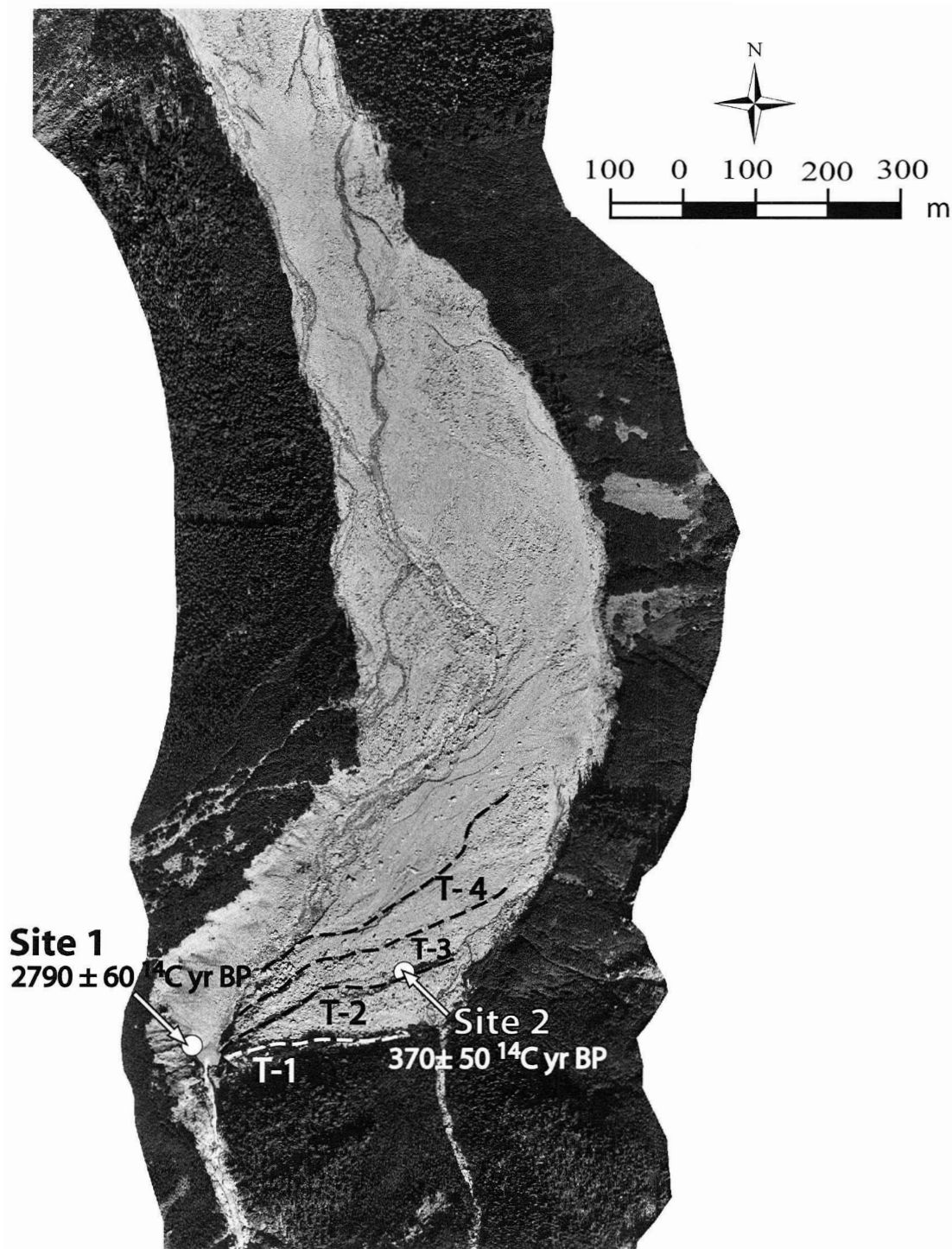


Figure 2.4 Four terraces (T-1, T-2, T-3, and T-4) inset into a gravel fan below the upper bedrock gorge. An *in situ* root in the scarp below T-2 yielded a radiocarbon age of 370 ± 50 ^{14}C yr BP (Kershaw, 2002).

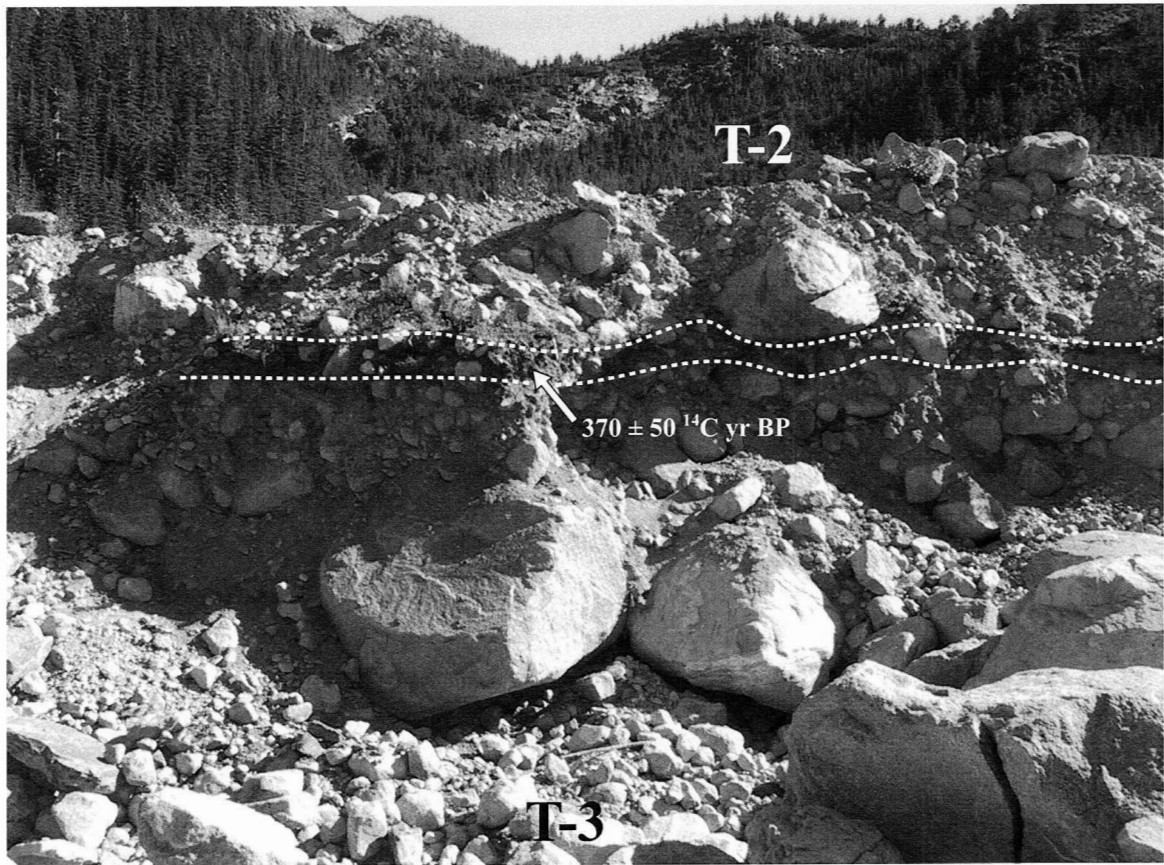


Figure 2.5 Scarp between terraces T-2 and T-3, showing a soil overlain and underlain by cobble-boulder gravel. Dashed lines bracket the dated soil.

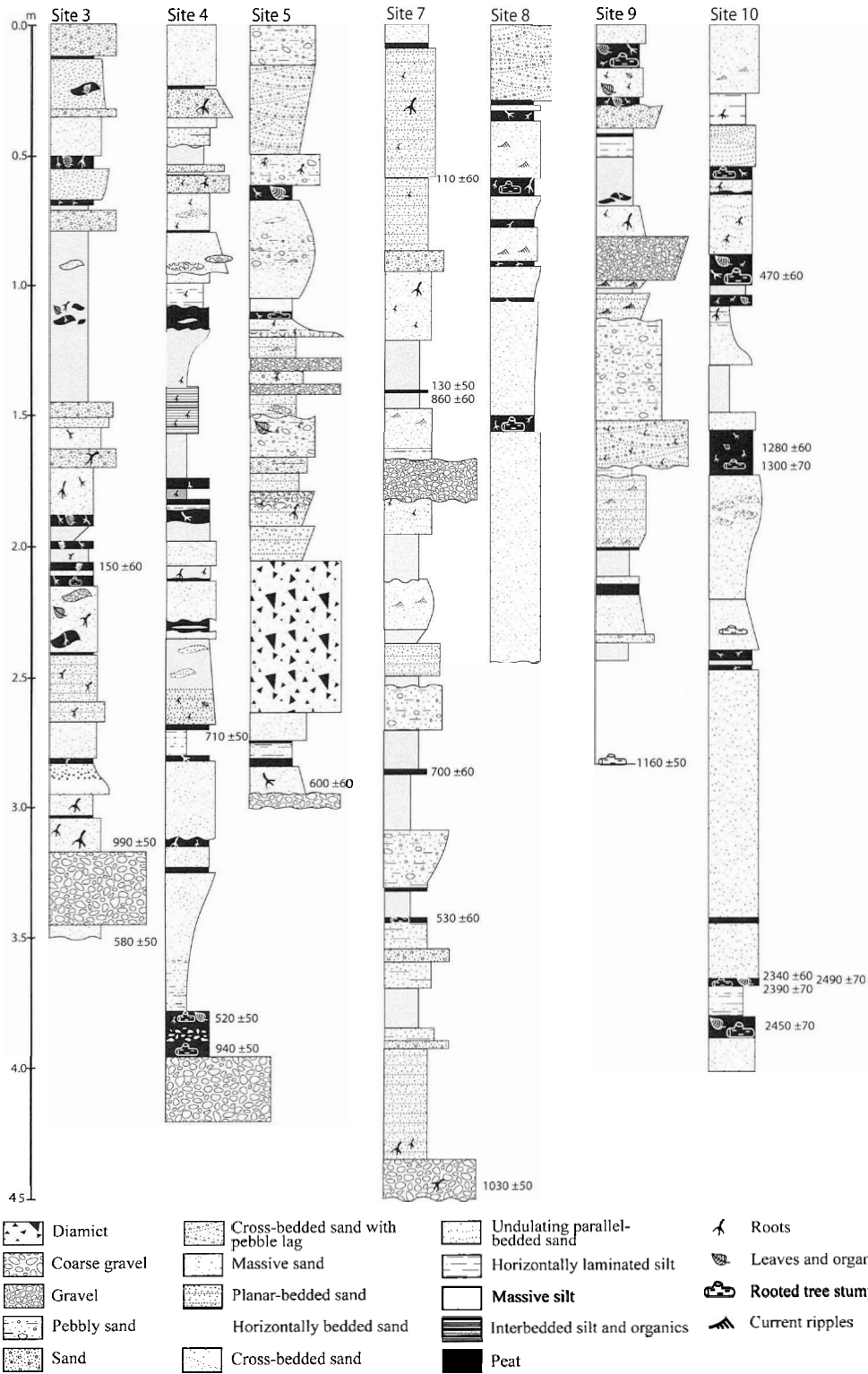


Figure 2.6 Lithostratigraphy of late Holocene sediments exposed at seven sites in the study area. Peat layers are shown in black. See Figure 2-2 and Appendix 1 for site locations and Table 2-1 for details on radiocarbon ages.

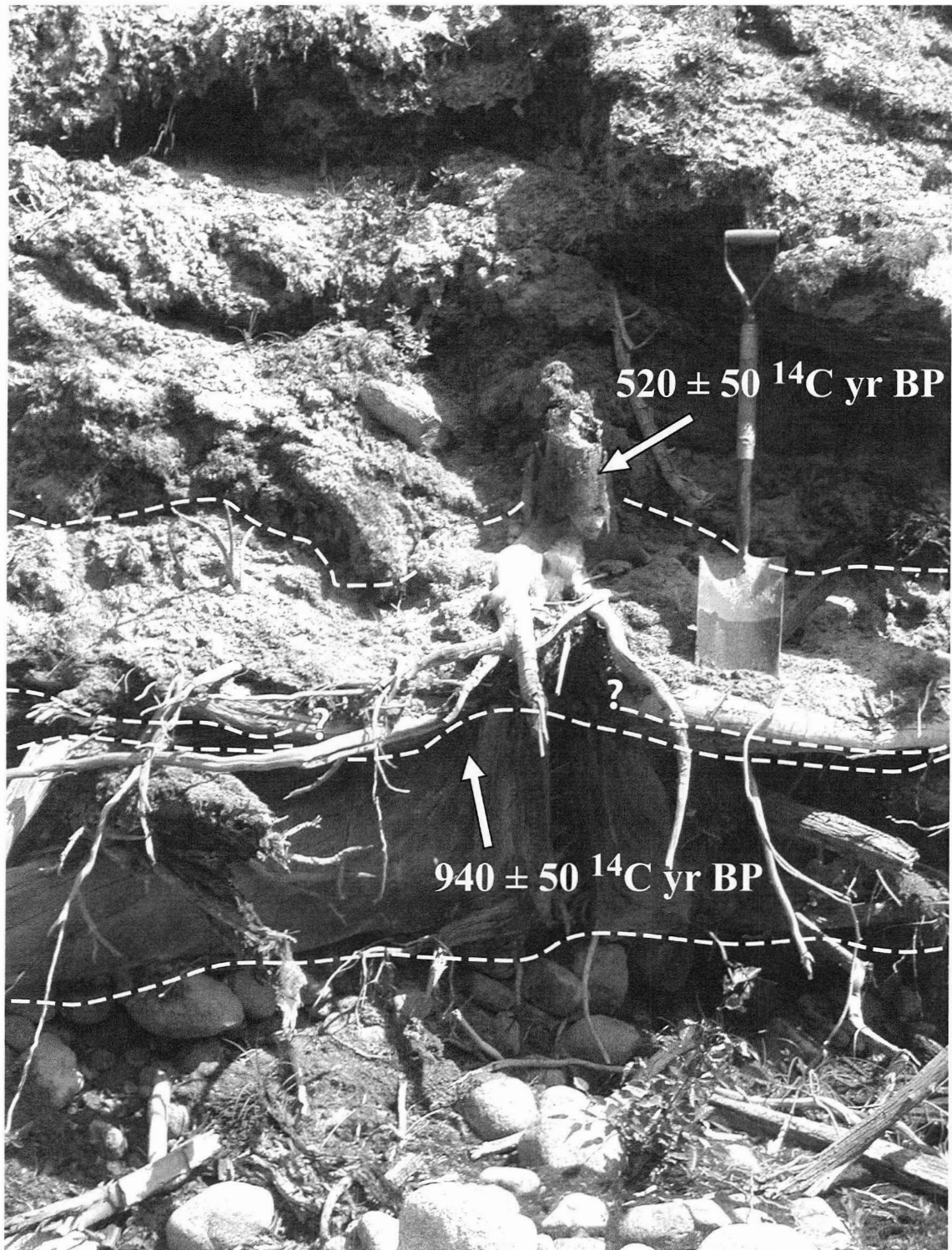


Figure 2.7 The lower part of the section at site 4. A basal gravel unit is sharply overlain by silt, sand, and peat beds. The two arrowed stumps are rooted in two peat beds separated by about 10 cm. Dashed lines delineate the peat beds. Radiocarbon ages of 520 ± 50 and 940 ± 50 ^{14}C yr BP (Table 2-1) were obtained on the outer rings of the two stumps.



Figure 2.8 Upper 2 m of the section at site 5, showing a prominent peat layer, an *in situ* stump, and an oxidized horizon (directly above the trowel). The heavy dashed lines delineate the peat layer, and the dotted lines mark contacts between sand beds. A radiocarbon age of 600 ± 60 ^{14}C yr BP was obtained on a root near the base of this section (not shown in the photograph).



Figure 2.9 Upper 2 m of the section at site 9, showing peat layers, laminated fine-grained sediments, and a stump in growth position. Dashed lines delineate peat layers. A radiocarbon age of 470 ± 60 ^{14}C yr BP was obtained on an *in situ* root (not shown in the photograph).

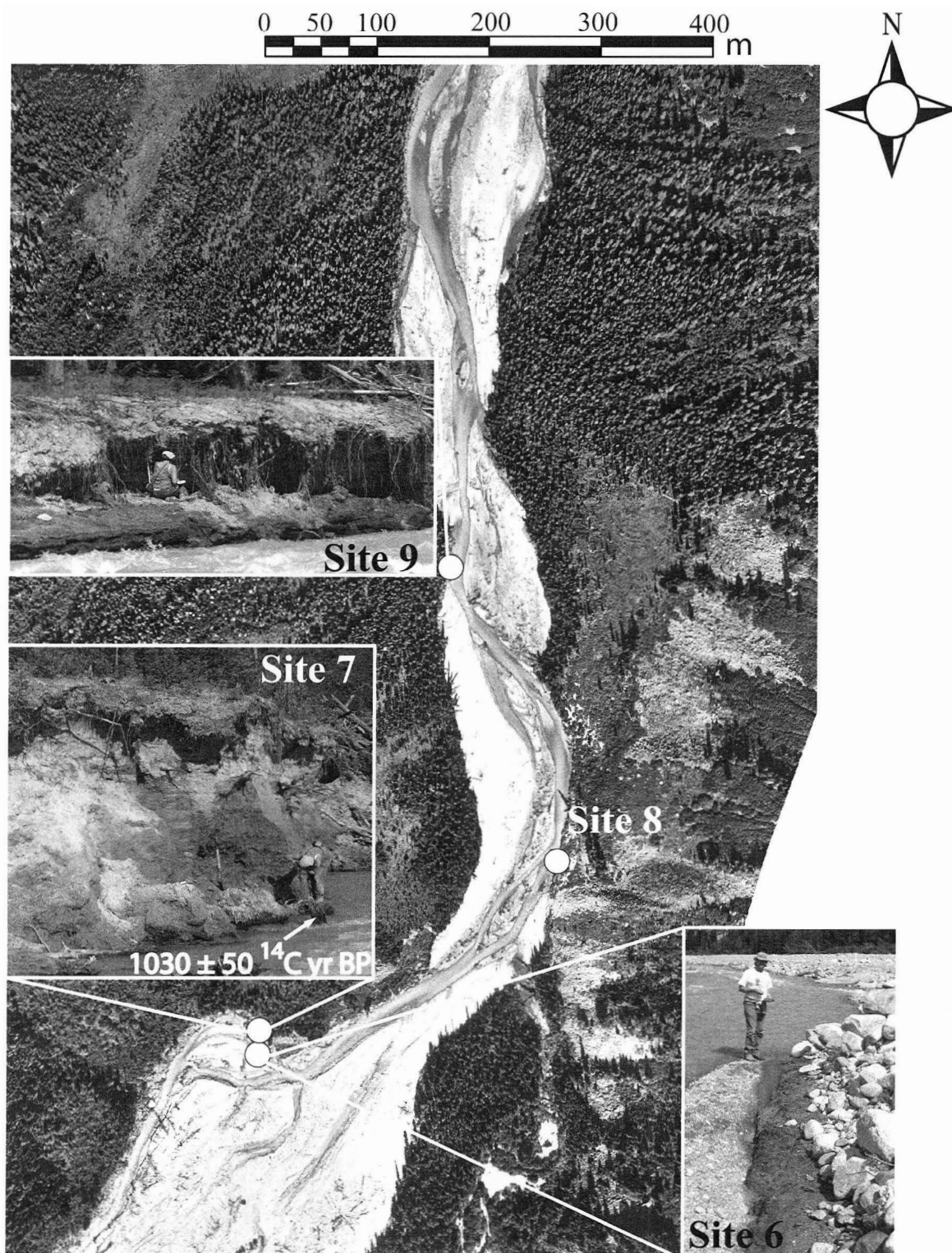


Figure 2.10 Examples of sections documented in this study.

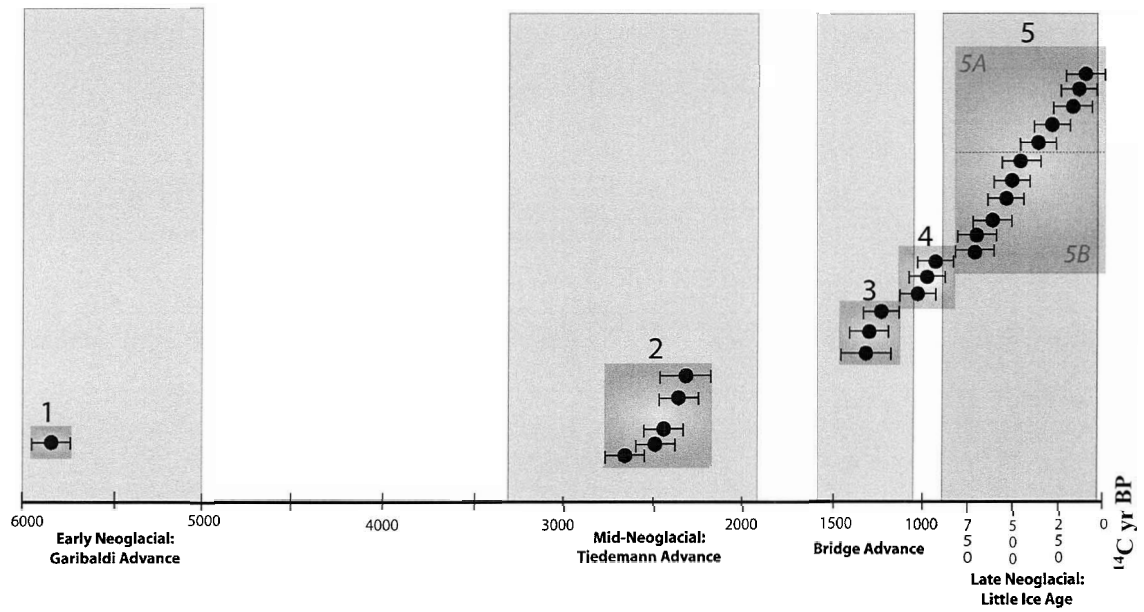


Figure 2.11 Plot of radiocarbon ages obtained for this study and their relation to independently dated Holocene glacier advances in western Canada. Cluster analysis placed the radiocarbon ages into five groups. The youngest group consists of two sub-groups (5A and 5B).

2.10 Tables

Table 2.1 Radiocarbon ages from the west fork of Nostetuko River valley.

Radiocarbon age ^b (¹⁴ C yr BP)	Calibrated age ^c (cal yr BP)	Laboratory no. ^a	Site no.	Location		Elevation (m)	Material	Chrono-stratigraphic units
				Lat. (N)	Long. (W)			
110 ± 60	280-0	TO-8935	7	51° 17' 24"	124° 30' 18"	1378	<i>in situ</i> root	
130 ± 50	290-0	Beta-200730	7	51° 17' 24"	124° 30' 18"	1378	<i>in situ</i> root	
150 ± 60	290-0	TO-8932	3	51° 17' 10"	124° 30' 31"	1386	outer rings of <i>in situ</i> stump	5a
270 ± 50	460-270; 180-150; 10-0	Beta-200723	6	51° 16' 45"	124° 30' 18"	1408	outer rings of <i>in situ</i> stump	
370 ± 50	560-360	TO-8923	2	51° 16.3'	124° 30.1'	1493	<i>in situ</i> root	
470 ± 60	640-590; 570-420; 390-310	TO-8942	10	51° 18' 40"	124° 30' 05"	1294	<i>in situ</i> root	
520 ± 50	640-590; 560-500	Beta-200727	4	51° 17' 10"	124° 30' 32"	1388	outer rings of <i>in situ</i> stump	
530 ± 60	660-480	TO-8933	7	51° 17' 24"	124° 30' 18"	1373	outer rings of <i>in situ</i> stump	
580 ± 50	660-520	Beta-200726	3	51° 17' 10"	124° 30' 31"	1384	herbaceous organics	5b
600 ± 60	670-520	Beta-200733	5	51° 17' 13"	124° 30' 29"	1384	root	
620 ± 50	670-530	Beta-200725	6	51° 17' 18"	124° 30' 18"	1377	outer rings of <i>in situ</i> stump	
700 ± 60	720-550	Beta-200729	7	51° 17' 24"	124° 30' 18"	1376	<i>in situ</i> root	
710 ± 60	730-550	Beta-200734	4	51° 17' 10"	124° 30' 32"	1387	outer ~10 rings of log	
940 ± 50	950-740	Beta-200728	4	51° 17' 10"	124° 30' 32"	1388	outer rings of <i>in situ</i> stump	
990 ± 50	1050-1030; 990-760	TO-8931	3	51° 17' 10"	124° 30' 31"	1385	root	4
1030 ± 50	1050-900; 850-810	Beta-200731	7	51° 17' 24"	124° 30' 18"	1375	outer rings of <i>in situ</i> stump	
1160 ± 50	1180-960	Beta-200732	9	51° 17' 37"	124° 30' 09"	1395	outer rings of <i>in situ</i> stump	
1280 ± 60	1300-1060	Beta-200736	10	51° 18' 40"	124° 30' 05"	1293	outer rings of <i>in situ</i> stump	3
1300 ± 70	1340-1050	TO-8941	10	51° 18' 40"	124° 30' 05"	1293	outer rings of <i>in situ</i> stump	
2340 ± 60	2690-2660; 2480-2310; 2240-2180	Beta-200737	10	51° 18' 40"	124° 30' 05"	1291	outer rings of <i>in situ</i> stump	
2390 ± 70	2750-2300	TO-8943	10	51° 18' 40"	124° 30' 05"	1292	<i>in situ</i> root	
2450 ± 70	2730-2350	TO-8939	10	51° 18' 40"	124° 30' 05"	1291	twig	2
2490 ± 70	2740-2360	TO-8940	10	51° 18' 40"	124° 30' 05"	1291	<i>in situ</i> root	
2790 ± 60	3050-2770	Beta-200721	1	51° 16' 10"	124° 30' 24"	1494	branch	
5810 ± 70	6750-6440	Beta-200735	11	51° 18' 53"	124° 30' 04"	1285	outer rings of <i>in situ</i> stump	1

^a Laboratories: Beta- Beta Analytic Inc.; TO- IsoTrace Laboratory (University of Toronto).

^b Ages have been corrected for natural and sputtering fractionation to a base of $\delta^{13}\text{C} = -25.0\text{‰}$.

^c Determined from atmospheric decadal data set of Stuiver et al. (1998) using the program OxCal v.3.9. The range represents 95.4% confidence limits.

CHAPTER 3

CONSTRAINING TIMES

OF LITTLE ICE AGE GLACIER FLUCTUATIONS

IN THE WEST FORK OF NOSTETUKO RIVER VALLEY

USING DENDROCHRONOLOGY

3.1 Abstract

Little Ice Age glacier fluctuations and river bed changes are recorded in the downvalley fluvial sequence of the west fork of Nostetuko River, southern Coast Mountains, British Columbia. Periods of floodplain stability are indicated by peat layers and forest growth. Episodes of valley-wide aggradation coincide with times of glacier advance. In the last 1000 years, four periods of peat formation, spanning more than 142, 102, 254, and 182 years, are separated by three intervals of aggradation. Calendric ages of floating tree-ring chronologies derived from *in situ* buried tree stumps were constrained by cross-dating a tree rooted in the youngest peat with a previously established regional living master chronology and by radiocarbon dating fossil wood. The three major periods of floodplain aggradation occurred sometime between AD 1650 and 1826, AD 1534 and 1724, and AD 1280 and 1336. These periods coincide with independently documented local and regional glacier advances. The results imply that the west fork of Nostetuko River is sensitive to upvalley glacier fluctuations and, indirectly, to relatively minor changes in climate.

3.2 Introduction

The Little Ice Age spans much of the last millennium and is a time of generally cooler climate and repeated advances of glaciers (Grove, 1988; Luckman, 2000). Little Ice Age glacier advances in western Canada are broadly synchronous with advances in other parts of the world. Two major periods of glacier advance are recognized: an early interval between ~AD 1200 and 1300, and a late interval from about AD 1500 to 1900 when most glaciers advanced to their maximum Holocene limits (Ryder and Thomson, 1986; Ryder, 1987; Luckman, 1988, 1993, 2000; Desloges and Ryder, 1990; Wiles and Calkin, 1990; Clague and Mathewes, 1996; Smith and Larocque, 1996; Wiles et al., 1999; Smith and Desloges, 2000; Luckman and Villalba, 2001; Clague et al., 2004; Lewis and Smith, 2004).

Much of the evidence for past episodes of glacier activity in western Canada comes from moraine records (Osborn, 1986; Ryder and Thomson, 1986; Osborn and Luckman, 1988; Desloges and Ryder, 1990; Luckman, 1993, 2000; Osborn et al., 2001; Reyes and Clague, 2004). However, poor dating control and the fragmentary nature of these records have restricted reconstruction of Holocene glacier histories to centennial and millennial timescales. High-resolution archives, such as lake sediment cores and tree-ring series, provide proxy records of changing environmental conditions and allow more detailed reconstructions of past glacial events than is possible from the moraine record alone (Fritts, 1976; Karlén, 1976; Leonard, 1986, 1997; Schweingruber, 1988; Karlén and Matthews, 1992; Souch, 1994; Stokes and Smiley, 1996; Wiles et al., 1996; Leonard and Reasoner, 1999; Nesje et al., 2000; Menounos, 2002; Menounos et al., 2004). Studies of proglacial lake sediments in the Coast and Rocky Mountains indicate that sedimentation

increases during and following glacier advances and is low during periods of reduced ice extent (Leonard, 1986, 1997; Menounos, 2002; Menounos et al., 2004).

Dendrochronologic and dendroclimatic reconstructions in the southern Canadian Cordillera indicate major climatic shifts coincident with episodes of glacier expansion and retreat, and allow assignment of calendar ages to these events (Luckman, 2000; Larocque and Smith, 2003, 2005a, 2005b; Lewis and Smith, 2004; Luckman and Wilson, 2005).

Other paleoenvironmental records provide information on Holocene glacier fluctuations. One such record, the subject of this paper, is proglacial fluvial sediments and landforms (Maizels, 1979, 1995; Fenn and Gurnell, 1987; Marren, 2005). Glacier advance and retreat affect rates and amounts of sediment and meltwater delivered to mountain rivers (Maizels, 1979). Changes in sediment flux, in turn, can lead to aggradation or incision of the floodplains of these rivers.

This paper documents late Holocene changes in bed elevation in the west fork of the Nostetuko River valley in the southern Coast Mountains of British Columbia (Fig. 3.1) through stratigraphic and dendrochronological analyses of *in situ* forest remains in buried peat layers. Development of 'floating' chronologies and cross-dating of a tree rooted in the youngest peat with a regional master chronology allow assignment of ages to periods of surface stability and forest growth. Radiocarbon ages on *in situ* buried fossil wood provide further constraints on periods of aggradation.

3.3 Study area

The west fork of Nostetuko River flows 11 km north and east from its source to the main stem of Nostetuko River (Fig. 3.2). It is fed by meltwater from valley glaciers at the edge of Homathko Ice Field. A major tributary of the west fork flows from Queen Bess Lake, a moraine-dammed lake that partially drained during an outburst flood in August 1997 (Kershaw, 2002; Kershaw et al., 2004).

Queen Bess Lake is impounded by a large composite moraine produced by at least two Neoglacial advances of Diadem Glacier (Fig. 3.2; Kershaw, 2002). The lake formed behind the composite moraine during glacier retreat in the late 1800s and early 1900s. In 1997, a large ice avalanche fell into Queen Bess Lake from the toe of Diadem Glacier and produced displacement waves that overtopped and incised the moraine. The resulting flood eroded sediments in the valley below the moraine dam, creating exposures of the upper part of the valley fill in which there are numerous peat layers containing *in situ* tree stumps (Fig. 3.3).

Subalpine forest in the study area is dominated by subalpine fir (*Abies lasiocarpa*) and whitebark pine (*Pinus albicaulis*). Subalpine fir also extends onto the lower slopes of the study area, providing an opportunity to cross-date subfossil tree remains in the peat beds. Mountain hemlock (*Tsuga mertensiana*) and yellow cedar (*Chamaecyparis nootkatensis*) are also common on lower elevation slopes in this area (Larocque and Smith, 2003).

3.4 Methods

3.4.1 Sampling

Cross-section disks were collected from *in situ* fossil tree stumps and logs in exposed peat layers, mainly along the west side of the valley. Three stumps were sampled from the valley floor (KW15, KW17, and KW18; Table 3.1) and one from an eastern exposure (KW14a). Disks of three detrital tree stems (KW03, KW05, and KW09) were collected from an apron of colluvium at the Little Ice Age limit of a tributary glacier on the west side of the valley (Fig. 3.4). Tree stumps were rooted in peat layers and were sampled at the stem base where possible. The sampled trees are probably subalpine fir (*Abies lasiocarpa*) because this species dominates the forest around the sample sites. Field identification was limited, however, because the trees lacked bark.

3.4.2 Laboratory analysis

Samples were air-dried and sanded several times with progressively finer sand paper. Annual tree-ring widths were measured to the nearest 0.001 mm along up to four radii for each tree sample using a Velmex-type measuring stage, a Leitz stereomicroscope, and the Measure J2X measuring program. Attempts were made to cross-date samples and to establish floating chronologies by visually comparing marker rings and by employing the statistical correlation and verification procedures within the ITRDBL (International Tree-Ring Data Bank Library) tree-ring dating software program COFECHA (Holmes, 1999; Grissino-Mayer, 2001). Verification was based on 50-year dated segments with 25 year lags, significant at the 99% critical level of correlation of 0.3281 (Holmes, 1999; Grissino-Mayer, 2001). Low-frequency variance was removed by filtering with a cubic smoothing spline having a 50% cutoff of 32 years. An

autoregressive model was fit to the data to remove any remaining persistence within the smoothed series, and log transformation was performed to weight proportional differences in ring measurements more equally. Segments that were not significantly correlated were re-measured and corrected to account for radial growth anomalies and missing or false rings.

Tree-ring cross-dating done in this study involved two strategies. First, a regional living tree-ring chronology from Cathedral Glacier, about 23 km west of the study area (Laroque and Smith, 2003), was used to cross-date a tree in the youngest peat layer. Second, floating ring-width chronologies were constructed from trees rooted in several other peat layers, and ages were determined based on stratigraphic relations and radiocarbon dating of stumps and logs in the peats (Chapter 2).

Outer rings of three stumps analyzed for tree-ring dating and three additional logs, stumps, and roots in peat layers were radiocarbon dated. Dating was done at Beta Analytic and IsoTrace Laboratory (University of Toronto). Radiocarbon ages were calibrated using the program OxCal v.3.9 (Bronk Ramsey, 1995, 2001), which is based on the decadal dataset of Stuiver et al. (1998).

The age of the oldest living tree on a surface provides a minimum age for that surface, after corrections for local ecesis and sampling height have been made (McCarthy et al., 1991; Wiles et al., 1996). Ecesis, defined as the time between surface stabilization and germination of the first seedling, has been shown to range from 1 to 100 years in the Pacific Northwest (Sigafos and Hendricks, 1969; Desloges and Ryder, 1990; McCarthy et al., 1991; Smith et al., 1995; Wiles et al., 1999; Luckman, 2000; Lewis and Smith, 2004). Ecesis intervals of 1 to 4 years have been documented in the Coast Mountains at

Tiedemann Glacier (Larocque and Smith, 2003) and on Vancouver Island at Colonel Foster and Septimus glaciers (Lewis and Smith, 2004). Seedlings growing on the floodplain scoured by the 1997 Queen Bess outburst flood were no more than five years old at the time of field work in 2004, suggesting that ecesis in the west fork valley is 1-2 years. Two years were therefore added to the outer ring ages of *in situ* stumps to correct for ecesis (Table 3.1).

Sampling height errors occur when annual growth rings are lost due to sampling above the root crown (McCarthy et al., 1991). Larocque and Smith (2003) proposed a regional correction factor of 1.35cm a^{-1} for subalpine fir seedlings on valley floors in the Mount Waddington area. This correction was applied to all *in situ* stumps (Table 3.1). Sampling height corrections cannot be applied to detrital logs.

3.5 Results

3.5.1 Internal tree-ring correlations

Twenty-six radial series from ten trees were used in the analysis. Statistics were calculated to show the characteristics of the tree-ring chronologies (Table 3.1). The statistics include: mean sensitivity, a measure of annual variability in the tree rings; serial correlation, a measure of the amount of common signals among tree-ring sequences; and auto-correlation, a measure of the association between growth in the previous year and that in the current year. Series correlation values within each sample range from 0.341 to 0.671, and mean sensitivity values range from 0.144 to 0.356. For reference, low values of mean sensitivity range from 0.10 to 0.19, intermediate values from 0.20 to 0.29 and high values are more than 0.30 (Grissino-Mayer, 2001). High first-order coefficients

before detrending suggest that low sensitivity values are attributable to lag effects that may persist for up to three years in the case of subalpine fir (Peterson et al., 2002).

3.5.2 Stratigraphic correlations and ring-width chronologies

Initial cross-dating focused on the development of floating ring-width chronologies for peat layers correlated on the basis of stratigraphy and radiocarbon ages (Chapter 2, Appendix 2). Cross-dating of *in situ*, radiocarbon-dated samples with stratigraphically related fossil trees of unknown age provided a framework for the floating chronologies. Floating chronologies did not cross-date with one another, reducing the possibility of overlap in ages of peat layers and supporting the assumption of limited or no survival of trees between aggradation events.

Four ring-width chronologies were constructed based on stratigraphy and radiocarbon ages (Chapter 2). The oldest radiocarbon age (620 ± 50 ^{14}C yr BP; Table 3.2) is from the outer rings of an *in situ* stump in peat. Cross-dating of this sample with stratigraphically related samples ($r = 0.434$, significant at the 99% confidence level) produced an uncorrected floating chronology of 176 years (Fig. 3.5; Table 3.2). Corrections for ecesis and sampling height extended the interval by five years. Individual tree series end within 14 and 19 years of one another. The outer surfaces of the analyzed samples are weathered, consequently an unknown number of rings have been lost and precise kill dates are unknown. Further, the chronology provides only a minimum estimate of the age of the surface on which the peat was deposited.

A second floating chronology is associated with a radiocarbon age of 370 ± 50 ^{14}C yr BP on an *in situ* stump, reported by Kershaw (2002) (Table 3.2). A disk from an

adjacent stump on the same surface was cross-dated with two other samples 2 km downvalley ($r = 0.420$, significant at the 99% confidence level). The uncorrected chronology records surface stability for a minimum of 237 years. This interval was extended to 254 years by applying corrections for ecesis and sampling height (Table 3.1). The minimum kill dates are within 12 and 34 years of one another (Fig. 3.5).

The outer rings of a buried log in the moraine fan complex on the west side of the valley, 1.5 km north of Queen Bess Lake, yielded a radiocarbon age of 150 ± 60 ^{14}C yr BP (Table 3.2). Cross-dating of this log with two other buried logs at the same stratigraphic level ($r = 0.487$, significant at the 99% confidence level) produced an uncorrected chronology of 100 years, which was extended two years for ecesis (Fig. 3.5).

A radiocarbon age of 0 BP was obtained on a stump with 127 annual rings in the youngest peat layer (Table 3.2). The radiocarbon age indicates that the tree was living within the last 50 years. The dated sample yielded 104.02 ± 0.66 pMC (% of modern reference standard), suggesting that tree death was in the late 1960s (D. Hood, written communication, 2005). The ring series cross-dates with the living subalpine fir master chronology of Larocque and Smith (2003) ($r = 0.438$). With the addition of ecesis and sampling height corrections, the series dates to AD 1826-1968 (Table 3.1).

3.5.3 Ages of peat layers and aggradation

Calendar ages were assigned to peat and clastic sediment units based on (1) the assumed date of death of the youngest buried tree, (2) the lengths of the four ring series, and (3) radiocarbon ages on detrital logs and *in situ* stumps and roots in the four peat

layers (Fig. 3.6). Calibrated radiocarbon ages on *in situ* fossils are assumed to date the time of death and burial of the trees.

The elevation of the base of each peat layer is the maximum elevation of base level when that peat was deposited. The elevation of the base of each clastic unit is base level at the time the underlying peat was buried. The length of time spanned by each peat layer, which must equal or exceed the lifespan of its associated trees, provides a further constraint on the ages of stable and aggradation intervals. The length of time between the death of trees on a floodplain surface and the inception of tree growth on the next younger surface is a maximum for the period of intervening aggradation.

The four intervals of peat deposition and inferred surface stability span, from oldest to youngest, more than 182, 254, 102, and 144 years (i.e., the number of years in the floating tree-ring series; Fig. 3.5, Table 3.1). The four stable intervals are separated by three periods of aggradation, during which the peat layers were buried (Fig. 3.6).

The first interval of peat deposition and surface stability ended between AD 1280 and 1420 (2σ cal yr age range; 620 ± 50 ^{14}C yr BP). The second interval of peat deposition ended between AD 1390 and 1590 (2σ cal yr age range; 370 ± 50 ^{14}C yr BP). Both intervals can be more tightly constrained by considering (1) the extreme limits of the calendric age ranges and (2) the minimum duration of deposition of the second peat (254 years). The first interval thus must have ended after AD 1280 and before 1336 (AD 1590 - 254 years), and the second interval after AD 1534 (AD 1280 + 254 years) and before 1590. The period of aggradation separating these two stable intervals must be less than 56 years (AD 1590 - AD 1280 - 254 years). Further, the beginning of the first stable interval must date to before AD 1154 (AD 1336 - 182 years).

The third interval of peat formation and surface stability ended between AD 1650 and 1950 (2σ cal yr age range; 150 ± 60 ^{14}C yr BP), but it too can be more tightly constrained. It must have ended before AD 1826 because the fourth (last) interval of peat deposition began before that date. The third interval of peat deposition thus ended after AD 1650 but before AD 1826. It began after AD 1534 (constraint imposed by termination of next older peat), but before AD 1724 (AD 1826 - 102 years).

Given the above constraints, the period of aggradation separating the second and third intervals of peat deposition is less than 190 years long, perhaps considerably less. Similarly, the period of aggradation separating the third and fourth (youngest) intervals of peat deposition is less than 176 years. The end of the oldest interval of surface stability is constrained by a radiocarbon age of 620 ± 50 ^{14}C yr BP (AD 1280-1420) on an *in situ* root from the top of the oldest (first) peat.

In summary, the three intervals of floodplain aggradation are dated to between AD 1280 and 1336, AD 1534 and 1724, and AD 1650 and 1826 (Fig. 3.6). It must be emphasized that aggradation may have occurred over a shorter period within each of these intervals.

3.6 Discussion

3.6.1 Little Ice Age chronology

Periods of floodplain stability, recorded by peat layers and forest growth, coincide with times of base-level stability or lowering. The three major periods of valley-wide aggradation record increases in sediment delivery to the fluvial system and probably are times when glaciers were advancing and were more extensive than today (Chapter 2).

Lichenometric and radiocarbon ages indicate that Diadem Glacier retreated from its outermost moraine in the middle to late 1800s (Kershaw, 2002). The retreat appears to coincide with downvalley floodplain stability from AD 1826 to 1968. Dendroclimatic reconstructions for the Mount Waddington area show that mean summer temperature was relatively high from AD 1823 to 1867 and from AD 1902 to 1973 (Larocque and Smith, 2005b). Temperature and precipitation reconstructions for the Canadian Rockies based on tree rings also document higher temperatures and precipitation during the late 1800s and 1900s (Luckman et al., 1997; Luckman, 2000; Luckman and Wilson, 2005). Generally, glacier recession was most rapid during the first half of the 20th century (Gardner, 1972). Some glaciers readvance short distances during the period 1950-1970 (Luckman et al., 1987; Luckman, 2000; Koch et al., 2004b).

The period of valley-wide aggradation between AD 1650 and 1826 coincides with a Little Ice Age advance of Diadem Glacier and construction of its outermost moraine. The maximum age of this advance is constrained by a radiocarbon age of 490 ± 60 ¹⁴C yr BP (AD 1300-1620) on a peat clast within the moraine. A radiocarbon age of 180 ± 50 ¹⁴C yr BP (AD 1640-1960) from fine-grained fluvial sediments overlying ice-proximal outwash is a minimum date for retreat of the glacier from the moraine (Kershaw, 2002). Glacier mass balance and dendroclimatic reconstructions for the Mount Waddington area indicate cool and wet conditions throughout the 1700s and early 1800s (Larocque and Smith, 2005a, 2005b). Summer temperatures in the Rocky Mountains during the 1690s were the lowest of the last 1000 years (Luckman and Wilson, 2005). A major moraine-building episode in the late 1600s and early 1700s, documented in Alaska and the Coast and Rocky Mountains (Desloges and Ryder, 1990; Clague and Mathewes, 1996; Calkin

et al., 1998; Wiles et al., 1999; Luckman, 2000; Smith and Desloges, 2000; Koch et al., 2003; Larocque and Smith, 2003; Lewis and Smith, 2004) probably coincides with construction of the outer moraine of Diadem Glacier and the youngest aggradation period. Leonard (1997) documented high sedimentation rates at Hector Lake from the early 1700s until the mid-1800s, with a peak in the first two decades of the 1700s.

Floodplain stability commencing between AD 1534 and 1724 and ending between AD 1650 and 1826, lasted at least 102 years (Fig. 3.6) and probably corresponds to a documented warm interval within the Little Ice Age, separating early and late Little Ice Age glacier advances (Ryder and Thomson, 1986; Ryder, 1987; Desloges and Ryder, 1990; Clague and Mathewes, 1996; Calkin et al., 1998; Wiles et al., 1999; Luckman, 2000; Smith and Desloges, 2000; Koch et al., 2003; Larocque and Smith, 2003). Berendon Glacier in the northern Coast Mountains was less extensive than during the historic period between AD 1440 and 1660 (Clague et al., 2004). Dendroclimatic reconstructions for the Mount Waddington area extend back only to AD 1690, but relatively high reconstructed mean summer temperatures and below-average spring snowpack depths between AD 1690 and 1700 suggest this was a time of relatively warm climate (Larocque and Smith, 2005b). Temperatures in the Canadian Rocky Mountains during the 1500s and early and middle 1600s were also above average (Luckman, 2000; Luckman and Wilson, 2005).

Aggradation and forest burial began between AD 1534 and 1590 (Fig. 3.6) and ended before AD 1724 (maximum of 190 years). This aggradation event may coincide with construction of Little Ice Age moraines in the Mount Waddington area at about AD 1562-1575 and 1597-1621 (Larocque and Smith, 2003). A major moraine-building phase

on the wetter, west side of the Coast Mountains dates to AD 1690-1730 (Smith and Desloges 2000; Koch et al., 2004b; Lewis and Smith, 2004). This event is possibly the longest aggradation interval documented in this study and may include more than one glacier advance.

The second-oldest peat layer records floodplain stability for a minimum of 254 years after AD 1280. Base level was lower than the level of the peat from the early 1300s to the middle 1500s, coincident with the Medieval Warm Period (Grove, 1988; Houghton et al., 1990). Reconstruction of summer temperatures in the Canadian Rockies indicates warmth, comparable to that of the 20th century, in the first half of the eleventh century, the late 1300s, and early 1400s (Luckman and Wilson, 2005). The lowest sedimentation rates of the last millennium in Hector Lake, Alberta, occurred between AD 1375 and 1550 (Leonard, 1997). No evidence of glacier advances has been found in the Canadian Rockies during this period, and treeline was higher during this period than later in the Little Ice Age (Luckman, 1994).

The oldest, and temporally best constrained, Little Ice Age aggradation event occurred between AD 1280 and 1336. Dendrochronological dating in the Rocky Mountains indicates that Robson Glacier advanced over a forest between AD 1142 and 1350, Peyto Glacier advanced between about AD 1246 and 1375, and Stutfield Glacier advanced after AD 1272 (Luckman, 1995, 1996, 2000; Osborn et al., 2001). The aggradation event also coincides with a variable, but generally cool period in the Rocky Mountains between AD 1200 and 1350 (Luckman and Wilson, 2005).

3.6.2 Possible sources of error

The small number of analyzed samples is a potential source of error in this study, although limited sample size is compensated, to some extent, by temporal constraints imposed by the radiocarbon ages. Small samples are subject to considerable statistical random variation and lower reliability of all statistics. I have placed considerable weight on a statistically significant correlation between a single sample from the study area and the Cathedral Glacier master chronology, yet the minimum sample size recommended for dendrochronological reconstruction is 5-10 trees (Fritts, 1976). Each of the three floating chronologies used in this study was constructed from three samples, each with multiple radial series. Additional samples would strengthen the correlations and improve confidence in the cross-dates.

Another problem is that loss, through weathering, of the bark and outer rings of the stumps, made it impossible to assign precise dates to the death of the trees. Also, the floating ring series for the trees likely do not record the entire intervals spanned by the peat layers.

Errors in correlating tree-ring series may result from the lack of sensitivity, or complacency, of rings to changes in climate, to missing or false rings, and to natural ring-width variability along different radial sections of samples. Complacent ring series have little variation in ring width and result from non-stressed growing conditions (Stokes and Smiley, 1968; Fritts, 1976, Wiles et al., 1996). To overcome some of these problems, cross-sectional disks were collected instead of incremental cores and two to four radii were measured for each sample. Additional radii were added when visual inspection and statistical verification suggested the possibility of missing or false rings. Cross-dating of

trees in the same horizon helped to identify locations of missing and false rings within a series (Heikkinen, 1984).

Errors can also result from assumptions about sampling height and ecesis. Cores and disks should be taken as low on the stem as possible to maximize the number of rings sampled (McCarthy et al, 1991; Wiles et al., 1996). McCarthy et al. (1991) recommend sampling the lowest 10 cm of the stem. Tree stumps were sampled at the stem base where possible to minimize sampling height errors. The ecesis estimate used in this study is two years and may slightly underestimate the true ecesis interval.

The strength of correlations among correctly cross-dated trees may differ with tree species, geographic area, site homogeneity, amount of stand competition, and degree of disturbance (Grissino-Mayer, 2001). Low correlations between samples from the west fork of the Nostetuko River valley and the regional master chronology may stem from differences in tree species. Identification of samples to species level was not attempted, but as subalpine fir dominates the forest around the sample sites, it probably forms the majority of the fossil wood. Cross-dating was done using the nearest living-tree ring series (Cathedral Glacier; Larocque and Smith, 2003) to strengthen correlations and minimize these errors.

3.7 Conclusion

Four intervals of peat deposition, forest growth, and channel stability, interrupted by three aggradation intervals, are recorded in the upper part of the sediment fill in the west fork of the Nostetuko River valley. Channel aggradation coincides with times when glaciers were advancing or at advanced Holocene positions. Peat layers and periods of

floodplain stability coincide with times of warmer temperature and more restricted glaciers. Despite the limited sample size and several sources of error, the results show that proglacial fluvial sediments can record glacier fluctuations on decadal to centennial timescales.

3.8 Acknowledgements

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3.9 Figures

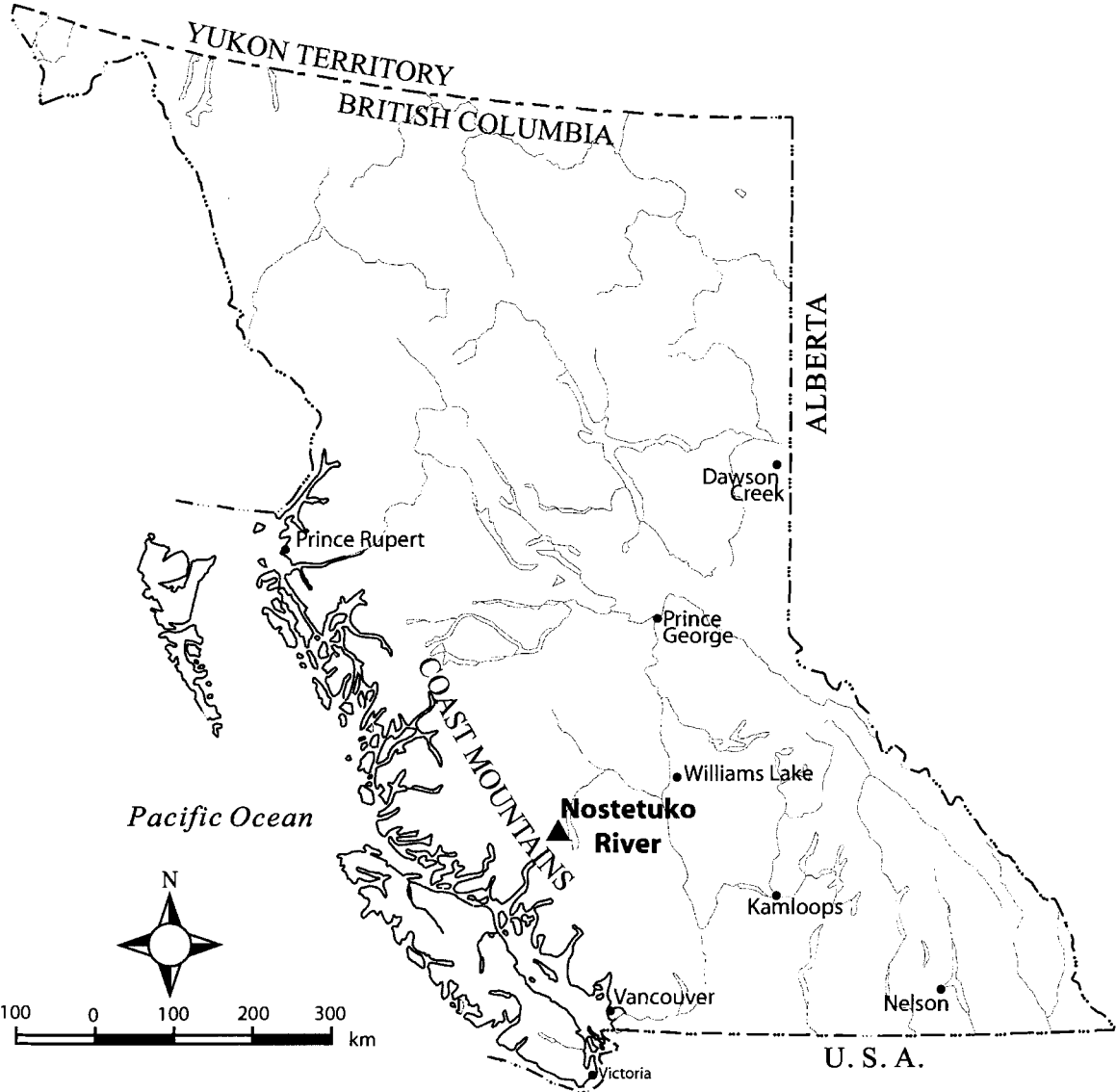


Figure 3.1 Location of the study area (modified from BMGS data; reproduced with permission of the Province of British Columbia).

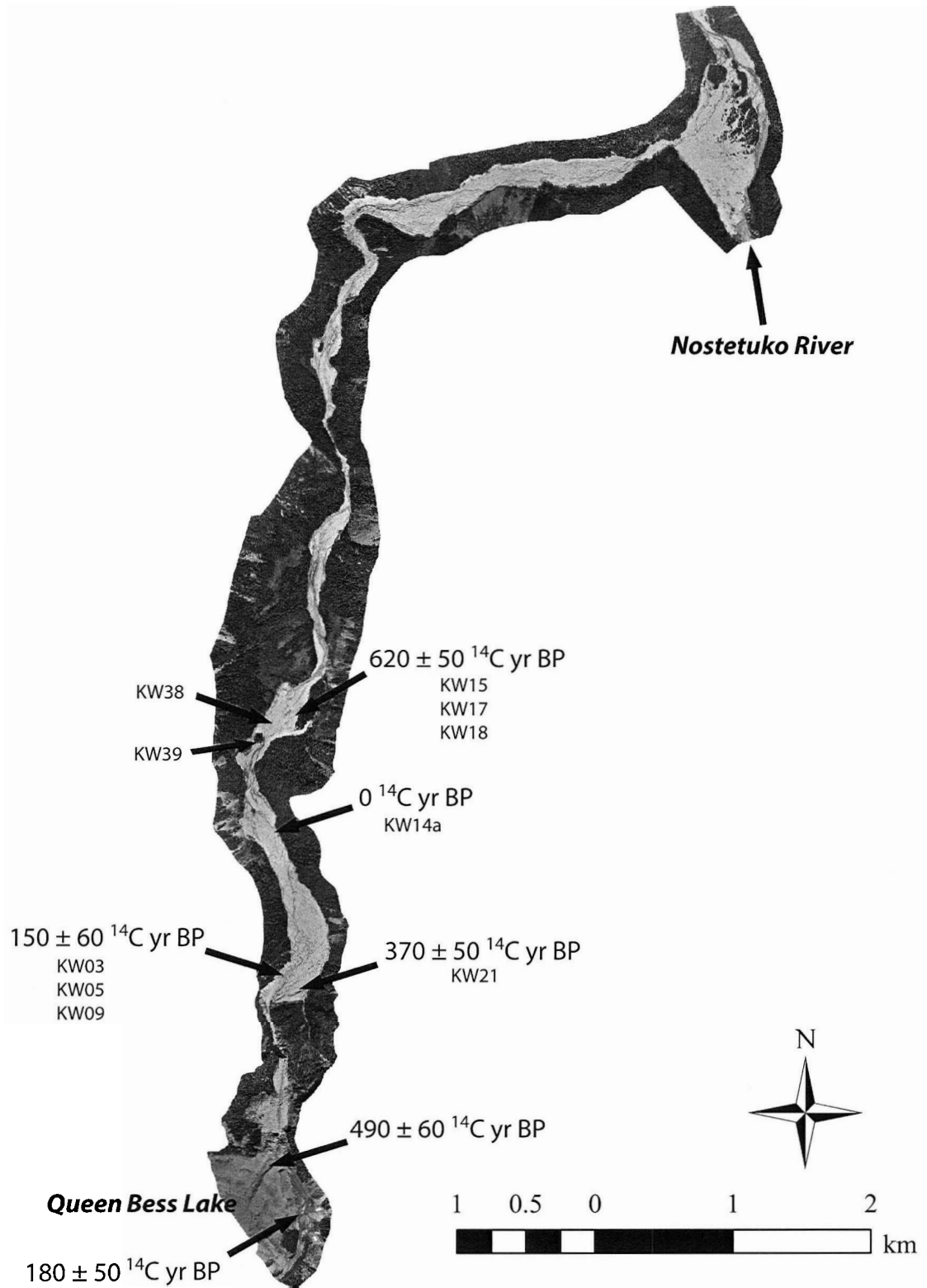


Figure 3.2 Aerial photomosaic of the valley of the west fork of Nostetuko River, showing locations of sampled trees and radiocarbon ages.

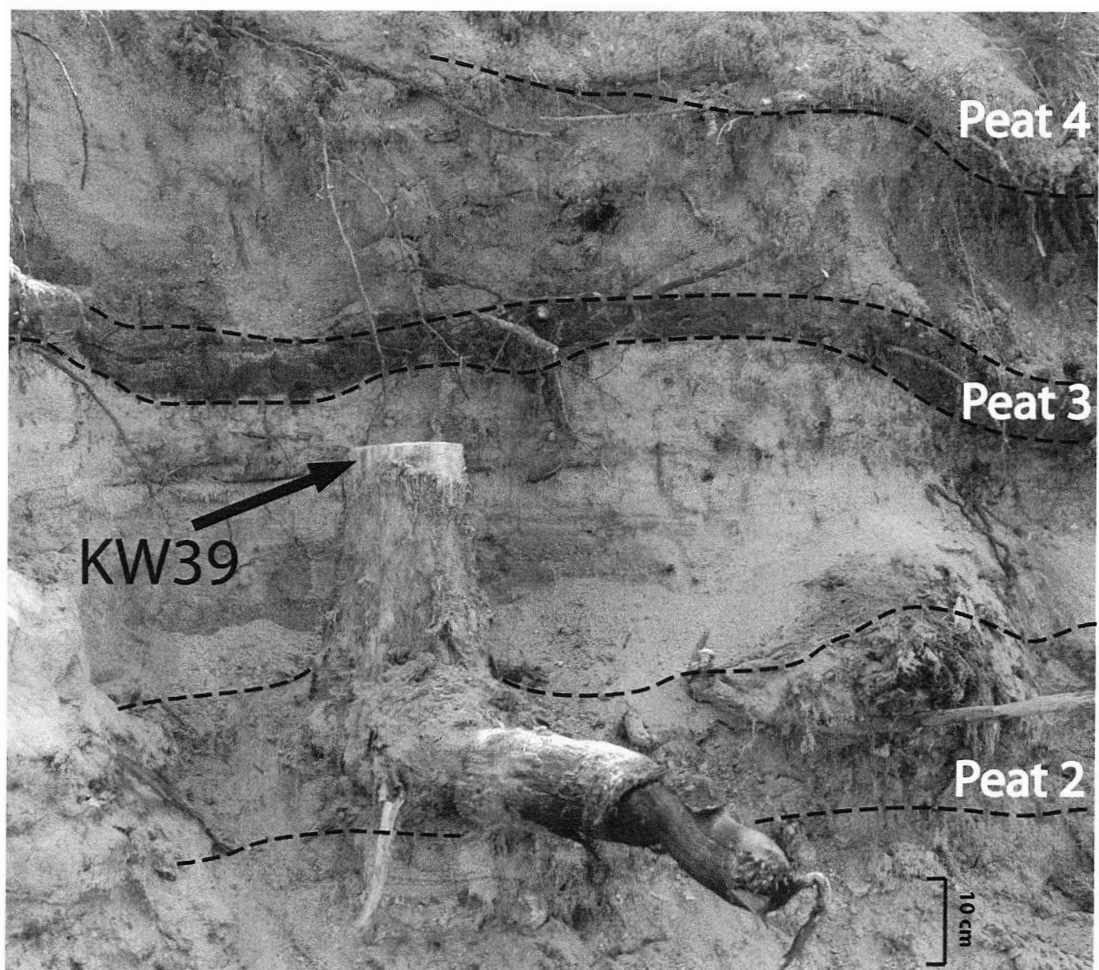


Figure 3.3 Upper 1.5 m of the section from which sample KW39 was collected. Dashed lines delineate peat layers. Sample KW39 is from a tree rooted in peat 2 (see text).

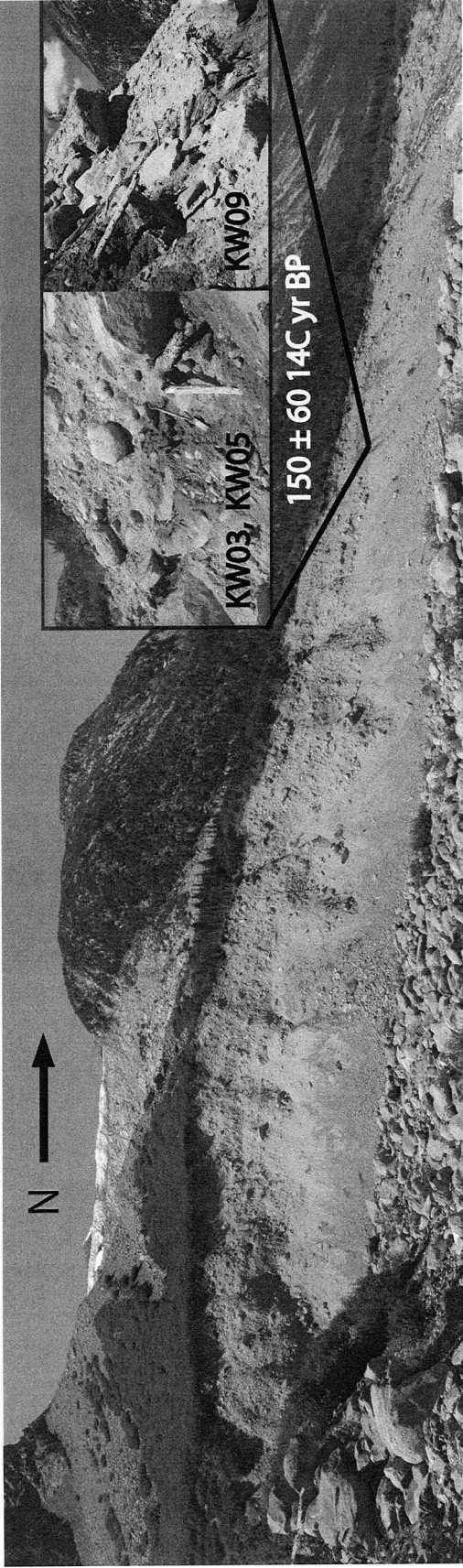


Figure 3.4 Photographic mosaic of the eroded distal face of a large moraine fan complex, 1.5 km north of Queen Bess Lake. The middle and late Holocene time. Inset photographs show logs associated with a paleosol within the fan. was built during

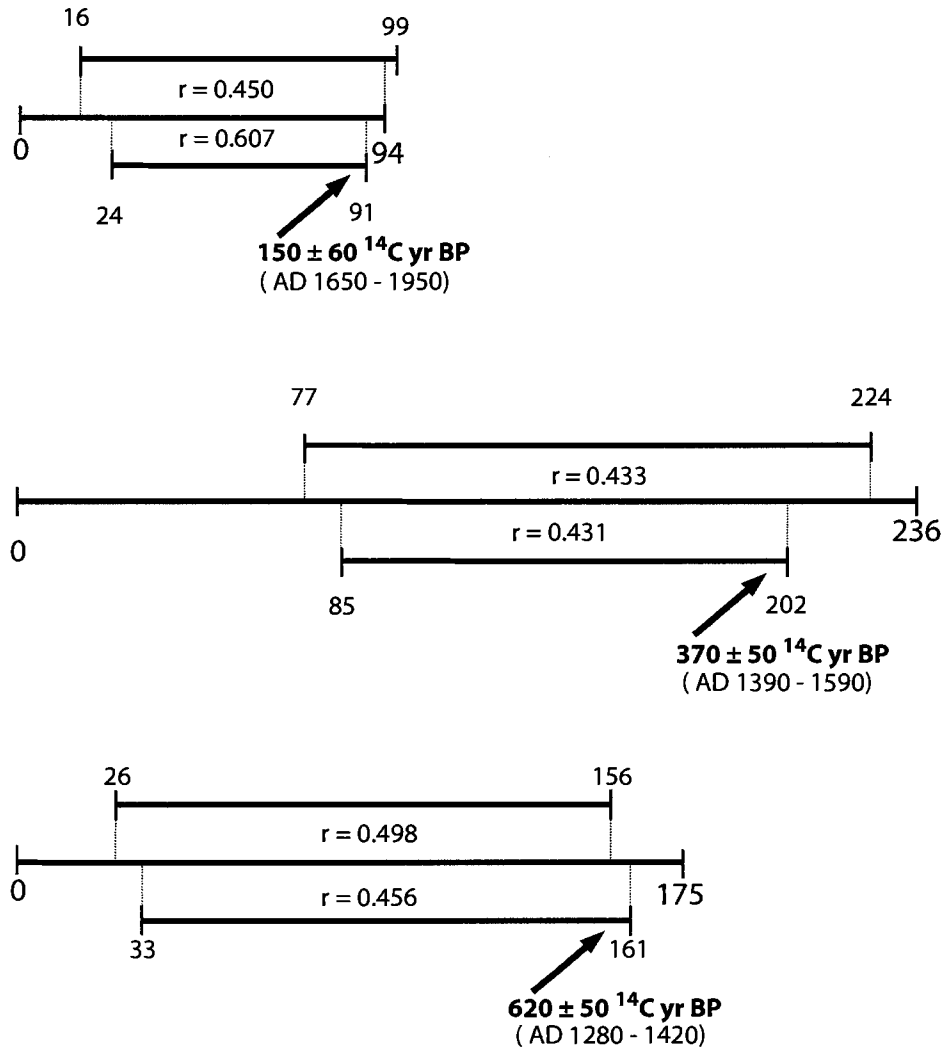


Figure 3.5 Floating tree-ring chronologies for peat layers in the west fork of Nostetuko River valley. Lengths of correlated tree-ring series, inter-series correlation values, and radiocarbon ages are shown for each peat layer.

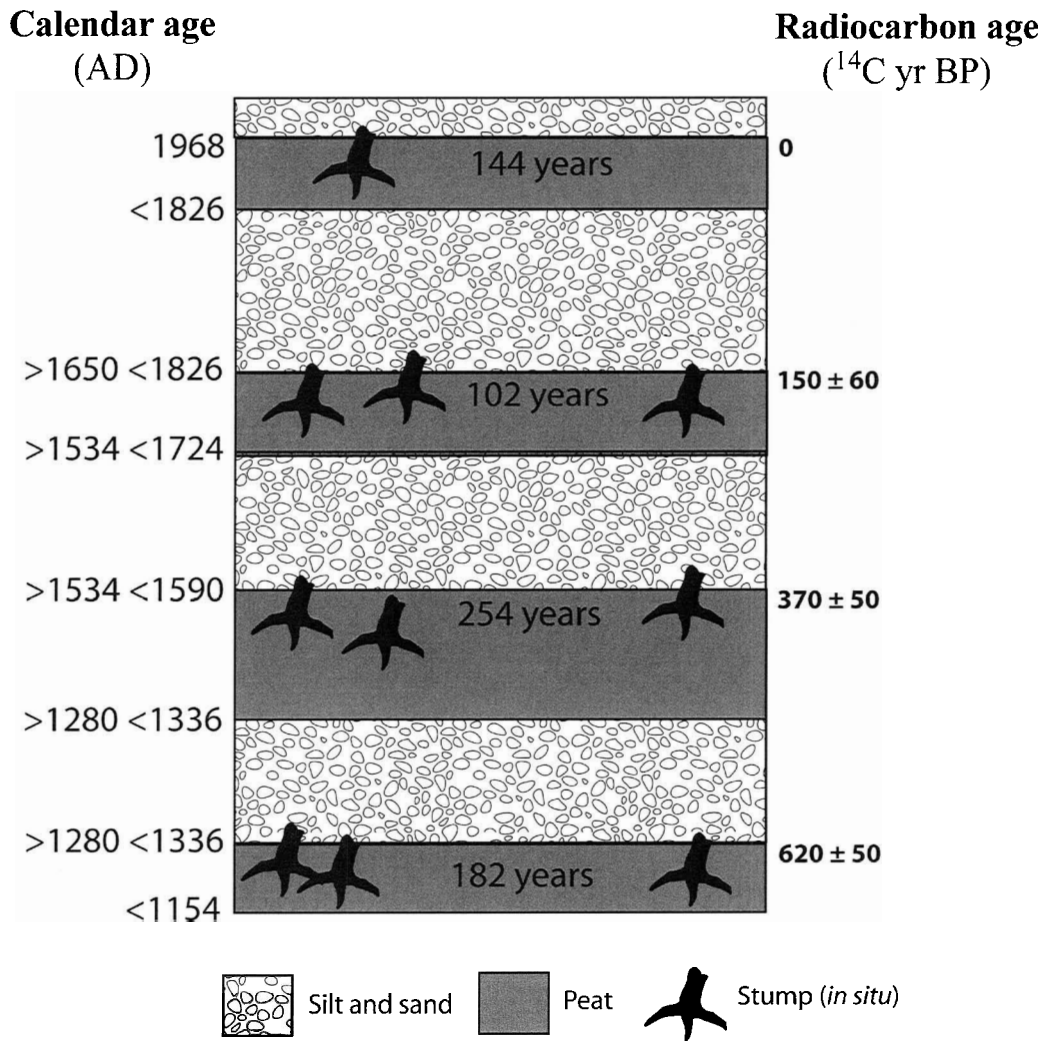


Figure 3.6 Schematic diagram showing chronological framework of late Holocene peat and clastic units in the west fork of Nostetuko River valley. Unit thicknesses are representative of those at measured sections.

3.10 Tables

Table 3.1 Summary of tree-ring statistics for samples from the study area.

Sample	Elevation (m asl)	Correction factors ^a (years)	Series length (years) ^b	N ^c	Mean sensitivity ^d	Serial correlation	First-order autocorrelation
KW14a	1396	2+15	127	3 (1)	0.301	0.527	0.775
Peat 4		17	127 (1826-1968)	1	0.883	0.438	0.413
KW03	1495	2+?	84	4 (3)	0.318	0.664	0.771
KW05	1495	2+?	68	2 (3)	0.223	0.470	0.691
KW09	1495	2+?	95	3 (3)	0.233	0.612	0.712
Peat 3		2	100	3	0.268	0.487	0.733
KW21	1493	2+0	118	2 (3)	0.356	0.671	0.744
KW38	1384	2+15	237	2 (3)	0.197	0.578	0.865
KW39	1387	2+20	148	2 (3)	0.144	0.341	0.687
Peat 2		17	237	3	0.219	0.420	0.782
KW15	1381	2+4	176	2 (3)	0.207	0.450	0.832
KW17	1381	2+8	131	3 (3)	0.224	0.577	0.733
KW18	1380	2+8	96	3 (3)	0.172	0.534	0.774
Peat 1		6	176	3	0.204	0.434	0.778

^a Ecesis estimate + sampling-height correction.

^b Number of years determined from annual growth rings. Minimum age range of floodplain stability in parentheses.

^c N is the number of radial series; number of trees used in floating chronologies in parentheses.

^d Indicator of annual variability in ring width: low sensitivity= 0.10 - 0.19; medium sensitivity= 0.20 - 0.29; high sensitivity >0.30.

Table 3.2 Radiocarbon ages bearing on floodplain stability.

Radiocarbon age (¹⁴ C yr BP) ^a	Calibrated age (cal AD) ^b	Laboratory no. ^c	Location		Elevation (m)	Material
			Lat (N)	Long (W)		
0 (104.02 ± 0.66 pMC)	1950-1960	Beta-200724	51° 17' 02"	124° 30' 30"	1480	outer rings of <i>in situ</i> stump
150 ± 60	1650-1950	TO-8932	51° 17' 10"	124° 30' 31"	1386	outer rings of <i>in situ</i> stump
180 ± 60	1640-1960	TO-8937	51° 15' 39"	124° 30' 19"	1690	root
370 ± 50	1390-1590	TO-8923	51° 16' 13"	124° 30' 11"	1493	<i>in situ</i> root
490 ± 60	1300-1620	TO-8930	51° 15' 39"	124° 30' 19"	1685	root
620 ± 50	1280-1420	Beta-200725	51° 17' 18"	124° 30' 18"	1377	outer rings of <i>in situ</i> stump

^a Ages have been corrected for natural and sputtering fractionation to a base of δ¹³C = -25.0‰.

^b Determined from atmospheric decadal data set of Stuiver et al. (1998) using the program OxCal v.3.9. The range represents 95.4% confidence limits.

^c Laboratories: Beta- Beta Analytic Inc.; TO- IsoTrace Laboratory (University of Toronto).

CHAPTER 4 CONCLUSIONS

4.1 Overview

Late Holocene glacier fluctuations produced complex downvalley fluvial responses in the Nostetuko River valley, southern Coast Mountains, British Columbia. Intervals of valley-wide aggradation correspond with periods of more extensive ice cover and independently documented glacier advances. Periods of floodplain stability and incision of the valley fill record times when glaciers were less extensive than today.

Five major periods of base-level rise and aggradation, determined from radiocarbon ages and stratigraphic analysis, coincide with independently dated glacier advances (Chapter Two). Radiocarbon ages on wood in peat layers, however, provide only maximum ages for the inception of aggradation and, by inference, glacier advance. Unavoidable uncertainties in radiocarbon ages impede precise dating of events, thus it is difficult to define the exact times within major Neoglacial 'advances' when aggradation occurred. The higher resolution afforded by dendrochronological analysis of the youngest peat layers (Chapter Three) allows more precise dating of Little Ice Age events. Therefore, at least in small, glacierized upland basins, it appears that glacier advances may increase sediment supply to the proglacial fluvial system, causing rivers to aggrade their beds.

These results have significance to questions about the sensitivity of the fluvial system to climate change and the value of proglacial fluvial sediments as proxy records of glacier fluctuations. The fluvial system has long been recognized to respond to climate change, but its sensitivity to relatively small climate perturbations on short timescales (10^1 - 10^2 yr) has been unclear. This research shows that, in at least some glacierized drainage basins, the fluvial system is sensitive to upvalley glacier fluctuations and low-magnitude climate change on centennial and perhaps shorter timescales. Fluvial sediments in these basins are a largely unexploited archive of proxy information on upvalley glacier fluctuations. Glacial histories derived from moraine records are generally incomplete, with recent and more extensive advances commonly obliterating evidence of earlier, less extensive advances. Proxy records, such as lake sediments may provide high resolution continuous records of past glacier fluctuations yet they remain distal to the ice margin and therefore subject some attenuation of the glacial signal. Glacial events probably cannot be resolved in the lacustrine record with a temporal precision better than about a century (Leonard and Reasoner, 1999). Study of proglacial fluvial sediment sequences may assist in interpreting former glacier front positions. These sediment sequences can potentially preserve and provide a more complete record of Holocene glacier fluctuations and climate change than deposits and landforms in glacier forefields.

4.2 Future work

More effort should be made to locate and study sequences of fluvial sediments in glacierized basins in British Columbia in order to test the broader significance of the results from this study. The chronology of events in the west fork of the Nostetuko River

valley could be more tightly constrained with additional radiocarbon dating and, especially, with a more thorough dendrochronological study. Larger samples of rooted trees in each of the peat layers and the development of a local living tree-ring chronology extending back several hundred to more than 1000 years would allow more precise dating of aggradation events.

REFERENCES

- Arsenault, T.A., 2004. Late Quaternary vegetation and climate change at Moraine Bog, Tiedemann Glacier, Coast Mountains, British Columbia, Canada. M.Sc. thesis, Simon Fraser University, Burnaby, BC.
- Ashmore, P.E., 1993. Contemporary erosion of the Canadian landscape. *Progress in Physical Geography* **17**, 190–204.
- Ashmore, P. and Church, M., 2001. The impact of climate change on rivers and river processes in Canada. *Geological Survey of Canada, Bulletin* **555**.
- Ballantyne, C.K., 2002. Paraglacial geomorphology. *Quaternary Science Reviews* **21**, 1935-2017.
- Ballantyne, C.K. and Benn, D.I., 1994. Paraglacial slope adjustment and re sedimentation following glacial retreat, Fabergstolsdalen, Norway. *Arctic and Alpine Research* **26**, 255–269.
- Ballantyne, C.K. and Benn, D.I., 1996. Paraglacial slope adjustment during recent deglaciation and its implications for slope evolution in formerly glaciated environments. In: *Advances in Hillslope Processes* (Eds. M.G. Anderson and S. Brooks). Wiley, Chichester, 1173– 1195.
- Beerbower, J.R., 1964. Cyclothems and cyclic depositional mechanisms in alluvial plain sediments. In: *Symposium on Cyclic Sedimentation* (Ed. D.F. Merriam). Kansas Geological Survey, Bulletin **169**, 31-42.
- Bluck, B.J., 1979. Structure of coarse grained braided alluvium. *Transactions of the Royal Society of Edinburgh: Earth Science* **70**, 181-221.
- Blum, M.D. and Törnqvist, T.E., 2000. Fluvial response to climate and sea level: A review and a look forward. *Sedimentology* **47**, 1-48.
- Brackenridge, G.R., 1988. River flood regime and floodplain stratigraphy. In: *Flood Geomorphology* (Eds. V.R. Baker, R.C. Kockel, and P.C. Patton). John Wiley & Sons, New York, 139-165.
- Brierley, G.J., 1991. Floodplain sedimentology of the Squamish River, British Columbia, relevance of element analysis. *Sedimentology* **38**, 735-750.
- Brierley, G.J., 1996. Channel morphology and element assemblages: A constructivist approach to facies modeling. In: *Advances in Fluvial Dynamics and Stratigraphy* (Eds. P.A. Carling and M.R. Dawson). John Wiley & Sons, Chichester, 263-298.
- Bronk Ramsey, C., 1995. Radiocarbon calibration and analysis of stratigraphy: The OxCal program. *Radiocarbon* **37**, 425-430.

- Bronk Ramsey, C., 2001. Development of the radiocarbon program OxCal. *Radiocarbon* **43**, 355-363.
- Brooks, G.R., 1994. The fluvial reworking of late Pleistocene drift, Squamish River drainage basin, southwest British Columbia. *Géographie physique et Quaternaire* **48**, 51-68.
- Brown, A.G., 1998. Fluvial evidence of the Medieval Warm Period and the late Medieval climatic deterioration in Europe. In: *Paleohydrology and Environmental Change* (Eds. G. Benito, V.R. Baker, and K.J. Gregory). John Wiley & Sons, Chichester, 43-52.
- Bull, W.B., 1991. *Geomorphic Responses to Climatic Change*. Oxford University Press, London.
- Calkin, P.E., 1988. Holocene glaciation of Alaska (and adjoining Yukon Territory, Canada). *Quaternary Science Reviews* **7**, 159-184.
- Calkin, P.E., Wiles, G.C., and Barclay, D.J., 2001. Holocene coastal glaciation in Alaska. *Quaternary Science Reviews* **20**, 449-461.
- Calkin, P.E., Kaufman, D.S., Przybyl, B.J., Whitford, W.B., and Peck, B.J., 1998. Glacier regimes, periglacial landforms, and Holocene climate change, Seward Peninsula, Alaska. *Arctic and Alpine Research* **30**, 154-165.
- Church, M., 1983. Pattern of instability in a wandering gravel bed channel. In: *Modern and Ancient Fluvial Systems* (Eds. J.D. Collinson and J. Lewin). International Association of Sedimentologists, Special Publication **6**, 169-180.
- Church, M.A., 2002. Fluvial sediment transfer in cold regions. In: *Landscapes of Transition: Landform Assemblages and Transformations in Cold Regions* (Eds. K. Hewitt, M.-L. Byrne, M. English, and G. Young). Kluwer Academic Publishers, Netherlands, 93-117.
- Church, M. and Ryder, J.M., 1972. Paraglacial sedimentation: A consideration of fluvial processes conditioned by glaciation. *Geological Society of America Bulletin* **83**, 3059-3072.
- Church, M. and Slaymaker, O., 1989. Disequilibrium of Holocene sediment yield in glaciated British Columbia. *Nature* **337**, 452-454.
- Church, M., Ham, D., Hassan, M., and Slaymaker, O., 1999. Fluvial clastic sediment yield in Canada: Scaled analysis. *Canadian Journal of Earth Sciences* **36**, 1267-1280.
- Clague, J.J., 1986. The Quaternary stratigraphic record of British Columbia- evidence of episodic sedimentation and erosion controlled by glaciation. *Canadian Journal of Earth Sciences* **23**, 885-894.
- Clague, J.J., 2000. Recognizing order in chaotic sequences of Quaternary sediments in the Canadian Cordillera. *Quaternary International* **68-71**, 29-38.

- Clague, J.J. and Mathewes, R.W., 1996. Neoglaciation, glacier-dammed lakes and vegetation change in northwestern British Columbia. *Arctic and Alpine Research* **28**, 10-24.
- Clague, J.J. and Mathews, W.H., 1992. The sedimentary record and Neoglacial history of Tide Lake, northwestern British Columbia. *Canadian Journal of Earth Sciences* **29**, 2383-2396.
- Clague, J.J., Turner, R.J.W., and Reyes, A., 2003. Record of recent river channel instability, Cheakamus Valley, British Columbia. *Geomorphology* **53**, 317-332.
- Clague, J.J., Wohlfarth, B., Ayotte, J., Eriksson, M., Hutchinson, I., Mathewes, R.W., Walker, I.R., and Walker, L., 2004. Late Holocene environmental change at treeline in the northern Coast Mountains, British Columbia, Canada. *Quaternary Science Reviews* **23**, 2413-2431.
- Collinson, J.D., 1996. Alluvial sediments. In: *Sedimentary Environments: Processes, Facies and Stratigraphy*, 3rd ed. (Ed. H.G. Reading). Blackwell, Oxford, 37-82.
- Curry, A.M., 1999. Paraglacial modification of slope form. *Earth Surface Processes and Landforms* **24**, 1213-1228.
- Davies, T.R.H., Smart, C.C., and Turnbull, J.M., 2003. Water and sediment outbursts from advanced Franz Josef Glacier, New Zealand. *Earth Surface Processes and Landforms* **23**, 1081-1096.
- Desloges, J.R. and Church, M., 1987. Channel and floodplain facies in a wandering gravel-bed river. In: *Recent Developments in Fluvial Sedimentology* (Eds. F.G. Ethridge and R. M. Flores). Society of Economic Paleontologists and Mineralogists, Special Publication **39**, 99-109.
- Desloges, J.R. and Ryder, J.M., 1990. Neoglacial history of the Coast Mountains near Bella Coola, British Columbia. *Canadian Journal of Earth Sciences* **27**, 281-290.
- Fenn, C.R. and Gurnell, A.M., 1987. Proglacial channel processes. In: *Glacio-Fluvial Sediment Transfer* (Eds. A.M. Gurnell and M.J. Clark). John Wiley & Sons, Chichester, 205-210.
- Fritts, H.C., 1976. *Tree Rings and Climate*. Academic Press, London, UK.
- Gardner, J.S., 1972. Recent glacial history and some associated landforms in the Canadian Rocky Mountains. In: *Mountain Geomorphology*. (Eds. H.O. Slaymaker and H.J. McPherson). Tantalus Press, Vancouver, BC, 55-62.
- Gilbert, G.K. 1877. *Geology of the Henry Mountains: U.S. geographical and geological survey of the Rocky Mountain region: U.S. Government Printing Office, Washington, D.C.*
- Gottesfeld, A.S. and Johnson-Gottesfeld, L.M., 1990. Floodplain dynamics of a wandering river, dendrochronology of the Morice River, British Columbia, Canada. *Geomorphology* **3**, 159-179.

- Gregory, K.J., 1987. The hydrogeomorphology of alpine proglacial areas. In: *Glacio-Fluvial Sediment Transfer* (Eds. A.M. Gurnell and M.J. Clark). John Wiley & Sons, Chichester, 87-107.
- Grissino-Mayer, H.D., 2001. Evaluating cross-dating accuracy: A manual and tutorial for the computer program COFECHA. *Tree-Ring Research* **57**, 205-221.
- Grove, J.M., 1988. *The Little Ice Age*. Methuen, London, UK.
- Hallet, B., Hunter, L., and Bogen, J., 1996. Rates of erosion and sediment evacuation by glaciers: A review of field data and their implications. *Global and Planetary Change* **12**, 213-235.
- Harbor, J. and Warburton, J., 1993. Relative rates of glacial and nonglacial erosion in alpine environments. *Arctic and Alpine Research* **25**, 1-7.
- Harrison, S. and Winchester, V., 1997. Age and nature of paraglacial debris cones along the margins of the San Rafael Glacier, Chilean Patagonia. *The Holocene* **7**, 481-487.
- Hebda, R.J., 1995. British Columbia vegetation and climate history with focus on 6 ka BP. *Géographie physique et Quaternaire* **49**, 55-79.
- Heikkinen, O., 1984. Dendrochronological evidence of variations of Coleman Glacier, Mount Baker, Washington, U.S.A. *Arctic and Alpine Research* **16**, 53-64.
- Hicks, D.M., McSaveney, M.J., and Chinn, T.J.H., 1990. Sedimentation in proglacial Ivory Lake, Southern Alps, New Zealand. *Arctic and Alpine Research* **22**, 26-42.
- Holbrook, J.M. Willis, B.J., and Bhattacharya, J.P., 2003. The evolution of allocyclic and autocyclic sedimentary concepts. American Association of Petroleum Geologists Annual Meeting, Expanded Abstracts, Salt Lake City, UT, **12**, 79.
- Holbrook, J.M., Scott, R.W., and Oboh-Ikuemobe, F.E., 2006. Base-level buffers and buttresses: A model for upstream versus downstream control on fluvial geometry and architecture within sequences. *Journal of Sedimentary Research* **76**, 162-174.
- Holmes, R.L., 1999. Documentation for Programs in the Dendrochronology Program Library and the Dendroecology Program Library. University of Arizona, Tucson, AZ.
- Houghton, J.T., Callendar, B.A., and Varney, S.K., 1990. *Climate Change, The IPCC Scientific Assessment*. Cambridge University Press, Cambridge.
- Karlén, W., 1976. Lacustrine sediments and tree limit variations as evidence of Holocene climatic variations in Lapland, northern Sweden. *Geografiska Annaler* **58(A)**, 1-34.
- Karlén, W. and Matthews, J.A., 1992. Reconstructing Holocene glacier variations from glacial lake sediments: Studies from Nordvestlandet and Jostedalbreen-Jotunheimen, southern Norway. *Geografiska Annaler* **74(A)**, 327-348.
- Kershaw, J.A., 2002. Formation and failure of moraine-dammed Queen Bess lake, southern Coast Mountains, British Columbia. M.Sc. thesis, Simon Fraser University, Burnaby, BC.

- Kershaw, J.A., Clague, J.J., and Evans, S.G., 2004. Geomorphic and sedimentologic signature of a two-phase outburst flood from moraine-dammed Queen Bess Lake, British Columbia, Canada. *Earth Surface Processes and Landforms* **30**, 1-25.
- Knighton, D., 1998. *Fluvial Forms and Processes: A New Perspective*. Arnold, London.
- Knox, J.C., 1995. Fluvial systems since 20,000 years BP. In: *Global continental paleohydrology* (Eds. K.J. Gregory, L. Starkel, and Baker, V.R.). Wiley, Chichester, 87-108.
- Koch, J., Menounos, B.P., Osborn, G.D., and Clague, J.J., 2003. Holocene glacial history near Whistler, southern Coast Mountains, British Columbia. XVI INQUA (International Union for Quaternary Research) Congress, Reno, NV, Program with Abstracts, 163.
- Koch, J., Osborn, G.D., and Clague, J.J., 2004a. Evidence from Garibaldi Provincial Park, British Columbia, for a complex history of Holocene glacier fluctuations in the western Cordillera of the Americas. *Geological Society of America Abstracts with Programs* **36**, 346.
- Koch, J., Menounos, B.P., Clague, J.J., and Osborn, G.D., 2004b. Environmental change in Garibaldi Provincial Park, southern Coast Mountains, British Columbia. *Geoscience Canada* **31**, 127-135.
- Larocque, S.J. and Smith, D.J., 2003. Little Ice Age activity in the Mt. Waddington area, British Columbia Coast Mountains, Canada. *Canadian Journal of Earth Sciences* **40**, 1413-1436.
- Larocque, S.J. and Smith, D.J., 2005a. 'Little Ice Age' proxy glacier mass balance records reconstructed from tree rings in the Mt. Waddington area, British Columbia Coast Mountains, Canada. *The Holocene* **15**, 748-757.
- Larocque, S.J. and Smith, D.J., 2005b. A dendroclimatological reconstruction of climate since AD 1700 in the Mt. Waddington area, British Columbia Coast Mountains, Canada. *Dendrochronologia* **22**, 93-106.
- Leonard, E.M., 1986. Varve studies at Hector Lake, Alberta, Canada, and the relationship between glacial activity and sedimentation. *Quaternary Research* **25**, 199-214.
- Leonard, E.M., 1997. The relationship between glacial activity and sediment production: evidence from a 4450-year varve record of neoglacial sedimentation in Hector Lake, Alberta, Canada. *Journal of Palaeolimnology* **17**, 319-330.
- Leonard, E.M. and Reasoner, M.A., 1999. A continuous Holocene glacial record inferred from proglacial lake sediments in Banff National Park, Alberta, Canada. *Quaternary Research* **51**, 1-13.
- Lewin, J., Macklin, M.G., and Johnstone, E., 2005. Interpreting alluvial archives: sedimentological factors in the British Holocene fluvial record. *Quaternary Science Reviews* **24**, 1873-1889.
- Lewis, D.H. and Smith, D.J., 2004. Little Ice Age glacial activity in Strathcona Provincial Park, Vancouver Island, British Columbia, Canada. *Canadian Journal of Earth Sciences* **41**, 285-297.

- Luckman, B.H., 1986. Reconstruction of Little Ice Age events in the Canadian Rocky Mountains. *Géographie physique et Quaternaire* **40**, 17-28.
- Luckman, B.H., 1988. Dating the moraines and recession of Athabasca and Dome Glaciers, Alberta, Canada. *Arctic and Alpine Research* **20**, 40-54.
- Luckman, B.H., 1993. Glacier fluctuation and tree-ring records for the last millennium in the Canadian Rockies. *Quaternary Science Reviews* **16**, 441-450.
- Luckman, B.H., 1994. Climate conditions between ca 900–1300 AD in the southern Canadian Rockies. *Climate Change* **26**, 171–182.
- Luckman, B.H., 1995. Calendar-dated, early Little Ice Age glacier advance at Robson Glacier, British Columbia, Canada. *The Holocene* **5**, 149–159.
- Luckman, B.H., 1996. Dendroglaciology at Peyto Glacier, Alberta. In: *Tree Rings, Environment and Humanity* (Eds. J.S. Dean, D.S. Meko, and T.W. Swetnam). *Radiocarbon* **38**, 679–688.
- Luckman, B.H., 2000. The Little Ice Age in the Canadian Rockies. *Geomorphology* **32**, 357-384.
- Luckman, B.H. and Osborn, G.D., 1979. Holocene glacier fluctuations in the middle Canadian Rocky Mountains. *Quaternary Research* **11**, 52-77.
- Luckman, B.H. and Villalba, R., 2001. Assessing the synchronicity of glacier fluctuations in the western Cordillera of the Americas during the last millenium. In: *Interhemispheric Climate Linkages* (Ed. V. Markgraf). Academic Press, New York, 119-140.
- Luckman, B.H. and Wilson, R.J.S., 2005. Summer temperatures in the Canadian Rockies during the last millennium: A revised record. *Climate Dynamics* **24**, 131-144.
- Luckman, B.H., Briffa, K.R., Jones, P.D., and Schweingruber, F.H., 1997. Tree-ring based reconstruction of summer temperatures at the Columbia Icefield, Alberta, Canada, A.D. 1073-1983. *The Holocene* **7**, 375-389.
- Luckman, B.H., Harding, K.A., and Hamilton, J.P., 1987. Recent glacier advances in the Premier Range, British Columbia. *Canadian Journal of Earth Sciences* **24**, 1149-1161.
- Luckman, B.H., Holdsworth, G., and Osborn, G.D., 1993. Neoglacial glacier fluctuations in the Canadian Rockies. *Quaternary Research* **39**, 144-153.
- Mackin, J.H., 1948. Concept of the graded river: *Geological Society of America Bulletin* **59**, 463–512.
- Maizels, J.K., 1979. Proglacial aggradation and changes in braided channel patterns during a period of glacier advance: An alpine example. *Geografiska Annaler* **61**, 87-101.
- Maizels, J.K., 1983. Proglacial channel systems: Change and thresholds for change over long, intermediate and short timescales. In: *Modern and Ancient Fluvial Systems* (Eds. J.D. Collison and J. Lewin). Special Publication of the International Association of Sedimentologists, Special Publication **6**, 251-266.

- Maizels, J.K., 1990. Raised channel systems as indicators of palaeohydrologic change: A case study from Oman. *Palaeogeography, Palaeoclimatology, Palaeoecology* **76**, 241-277.
- Maizels, J.K., 1995. Sediments and landforms of modern proglacial terrestrial environments. In: *Modern Glacial Environments* (Ed. J. Menzies). Butterworth-Heinemann, Oxford, 365-416.
- Marren, P.M., 2005. Magnitude and frequency in proglacial rivers: A geomorphological and sedimentological perspective. *Earth Science Reviews* **70**, 203-251.
- Mathews, W.H., 1951. Historic and prehistoric fluctuations of alpine glaciers in the Mount Garibaldi map-area, southwestern British Columbia. *Journal of Geology* **59**, 357-380.
- McCarthy, D.P., Luckman, B.H., and Kelly, P.E., 1991. Sampling height-age error correction for spruce seedlings in glacial forefields, Canadian Cordillera. *Arctic and Alpine Research* **23**, 451-455.
- Menounos, B., 2002. Climate, fine-sediment transport linkages, Coast Mountains, British Columbia, Canada. Ph.D. thesis, University of British Columbia, Vancouver, BC.
- Menounos, B., Koch, J., Osborn, G., Clague, J.J., and Mazzucchi, D., 2004. Early Holocene glacier advance, southern Coast Mountains, British Columbia, Canada. *Quaternary Science Reviews* **23**, 1543-1550.
- Miall, A.D., 1977. A review of the braided river depositional environment. *Earth Science Reviews* **13**, 1-62.
- Miall, A.D., 1978. Lithofacies types and vertical profile models in braided river deposits: A summary. In: *Fluvial Sedimentology* (Ed. A.D. Miall). Canadian Society of Petroleum Geologists, Memoir **5**, 597-604.
- Miall, A.D., 1992. Alluvial deposits. In: *Facies Models: Response to Sea-level Change* (Eds. R.G. Walker and N.P. James). Geological Association of Canada, Toronto, ON, 119-142.
- Nanson, G.C. and Croke, J.C., 1992. A genetic classification of floodplains. *Geomorphology* **4**, 459-486.
- Nesje, A., Dahl, S.O., Andersson, C., and Matthews, J.A., 2000. The lacustrine sedimentary sequence in Sygneskardvatnet, western Norway: A continuous, high-resolution record of the Jostedalbreen ice cap during the Holocene. *Quaternary Science Reviews* **19**, 1047-1065.
- Oldfield, F., 1977. Lakes and their drainage basins as units of sediment based ecological study. *Progress in Physical Geography* **1**, 469-504.
- Orwin, J.F.O. and Smart, C.C., 2004. The evidence for paraglacial sedimentation and its temporal scale in the deglaciating basin of Small River Glacier, Canada. *Geomorphology* **58**, 175-202.
- Osborn, G., 1986. Lateral moraine stratigraphy and Neoglacial history of Bugaboo Glacier, British Columbia. *Quaternary Research* **26**, 171-178.

- Osborn, G. and Luckman, B.H., 1988. Holocene glacier fluctuations in the Canadian Cordillera (Alberta and British Columbia). *Quaternary Science Reviews* **7**, 115-128.
- Osborn, G., Robinson, B.J., and Luckman, B.H., 2001. Holocene and latest Pleistocene fluctuations of Stutfield Glacier, Canadian Rockies. *Canadian Journal of Earth Sciences* **38**, 1141-1155.
- Pellatt, M. and Mathewes, R.W., 1997. Holocene tree line and climate change on the Queen Charlotte Islands, Canada. *Quaternary Research* **48**, 88-99.
- Peterson, D.W., Peterson, D.L., and Ettl, G.L., 2002. Growth responses of subalpine fir to climatic variability in the Pacific Northwest. *Canadian Journal of Forest Research* **32**, 1503-1517.
- Porter, S.C. and Denton G.H., 1967. Chronology of Neoglaciation in the North American Cordillera. *American Journal of Science* **265**, 177-210.
- Reyes, A.V. and Clague, J.J., 2004. Stratigraphic evidence for multiple Holocene advances of Lillooet Glacier, southern Coast Mountains, British Columbia. *Canadian Journal of Earth Sciences* **41**, 903-918.
- Reyes, A.V., Wiles, G.C., Smith, D.J., Barclay, D.J., Allen, S., Jackson, S., Larocque, S., Laxton, S., Lewis, D., Calkin, P.E., and Clague, J.J., 2006. Expansion of alpine glaciers in Pacific North America in the first millennium AD. *Geology* **34**, 57-60.
- Rumsby, B.T and Macklin, M.G., 1996. River response to the last neoglacial in northern, western and central Europe. In: *Global Continental Changes: The Context of Palaeohydrology* (Eds. J. Branson, A.G. Brown, and K.J. Gregory). Geological Society Special Publication **115**, 217-233.
- Rust, B.R., 1972. Structure and process in a braided river. *Sedimentology* **18**, 221-245.
- Rust, B.R., 1984. Depositional models for braided alluvium. In: *Fluvial Sedimentology* (Ed. A.D. Miall). Canadian Society of Petroleum Geologists, Memoir **5**, 605-625.
- Rust, B.R. and Koster, E.H., 1984. Coarse alluvial deposits. In: *Facies Models*, 2nd ed. (Ed. R.G. Walker). Geological Association of Canada, Geoscience Reprint Series **1**, 53-69.
- Ryder, J.M., 1987. Neoglacial history of the Stikine Iskut area, northern Coast Mountains, British Columbia. *Canadian Journal of Earth Sciences* **24**, 1294-1301.
- Ryder, J.M. and Church, M.A., 1986. The Lillooet River terraces of Fraser River: A paleoenvironmental enquiry. *Canadian Journal of Earth Sciences* **23**, 869-884.
- Ryder, J.M. and Thomson, B., 1986. Neoglaciation in the southern Coast Mountains of British Columbia: chronology prior to the late Neoglacial maximum. *Canadian Journal of Earth Sciences* **23**, 273-287.
- Sambrook Smith, G.H., 2000. Small-scale cyclicity in alpine proglacial fluvial sedimentation. *Sedimentary Geology* **132**, 217-231.
- Schumm, S.A., Mosley, M.P., and Weaver, W.E., 1987. *Experimental Fluvial Geomorphology*. Wiley, New York.

- Schweingruber, F.H., 1988. A new dendrochronological network for western North America. *Dendrochronologia* **6**, 171-178.
- Shanley, K.W. and McCabe, P.J., 1994. Perspectives on the sequence stratigraphy of continental strata. *American Association of Petroleum Geologists Bulletin* **78**, 544- 568.
- Sigafoos, R.S. and Hendricks, E.L., 1969. The time interval between stabilization of alpine glacial deposits and establishment of tree seedlings. U.S. Geological Survey, Professional Paper **650-B**, B89-B93.
- Smith, D.G., 1983. Anastomosed fluvial deposits: modern examples from western Canada, In: *Modern and Ancient Fluvial Systems* (Eds. J.D. Collinson and J. Lewin). International Association of Sedimentologists, Special Publication **6**, 155-168.
- Smith, D.J., 2003. The Garibaldi Phase revisited: Mid-Holocene glaciation in the British Columbia Coast Mountains. *Canadian Association of Geographers, Victoria, BC, Program with Abstracts*, 34.
- Smith, D.J. and Desloges, J.R. 2000. Little Ice Age history of Tzeetsaytsul Glacier, Tweedsmuir Provincial Park, British Columbia. *Géographie physique et Quaternaire* **54**, 135-141.
- Smith, D.J. and Larocque, C.P., 1996. Dendroglaciological dating of a Little Ice Age glacial advance at Moving Glacier, Vancouver Island, British Columbia. *Géographie physique et Quaternaire* **50**, 47-55.
- Smith, D.J., McCarthy, D.P., and Colenutt, M.E., 1995. Little Ice Age glacial activity in Peter Lougheed and Elk Island provincial parks, Canadian Rocky Mountains. *Canadian Journal of Earth Sciences* **32**, 579-589.
- Souch, C., 1994. A methodology to interpret downvalley lake sediments as records of Neoglacial activity: Coast Mountains, British Columbia, Canada. *Geografiska Annaler* **76A**, 169-185.
- Stokes, M.A. and Smiley, T.L., 1968. *An Introduction to Tree-ring Dating*. The University of Arizona Press, Tucson, AZ.
- Straffin, E.C. and Blum, M.D., 2002. Holocene fluvial response to climate change and human activities; Burgundy, France. *Netherlands Journal of Geosciences* **3-4**, 417-430.
- Stuiver, M., Reimer, P.J., Bard, E., Beck, J.W., Burr, G.S., Hughen, K.A., Kromer, B., McCormac, G., van der Plicht, J., and Spurk, M., 1998. INTCAL98 radiocarbon age calibration, 24000-0 cal BP. *Radiocarbon* **40**, 1041-1083.
- Törnqvist, T.E., 1998. Longitudinal profile evolution of the Quaternary Rhine-Meuse river system during the last deglaciation: Interplay of climate change and glacio-eustasy? *Terra Nova* **10**, 11-15.
- Vandenberghe, J., 1995. Timescales, climate and river development. *Quaternary Science Reviews* **14**, 631-638.

- Vandenberghe, J., 2002. The relation between climate and river processes, landforms and deposits during the Quaternary. *Quaternary International* **91**, 17-23.
- Wiles, G.C. and Calkin, P.E., 1990. Neoglaciation in the southern Kenai Mountains, Alaska. *Annals of Glaciology* **14**, 319-322.
- Wiles, G.C. and Calkin, P.E. 1994. Late Holocene high resolution glacial chronologies and climate, Kenai Mountains, Alaska. *Geological Society of America Bulletin* **106**, 281-303.
- Wiles, G.C., Calkin, P.E., and Jacoby, G.C., 1996. Tree-ring analysis and Quaternary geology: Principles and recent applications. *Geomorphology* **16**, 259-272.
- Wiles, G.C, Post, A., Muller, E.H., Molnia, B.F., 1999. Dendrochronology and late Holocene history of the Bering Piedmont Glacier, Alaska. *Quaternary Research* **52**, 185-195.
- Wiles, G.C, Jacoby, G.C., Davi, N.K., and McAllister, R.P., 2002. Late Holocene glacier fluctuations in the Wrangell Mountains, Alaska. *Geological Society of America Bulletin* **114**, 896-908.
- Williams, P.F. and Rust, B.R., 1969. The sedimentology of a braided river. *Journal of Sedimentary Petrology* **39**, 649-679.
- Wolman, M.G. and Leopold, L.B., 1957. River flood plains: Some comments on their formation. U.S. Geological Survey, Professional Paper **282-C**, 87-107.
- Wright, V.P., 1986. *Palaeosols: Their Recognition and Interpretation*. Blackwell, Oxford.

APPENDICES

Appendix 1. Locations of studied sections.

Site no.	Location		Elevation (m asl)
	Lat. (N)	Long. (W)	
1	51° 16' 09.7"	124° 30' 23.7"	1494
2	51° 16' 13.1"	124° 30' 11.3"	1493
3	51° 17' 10.7"	124° 30' 31.9"	1388
4	51° 17' 10"	124° 30' 32"	1392
5	51° 17' 13.2"	124° 30' 29.8"	1387
6	51° 17' 23.7"	124° 30' 17.6"	1408
7	51° 17' 24.8"	124° 30' 18.1"	1378
8	51° 17' 30.2"	124° 30' 04.1"	1383
9	51° 17' 37.9"	124° 30' 09.9"	1398
10	51° 18' 40.3"	124° 30' 05.4"	1294
11	51° 17' 53.0"	124° 30' 04.0"	1285

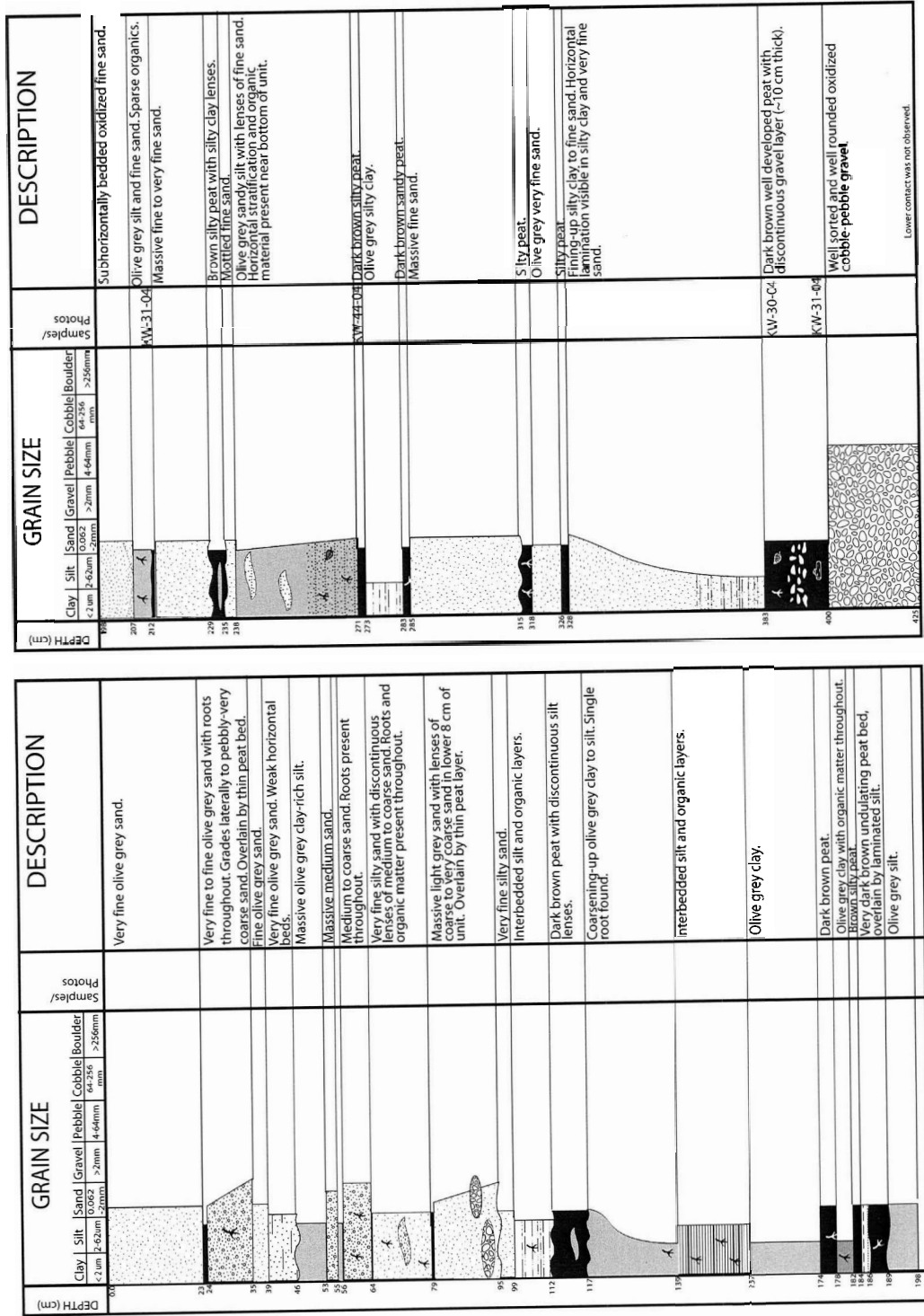
Appendix 2. Lithostratigraphic columns of studied sections.

Lithostratigraphic column for Site 3

DEPTH (cm)	GRAIN SIZE							Samples/ Photos	DESCRIPTION
	Clay <2 um	Silt 2-62um	Sand 0.062-2mm	Gravel >2mm	Pebble 4-64mm	Cobble 64-256 mm	Boulder >256mm		
12								KW-27-04	Olive grey, well sorted coarse sand.
13									
32									Massive fining-up medium to fine sand with discontinuous lenses of organic material.
35									Massive coarse sand.
50								KW-28-04	Massive mottled fine sand.
55									Medium brown fibrous peat.
67									Coarsening-up medium to fine sand.
71									Fibrous peat.
79									Massive rusty coarse sand with a sharp lower contact.
110									Olive grey massive silt with medium sand lenses. Prominent discontinuous rooting horizon at 110cm.
145									Rusty medium to coarse sand.
151									Medium sand with gradational lower contact.
155									Massive fine sand with scattered roots.
163								KW-25-04	Very coarse sand with scattered roots in upper part of unit.
170									Very fine sand and silt, massive bedding with scattered roots.
188								KW-26-04	Sharp contacts.
192									Fibrous peaty rooting horizon with abundant large roots.
198									Olive grey clayey silt.
201									Silty peat.
206									Organic-rich silt with minor small roots.
209									Silty fibrous peat.
211									Dark brown fibrous peat with abundant roots and stumps.
215								KW-24-04	Massive silt and fine sand with discontinuous root mats, organic material and fine sand lenses.
240									Horizontally stratified fine and very fine sand. Scattered roots throughout.
241									
259									Horizontally stratified fine and medium sand with scattered roots. Mottled, oxidized.
267									Fine and very fine sand, mottled appearance.
281									
283									Fining-up massive medium to very fine sand.
295									Very fine silty sand with scattered roots.
303									Oxidized fine sand with scattered roots.
304									
317									Well oxidized well sorted, rounded pebble gravel.
345									
350									Lower contact was not observed.

KW-23-04

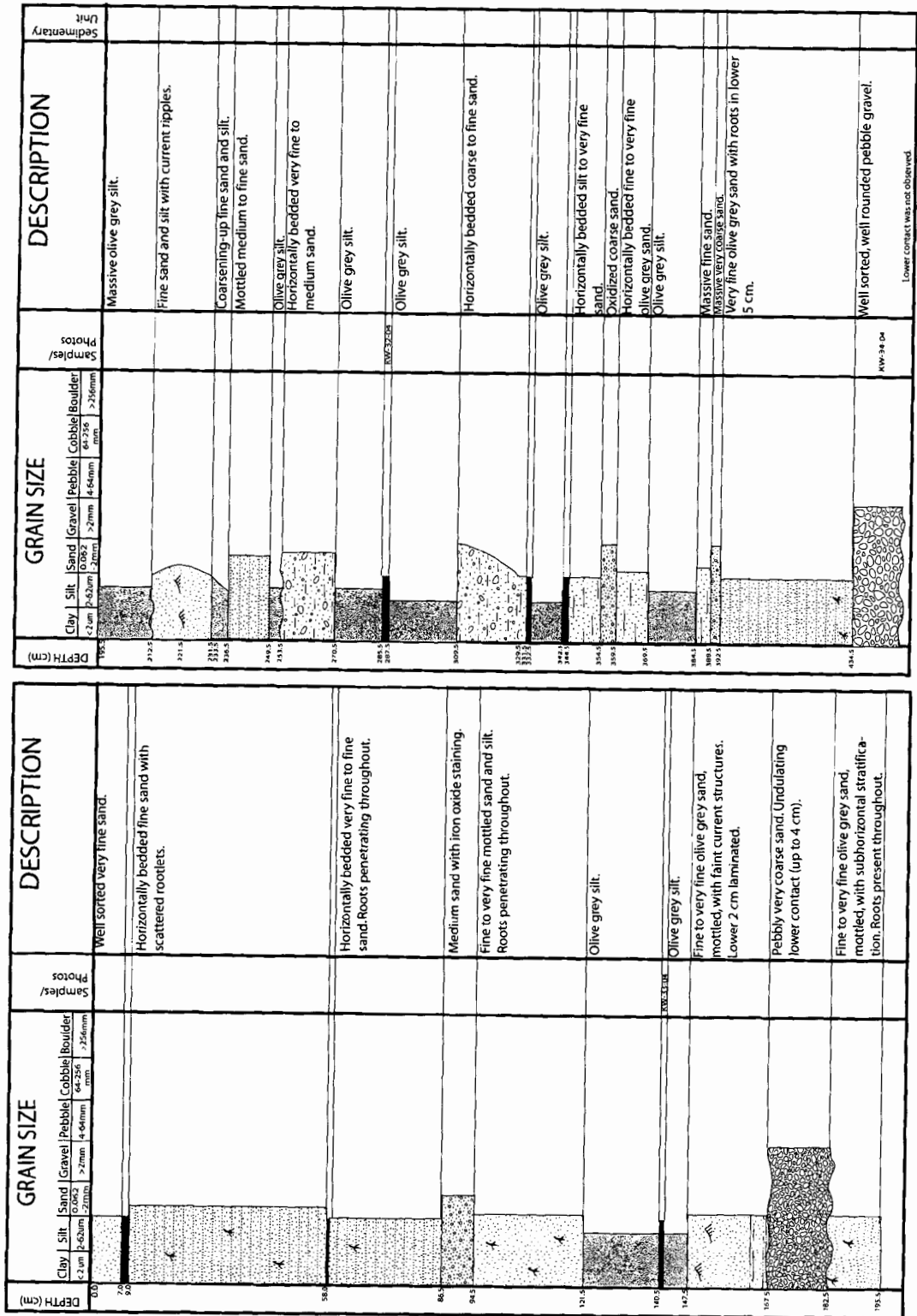
Lithostratigraphic column for Site 4



Lithostratigraphic column for Site 5

DEPTH (cm)	GRAIN SIZE							Samples/ Photos	DESCRIPTION
	Clay	Silt	Sand	Gravel	Pebble	Cobble	Boulder		
	<2 um	2-62um	0.062-2mm	>2mm	4-64mm	64-256 mm	>256mm		
0.0									Poorly sorted coarse sand and pebbles.
15									Coarsening-up fine to coarse sand with weak inclined bedding.
49									Weakly cross-bedded very coarse sand with roots.
60									Oxidized silty peat.
67									Fine to coarse sand with faint current structures.
105									Massive very fine sand.
110									Dark brown peat.
113									Light brown massive silt overlying ~1 cm of pebbles.
120								KW-39-04	Horizontally bedded very fine to fine sand.
128									Subrounded to subangular pebble gravel.
133									Well sorted medium to fine sand.
138									Subrounded to subangular pebble gravel.
142									Planar bedded very fine to fine sand with pebble gravel lenses.
150									Horizontally bedded very coarse to medium sand with abundant roots.
166									Horizontally bedded medium to fine sand.
172									Horizontally bedded fine sand.
179									Coarsening-up horizontally stratified fine and medium sand to coarse sand.
192									Coarsening-up horizontally stratified medium to coarse sand.
205								KW-41-04	Massive olive grey sandy diamicton.
263									Well sorted massive medium sand.
274									Olive grey laminated silt.
275									Dark brown peat.
281								KW-42-04	Finning-up medium to fine sand.
284									Oxidized, well rounded cobble-pebble gravel.
295									Oxidized, well rounded cobble-pebble gravel.
300									Lower contact was not observed.

Lithostratigraphic column for Site 7



Lithostratigraphic column for Site 8

DEPTH (cm)	GRAIN SIZE							Samples/ Photos	DESCRIPTION
	Clay	Silt	Sand	Gravel	Pebble	Cobble	Boulder		
	<2µm	2-62µm	0.062-2mm	>2mm	4-64mm	64-256mm	>256mm		
0.0									Cross-stratified medium to coarse white sand with current ripples.
28									Dark brown peat.
29									Massive fine sand.
32									Dark brown silty peat with abundant roots.
36									White fine sand with faint current structures and minor oxidation.
58								KW-20-04	Very dark brown rooty silty soil with <i>in situ</i> stumps and roots.
65									Coarsening-up silt and fine sand.
74									Light olive brown silty soil with roots.
77									Fine to very fine sand with current structures.
90									Brown silty soil with roots.
92									Coarsening-up silt and fine sand.
104									Coarsening-up olive grey sandy silt and fine sand.
107									Coarsening-up olive grey sandy silt and fine sand.
149								KW-19-04	Olive brown soil with abundant roots.
156									Olive grey fine sand which extends below river level.
243									Lower contact was not observed.

Lithostratigraphic column for Site 9

H (cm)	GRAIN SIZE							Samples/ Photos	DESCRIPTION
	Clay	Silt	Sand	Gravel	Pebble	Cobble	Boulder		
	<2um	2-62um	0.062-2mm	>2mm	4-64mm	64-256mm	>256mm		
0.0									Massive olive grey fine sand.
7									Light brown fibrous peat.
16									Light brown very fine to fine sand. Roots present throughout.
27									Dark brown silty peat.
30									Coarsening-up medium to coarse sand.
39									Horizontally laminated silt interbedded with light brown silty peat.
50									Olive grey silt with discontinuous lenses of dark brown peat in bottom 1 cm.
69									Fine sand fining-up to very fine sand and silt. Roots present throughout.
81									Coarsening-up medium sand to pebble gravel.
98									Very fine to fine sand with current ripples.
101									Olive grey silt.
103									Fining-up fine to medium sand with current ripples.
112									Horizontally bedded coarse to very coarse and pebbly sand.
152									Coarse sand and weakly imbricated pebble gravel. Unit is slightly oxidized with a sharp lower contact and roots present throughout.
170									Laminated olive grey silt.
173									Horizontally bedded fine sand with weak current structures. Gradational contact with underlying thin dark brown peat bed.
200									Massive olive grey silt.
211									Massive very fine sand.
214								KW-36-04	Dark brown fibrous peat.
218									Fining-up olive grey fine to very fine sand.
233									Medium sand with minor oxidation.
236									Olive grey clayey silt.
243									Lower contact was not observed.
281									

Lithostratigraphic column for Site 10

DEPTH (cm)	GRAIN SIZE							Samples/ Photos	DESCRIPTION
	Clay	Silt	Sand	Gravel	Pebble	Cobble	Boulder		
	<2 um	2-62um	0.062-2mm	>2mm	4-64mm	64-256 mm	>256mm		
0.0	[Diagram: Fine to very fine sand with current ripples]								Fine to very fine sand with current ripples.
26	[Diagram: Horizontally laminated very fine sand and silt]								Horizontally laminated very fine sand and silt.
38	[Diagram: Cross-bedded well sorted fine sand]								Cross-bedded well sorted fine sand.
54	[Diagram: Dark brown peat with abundant roots]								Dark brown peat with abundant roots.
59	[Diagram: Very fine sand and silt with parallel lamination and interbedded with brown silty peat layers]								Very fine sand and silt with parallel lamination and interbedded with brown silty peat layers.
88	[Diagram: Dark brown peat with abundant roots]								Dark brown peat with abundant roots.
100	[Diagram: Massive olive grey silt]								Massive olive grey silt.
104	[Diagram: Dark brown peat with abundant roots]								Dark brown peat with abundant roots.
108	[Diagram: Olive grey fine sand fining-up to olive grey clayey silt]								Olive grey fine sand fining-up to olive grey clayey silt.
131	[Diagram: Massive olive grey silty clay]								Massive olive grey silty clay.
149	[Diagram: Massive very fine to fine sand]								Massive very fine to fine sand.
156	[Diagram: Dark brown peat with abundant roots and in situ tree stumps]								Dark brown peat with abundant roots and <i>in situ</i> tree stumps.
173	[Diagram: Fine and very fine sand coarsening-up with lenses of medium to coarse sand in upper ~20 cm of unit]								Fine and very fine sand coarsening-up with lenses of medium to coarse sand in upper ~20 cm of unit.
220	[Diagram: Fining-up medium to very fine sand. Tree root extends through visible unit]								Finning-up medium to very fine sand. Tree root extends through visible unit.
239	[Diagram: Brown silty peat interbedded with olive grey silt]								Brown silty peat interbedded with olive grey silt.
247	[Diagram: Massive olive grey fine sand]								Massive olive grey fine sand. Unit extends below river level.
367	[Diagram: Lower contact was not observed]								Lower contact was not observed.