LATE GLACIAL LAKES OF THE THOMPSON BASIN,
SOUTHERN INTERIOR OF BRITISH COLUMBIA:
PALEOGEOGRAPHY AND PALEOENVIRONMENT

by

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Abstract

During the decay of the Cordilleran Ice Sheet (CIS), ~10 to 12 ka \(^{14}\text{C}\) BP, numerous ribbon lakes developed within the moderately deep valleys of the Interior Plateau of British Columbia. A rich geomorphic and sedimentary record of these lakes remains. This study integrates geomorphology, sedimentology, aerial photographs, differential global positioning system data, ground penetrating radar data and a digital elevation model (DEM) in a geographic information system (GIS) to (i) investigate, survey and correlate paleolake levels, (ii) reconstruct paleolake geography, evolution and environment, and (iii) reconstruct glacio-isostatic rebound.

Two definable glacial paleolake levels were identified, associated with Glacial Lake Thompson and Glacial Lake Deadman. DEMs of paleolake levels, inferred lake bottom and modern topography were integrated in a GIS to quantify lake parameters. Lakes were ribbon-shaped (width to length ratio of \(~3:100\) ), deep (\(~140\) and \(~50\) m, respectively), and of significant volumes (84 and 24 km\(^3\), respectively). Glacio-isostatic tilts of these lake shorelines (\(1.8 - 1.7\) m km\(^{-1}\)) are among the highest measured in the world and are related to a thin lithosphere, a low viscosity mantle and rapid deglaciation. Glacio-isostatic depression in the interior was likely hundreds of metres.

The sedimentary record of these lakes reflects the adjustments of a landscape undergoing deglaciation. Seventeen glaciolacustrine lithofacies were identified and record deltas, subaqueous fans, high rates of sedimentation, numerous hyperpycnal flows and a diversity of sediment dispersal and deposition processes. High sedimentation rates and numerous hyperpycnal flows suggest that ribbon lakes likely received their meltwater and sediment supply from ice remnant on the plateau.

Glacial Lake Deadman drained catastrophically with the breach of an ice dam, producing drainage bedforms and erosional surfaces within the basin, and discharging \(~20\) km\(^3\) of water. It is possible that this event may have triggered the failure of glacial lakes downstream or upstream in the Fraser River system. Eventually the floodwaters reached the Strait of Georgia, a distance of \(~250\) km. Here exotic sediments dated between \(~9,200\) \(^{14}\text{C}\) yr BP and \(10,800\) \(^{14}\text{C}\) yr BP may record this jökulhlaup.
Dedicated to all who are brave enough to learn about the world and ultimately their place within it.
Geographers are interested in exploring the reasons for inter-relationships in space. Why does that cliff become steep here and not there? Why does the river turn like this? How did human development occur in that particular way? Geography is like a detective story, with the answers laid out before us, if only we can read the clues.

Robert Bateman

I have always used Chamberlin’s method of multiple working hypotheses. I applied Ockham’s razor to select the most appropriate hypothesis, but always with due regard for possible dull places in the tool.

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While my degree is in physical geography the people of the Ashcroft area reminded me that the other half of geography is human geography. The traveller's spirit in me was satisfied. A few particularly stand out in memory (sorry for any spelling errors): Bill "landslide" Stewart, the whole Schalles family, Kevin "Ginseng/Shark" Hughes, Al "Big Al" Midley, Anita Dick and associates (Ashcroft Indian Band), Chief David Walkem, Earl Smith and Pearl Hewitt (Cook's Ferry Indian Band), Chief Robert "You're lucky I didn't hit you with my bucket!" Pasco (Oregon Jack Reserve), Chief Ron Ignace and Mary Anne Ignace, and Doug Brown (Deadman Indian Band), Tracy Ens and Ken Anderson, Mark Stoelwinder (WASTECH), Barry Tait (The Journal), Jack and Charlie Christian (taller than me!), Marcus Lowe and Trish (Namaste!), Joe Paulos (CNR), Tracy Murdock, Wayne (Gardner Ranch), John Zahradnik and Mariette Luce, Claire Louis (former land lady for Lesley Anderton in 1969!), Paul Ford (Ashcroft Ranch), Edward Villiers (I love kettles too Ed), Jim Farmer and family, Trevor and Flora Parker (my apologies for being too busy to have a visit), Joe Valentinuzzi (MoF), and Mareah "I think the river was up here" Franes. Thank you for your hospitality. My apologies for those I forgot.
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Introduction
Chapter 1

Introduction

1.1. Introduction

Proglacial lakes were common in British Columbia during Late Wisconsinan deglaciation of the Cordilleran Ice Sheet (CIS; Eyles and Clague 1991, Ryder et al. 1991). They have mostly been inferred from regional mapping (e.g. Mathews 1944, Fulton 1969, Clague 1987) or stratigraphic studies (e.g. Fulton and Smith 1978). Within the southern interior of British Columbia, regional mapping has provided information on the paleogeography of many lakes (e.g. Mathews 1944, Fulton 1965, 1969). The temporal context of the general depositional environment has been inferred from stratigraphic studies (e.g. Fulton and Armstrong 1965, Fulton and Smith 1978, Ryder 1976, Clague 2000). To date, sedimentological research has been scarce (e.g. Fulton 1965, Shaw 1977, Shaw and Archer 1979). Sedimentological, paleogeographic and stratigraphic knowledge can inform our understanding of the environmental controls on lake sedimentation and thus deepen our understanding of deglaciation.

This project seeks to refine the paleogeography of late glacial lakes in part of the Thompson Basin of the southern interior of British Columbia and to investigate the environmental controls on lake sedimentation. Specifically, the research objectives are:

1. To determine lake paleogeography, and

2. To investigate the paleoenvironmental controls on sedimentation.

More generally, this research expands our general knowledge of deglacial lake processes in the British Columbia Interior, and extends our understanding of deglaciation and isostatic rebound. The Thompson Basin is a good candidate for glaciolacustrine study as, (1) the stratigraphic context is known (Fulton and Armstrong 1965, Ryder 1976, Clague 2000), (2) a number of exposures with easy access are found throughout the valley, (3) little is known regarding both the paleogeography and sedimentary environments of lakes within the western portion of the basin, and (4) the lakes record deglaciation near the centre of the CIS (Prest et al. 1968, Clague 1989, Ryder et al. 1991). As well, this project offers a unique opportunity to integrate knowledge of
landforms, stratigraphy, sedimentology and digital terrain data. Today, the availability of
digital elevation models (DEMs), differential global positioning systems (DGPS), ground
penetrating radar (GPR) and geographic information systems (GIS) makes more detailed
explorations of deglacial lakes an exciting possibility. This study is the first integrated
application of these technologies for glacial lake reconstruction in the Cordillera.

This chapter introduces the study area and briefly discusses present knowledge of
(1) Quaternary history of the Thompson Valley, (2) glacial lake paleogeography and
evolution in the Thompson Basin, (3) glacial lake environments, processes and products,
and (4) the research project strategy.

1.2. Study area

Paleogeographic research was completed in the Thompson Basin (Thompson,
South Thompson, and the lower North Thompson river valleys, Fig. 1.1). The study area
is bordered by the communities of Chase to the east, Lytton to the southwest, and
McClure to the north (Fig. 1.1b). Altogether this area includes 220 km of valleys. Within
this area research was concentrated along the Spences Bridge-Ashcroft-Kamloops valley
corridor (Fig. 1.1a). Not included in this study, and located within the Thompson Basin, is
the Nicola River Valley and the Merritt Basin (Fig. 1.1b), previously studied by Anderton
(1970) and Fulton and Walcott (1975), respectively, and the upper North Thompson
Valley north of McClure.

Paleoenvironmental research was concentrated along a 75 km reach of the
Thompson Valley between the outlet of Kamloops Lake (Savona) and the confluence
of the Nicola and Thompson rivers (Spences Bridge) and is centred about the Village of
Ashcroft (Fig. 1.1a). The area of paleoenvironmental research is confined to the
Thompson River Valley as (1) this area has large cliffs (up to 100 m high) of deglacial
lacustrine sediments that occur in a continuous swath for ~75 km, and (2) it is the only
portion of the Thompson Basin containing such large cliffs that have not been studied in
detail.

The study area lies on the western portion of the Thompson Plateau physiographic
region, within the rain shadow of the Coast Mountains to the west, and neighbours the
Fraser Plateau to the northwest and the Cascade Range to the south (Fig. 1.1b; Holland
1964). Fluvial and glacial erosion have dissected the Thompson Plateau resulting in deep
valleys (1400 to 1600 m relief), today filled with thick (100 to >150 m) Quaternary
deposits.
Figure 1.1: The study area. (a) The Spences Bridge-Kamloops corridor. (b) The regional context of the study area. (c) Location of study area in British Columbia, Canada. Li = Lillooet, Ly = Lytton, SB = Spences Bridge, M = Merritt, Ash = Ashcroft, CC = Cache Creek, K = Kamloops, Mc = McClure, Ch = Chase, SA = Salmon Arm, Ar = Armstrong, and V = Vernon. Th.R = Thompson River, N. Th.R = North Thompson River, and S. Th.R = South Thompson River.
1.3. Quaternary history of the Thompson Valley

Multiple glacial and non-glacial intervals have been proposed for the British Columbia Cordillera (Tipper 1971, Fulton and Smith 1978, Clague et al. 1987). The Late Wisconsinan glaciation is inferred to have been initiated by the expansion and coalescence of alpine glaciers that advanced across the plateaus and eventually grew into large confluent masses that covered most of British Columbia (Ryder et al. 1991). During the final phase of glaciation, ice further thickened to form domes with regional surface flow unconstrained by topography (Dawson 1881, 1891, Kerr 1934, Mathews 1955, Wilson et al. 1958, Fulton 1967, Flint 1971, Stumpf et al. 2000).

During the climax of the last glaciation (Fraser) a large ice mass covered the entire British Columbia Cordillera (Davis and Mathews 1944, Ryder et al. 1991). In the Interior, deglaciation was characterized by downwasting of plateau areas leaving ice-tongues in valleys (Fulton 1967, 1991, Clague 1981) and forming a number of ice- and sediment-dammed lakes (Mathews 1944, Armstrong and Tipper 1948, Fulton 1967, 1969). This study focuses on the Late Wisconsinan (Fraser) deglacial lakes of the Thompson Basin (Fig. 1.2).

Understanding of Quaternary history in the Thompson Valley, west of Savona (Fig. 1.1), is based on limited studies of stratigraphy, landforms and surficial geology mapping (Fulton and Armstrong 1965, Ryder 1976, 1981, Clague 2000, Clague and Evans 2003). The stratigraphic record is one of repeated cycles of glacial, glaciolacustrine and non-glacial fluvial events. The ages of these events are poorly constrained for most of the British Columbia Interior (Lian 1997).

Sediments associated with the late glacial lakes of the Thompson Valley occur near the top of the preserved glacial sequence, overlie glacial sediments and underlie non-glacial Holocene sediments (Fig. 1.3). This stratigraphic relationship confirms the inference that these lakes were associated with the retreat phase of the last glaciation. Holocene events included (1) the development of alluvial fans that frequently built upon Holocene river terrace gravel overlying lacustrine sediments (Ryder 1970), and (2) fluvial incision >100 m producing large (>100 m tall) cliffs of deglacial lacustrine sediment.

East of Savona, sediments older than the last glacial maximum have been mapped in a few locations (Fulton 1975). Exposure descriptions have included tephras and bones of several extinct mammals of pre-Fraser Glaciation age (Fulton and Smith 1978, Fulton 2000). The most dominant deposit within the eastern portion of the study area is Late Wisconsinan (Fraser) deglacial lake sediment. In the South Thompson Valley, cliffs of glaciolacustrine silt tower over 100 m above the present river level and dominate
Figure 1.2: Evolution of deglacial lakes in the southern interior of British Columbia as proposed by Fulton (1969) and Ryder (1981). Regional stagnation of the Cordilleran Ice Sheet led to the development of glacial lakes impounded by valley-occupying ice masses. Lake basins lengthened coeval with recession of the fronts of these ice masses. Two glacial lakes and four glacial lake stages are inferred (a-d). Present drainage shown in (e).
Figure 1.3: The stratigraphic context of lacustrine sediments (unit 6) examined in this study. Stratigraphic position implies that these lacustrine sediments are deglacial in origin (Clague 2000, Clague and Evans 2003) following the Fraser Glaciation (oxygen isotope stage 2). This section records an inset valley-fill stratigraphy. Elsewhere, unit 6 sediments typically occur in units much thicker (>80 m) than at this section, and form the dominant valley fill.
the landscape. Cliff exposures and well logs indicate that deglacial lake sediments exceed 140 m in thickness (Fulton 1965).

1.4. Glacial lake paleogeography and evolution

The paleogeography and evolution of glacial lakes in the Thompson Basin have been inferred from (1) the spatial distribution of lake and glacial sediments and landforms, (2) the elevations of deltas, river terraces, meltwater channels, and ice-marginal deposits and landforms, and (3) the isostatic tilt of shorelines (Mathews 1944, Fulton 1965, 1967, 1969). Water plane features were correlated using a simple 3-point geometric method along linear valleys (R.J. Fulton personal communication 2001).

Previous research presents a generalized picture of lake evolution for the southern interior of British Columbia. Four stages of two eastward draining deglacial lakes, Glacial Lake Thompson and Glacial Lake Deadman, are defined in the study area (Fig. 1.2; Fulton 1969, Ryder 1976, 1981). Changes in lake name imply a change in outlet position; stage names define stillstands (Fulton 1969).

The lakes and stages are characterized by progressive westward lake extension and relocation, and lake level lowering. The South Thompson Valley between Chase and Kamloops became ice-free first and was occupied by Glacial Lake Thompson - South Thompson stage dammed by ice at both ends (Fig. 1.2a). This lake lengthened and lowered forming Glacial Lake Thompson - Niskonlith stage (Fig. 1.2b) as ice tongues at its western and eastern ends and in the lower North Thompson Valley receded. With continued westward ice recession and further lake level lowering, high elevation lake bottom sediments in the South Thompson Valley became exposed, separating lakes in the Shuswap Basin from those in the Thompson Basin by an eastward draining river. This new outlet, east of Kamloops, defines Glacial Lake Deadman - Tranquille stage (Fig. 1.2c) and with further lake lowering and southward growth, Glacial Lake Deadman - Durand stage (Fig. 1.2d). Following lake drainage, the Thompson River reversed course and flowed south (Fig. 1.2e). In the early Holocene, a high stage of Kamloops Lake called Cherry Creek stage (not shown in Fig. 1.2) developed behind raised delta sediments at Savona.

There are no dates available that define specific lake stages (Fulton 1969). All glacial lakes were drained no later than 8,900 ±150 radiocarbon years (GSC-193, organic silt, Dyck et al. 1965, Fulton 1969). This date is from a spillway near Armstrong that once connected the Shuswap and Okanagan Basins (~200 km east of Ashcroft, Fig. 1.2).
1.4.1. Paleogeography: research questions and methods

Knowledge of the paleogeography of Glacial Lake Thompson and Glacial Lake Deadman is inexact and requires refinement. Refinement of the paleogeography requires the following questions to be answered:

- What was the geometry (i.e. areal extent, elevation, depth and isostatic-tilt) of Glacial Lake Thompson and Glacial Lake Deadman?
- How did the lake geometries change through time?
- How and where were the lakes dammed?
- What was the style of final lake drainage?

Methods employed to answer these questions include (1) aerial photograph interpretation and field mapping of landforms and sediments, (2) field surveying to determine accurate elevations and possible glacio-isostatic tilting, and (3) the integration of this information with a digital elevation model to determine lake geometry and visualize relationships between lake stages and associated landforms and sediments. Field surveying involved the use of a differential global positioning system (DGPS) to accurately survey features and a ground penetrating radar (GPR) system to assist in inference of landform genesis.

1.5. Glacial lake environments, processes and products

In order to infer depositional processes and environments from the character of glacial lake sediments it is important to (1) consider the various types of glacial lake environments, and (2) understand the relationships between processes and products.

Glacier lakes are defined as lakes fed either directly or indirectly by glacier meltwater (Smith and Ashley 1985). Thus glacial lakes can form on top, below, in front, and beside a glacier (Table 1.1). Studies of modern glacial lakes have provided a sound understanding of the linkages between environments, processes and products. This understanding provides the basis for inferring environments and processes from the sediments and landforms of past glacial lakes. Controls on glacial lake sedimentation and the sediments and landforms associated with each type of glacial lake are summarized in Table 1.2.

The present understanding of glacial lake environments depends on research largely completed on proglacial lakes. It seems likely that Glacial Lake Thompson and Glacial Lake Deadman were proglacial lakes due to the presence of deltas, and wave-cut
benches and stratigraphic position (Fulton 1969, Ryder 1976, 1981, Clague 2000). Therefore, the remaining discussion focuses on the proglacial lake environment.

Table 1.1: Classification of glacial lake types including possible containment mechanisms (after Shaw 1977, Smith and Ashley 1985, Eyles et al. 1987)

<table>
<thead>
<tr>
<th>Type</th>
<th>Other names</th>
<th>Position with respect to glacier</th>
<th>Dam type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Proglacial lake</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ice-contact</td>
<td></td>
<td>At the front or side of glacier</td>
<td>Topography, moraine, and/or ice</td>
</tr>
<tr>
<td>Ice-frontal</td>
<td>Ice-dammed, Ice-proximal</td>
<td>Front</td>
<td>Topography, moraine, and/or ice</td>
</tr>
<tr>
<td>Ice-marginal</td>
<td>Ice-dammed</td>
<td>Side</td>
<td>Topography, moraine, and/or ice</td>
</tr>
<tr>
<td>Non-ice-contact</td>
<td>Glacier-fed, Ice-distal</td>
<td>Separated from glacier by outwash or another glacial lake</td>
<td>Topography and/or moraine</td>
</tr>
<tr>
<td>Supraglacial lake</td>
<td>Ice lake, Ice cauldron</td>
<td>On top of glacier</td>
<td>Ice topography</td>
</tr>
<tr>
<td>Supramarginal</td>
<td>Ice/land lake</td>
<td>Straddles glacier surface and land</td>
<td>Topography, moraine, and/or ice</td>
</tr>
<tr>
<td>Englacial reservoir</td>
<td></td>
<td>Reservoir within a glacier</td>
<td>Ice</td>
</tr>
<tr>
<td>Subglacial reservoir</td>
<td>Water cupola</td>
<td>At bed of glacier</td>
<td>Bed topography, cold ice, high pressure</td>
</tr>
</tbody>
</table>

Numerous factors control sedimentation in proglacial lake environments: basin depth and geometry, sources of water, timing and intensity of water input, sources of sediment, lake water density stratification, lake circulation, littoral processes and subaqueous mass wasting (Table 1.2; Gilbert 1975, Smith 1978, 1981, Smith et al. 1982, Gilbert and Desloges 1987, Gilbert et al. 1997, Desloges and Gilbert 1998). Water inputs can vary on many timescales: decadal, annual, seasonal, diurnal and episodically (Smith 1978). Occasionally, high discharge events can occur related to weather events (Gilbert et al. 1997) or jökulhlaups (Clague and Evans 1997, Gilbert et al. 1997, Gilbert and Desloges 1987). Proglacial ice-contact lakes may drain suddenly via subglacial or englacial conduits (Gilbert and Desloges 1987, Liverman 1987), or by the breaching of moraines (Gilbert and Desloges 1987) or other sediment piles (e.g. deltas, alluvial fans).
Table 1.2: Relative significance of controls on glacial lake sedimentation and of sediments and landforms for different glacial lake types.

<table>
<thead>
<tr>
<th>CONTROLS ON SEDIMENTATION</th>
<th>Proglacial</th>
<th>Ice-contact</th>
<th>Non-ice-contact</th>
<th>Supraglacial²</th>
<th>Subglacial</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Depth and Geometry</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Shallow</td>
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<td>present</td>
<td>present</td>
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</tr>
<tr>
<td>Deep</td>
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</tr>
<tr>
<td>Long</td>
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<td>present</td>
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<td>present</td>
<td>present</td>
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<td><strong>Water sources</strong></td>
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<td></td>
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<td></td>
</tr>
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<td>present</td>
<td>present</td>
<td>present</td>
<td>present</td>
</tr>
<tr>
<td>Groundwater</td>
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<td>present</td>
<td>present</td>
<td>present</td>
</tr>
<tr>
<td>Subglacial</td>
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<td>absent</td>
<td>present</td>
<td>present</td>
<td>present</td>
</tr>
<tr>
<td>Englacial</td>
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<td>absent</td>
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<td>present</td>
<td>present</td>
</tr>
<tr>
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<td>absent</td>
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<td>present</td>
<td>present</td>
</tr>
<tr>
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<td>present</td>
</tr>
<tr>
<td>Geothermal melting</td>
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<td>Upstream glacial lakes, rivers</td>
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<td>Non-glacial rivers</td>
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</tr>
<tr>
<td><strong>Timing of water input</strong></td>
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<td></td>
<td></td>
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<td><strong>Sediment sources</strong></td>
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<td>Atmospheric (aeolian)</td>
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<td>Groundwater (precipitates)</td>
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</tr>
<tr>
<td>Valleyside streams</td>
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<td>present</td>
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<td>absent</td>
</tr>
<tr>
<td>Rockfalls</td>
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<tr>
<td>Subaqueous gravity flows</td>
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</tr>
<tr>
<td>Turbidity flows</td>
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</tr>
<tr>
<td>Hyperconcentrated flows</td>
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</tr>
<tr>
<td>Debris flows</td>
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<tr>
<td><strong>Ice derived</strong></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Ice front (dump and flow)</td>
<td>present</td>
<td>absent</td>
<td>present</td>
<td>present</td>
<td>absent</td>
</tr>
<tr>
<td>Iceberg (dump and rainout)</td>
<td>present</td>
<td>absent</td>
<td>absent</td>
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<td>absent</td>
</tr>
<tr>
<td>Loeself (rainout)</td>
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<td>absent</td>
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<td>Direct meltout</td>
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<tr>
<td>Lake ice rainout (colluvial)</td>
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¹ "Present" indicates that the process, landform or sediment may be present; "absent" indicates that the process, landform or sediment is not be present. Derived from inference and numerous sources: Ashley 1975; Clague and Evans 1997; Desloges and Gilbert 1996; Fulton 1987; Gilbert 1975; Gilbert and Desloges 1987; Gilbert and Shaw 1981; Gilbert et al. 1997; Gustavson 1975; Huntley and Broster 1994; Leonard 1985; Liverman 1987; Munro-Stasiuk 1999; Østrem and Olsen 1987; Pickrill and Irwin 1983; Powell 1990; Shaw et al. 1978; Shaw 1977; Smith 1978; Smith 1981; Smith et al. 1982; Smith and Syvitski 1982

² The environmental settings of supraglacial lakes can vary greatly (e.g. located on the interior of ice sheets, located in mountain valleys)
Patterns of sediment transport and water flow in lakes are strongly dependent on the vertical temperature profile, or thermal structure, of the lake water (Gilbert 1975, Smith et al. 1982). Occasionally sediment (Gustavson 1975, Smith 1981) or chemical stratification (Smith 1978) can be significant. The relationship between the densities of water and sediment input and of basin water affect the dispersal of sediments by: overflow (hypopycnal flow), interflow and underflow (hyperpycnal flow; Gilbert 1975). Underflows can also be initiated by subaqueous mass wasting events (Gilbert 1975). Secondary sources of sediment derived from tributary streams can cause cross-lake trends in lake sedimentation. These trends can be superimposed on downlake proximal-distal trends related to a glacier front (Ashley 1975, Pickrell and Irwin 1983, Desloges and Gilbert 1998).

Three distinct sedimentary environments are related to basin geometry, ice position and inflow position, (1) proximal sedimentation (deltas, fans, and ice-contact and ice-shelf deposits), (2) lake bottom sediments (rhythmites, turbidites, mass wasting and iceberg dumping structures), and (3) littoral sediments and landforms (erosional benches, and beaches; Gilbert 1975, Smith et al. 1982, Pickrell and Irwin 1983).

Sedimentary structures found in the proximal and lake bottom environments can be used to infer energy conditions (i.e. hydrodynamics, rate of sedimentation, etc.) at the time of deposition (Allen 1982, Ashley et al. 1982). Temporal and spatial patterns of these structures can be used to infer energy patterns that reflect changes in the controls of sedimentation (i.e. retreating ice-tongue, rhythmic water inputs, etc.; Ashley 1975, Gilbert 1975, Desloges and Gilbert 1998). Turbidity currents can reflect temporal (vertical in section) changes in flow conditions (Bouma 1962, Middleton and Hampton 1976). These events are expected to be common for the environments of glacial lakes in the Thompson Valley given the high-energy conditions of deglaciation, moderately high relief and the existence of numerous tributaries. Deformation of sediments can produce a number of structures (e.g. dewatering, loading and slumping structures, etc.) that can be used to further refine understanding of the depositional environment (Allen 1982).

Various environmental controls determine the temporal and spatial patterns and character of sediments for each of the three sedimentary environments. By examining the character of Glacial Lake Thompson and Glacial Lake Deadman sediments and landforms, it may be possible to infer formative processes and environmental controls on sedimentation.
1.5.1. Environmental controls: research questions and methods

Stratigraphic work (Fulton and Armstrong 1965, Ryder 1976, Clague 2000) has clearly demonstrated that Glacial Lake Thompson and Glacial Lake Deadman were associated with the retreat phase of the last glaciation (Fig. 1.3). As well, observations of deltas and an inferred westward and then southward retreating ice-tongue suggest that both tributary and ice-front sedimentation patterns are likely (Fulton 1969, Ryder 1976). However, very little is known of the sedimentary environments and related environmental controls on sedimentation for these lakes. A number of research questions need to be answered:

- What were the styles of sediment deposition?
- What were the sediment transport dispersal mechanisms?
- What were the sources of water and sediment?
- Are there spatial patterns in lake sedimentation?
- Are there temporal patterns in lake sedimentation?
- What was the paleohydrologic regime of the lake?
- Can tributary sedimentation patterns be divorced from ice-front influxes?
- Do conclusions corroborate and/or extend our understanding of glacial history around Ashcroft?

Methods employed to answer these questions included standard field sedimentology methods and grain size analysis. Sedimentological analysis included observations of texture, structure, paleoflow, bed contact relationships, lateral continuity and thickness. Where centimetre-decimetre observation was not possible (i.e. very large and/or unsafe sections) sections were mainly described in a stratigraphic manner with portions described in detail. Grain size analysis was conducted at selected sites to confirm the texture of apparently clayey sediments. Radiocarbon dating was not possible as no organic material was found. Previous research has shown that the British Columbia Interior is essentially barren of preserved organics (Lian et al. 1999).
Thesis format

This thesis has been written in paper format. Each substantive chapter is written in a stand-alone style and as a result the reader will find some information between chapters repetitive. Chapter 2 discusses the paleogeography, evolution and glacio-isostatic significance of late glacial ribbon lakes. Chapter 3 discusses the environment in and around late glacial ribbon lakes. Chapter 4 discusses millennia-scale ice preservation and kettle development following deglaciation in the Canadian Cordilleran Interior. Finally, Chapter 5 summarizes the conclusions of this research project and presents future research ideas.
References


Dawson, G.M. 1891. On the later physiographical geology of the Rocky Mountain region in Canada, with special reference to changes in elevation and to the history of the glacial period. Royal Society of Canada, Transactions (1890), 8: 3-74.


Paleogeography
Chapter 2
Late glacial lakes in the Thompson Basin, British Columbia: paleogeography, evolution and glacio-isostatic significance

2.1. Introduction

In North America, research on Late Wisconsinan deglacial lakes has emphasised lake development and environments for the generally low relief setting of the Laurentide Ice Sheet. In contrast, few detailed studies have been completed for the Late Wisconsinan (Fraser Glaciation) deglacial lakes of the moderately high relief setting of the Cordilleran Ice Sheet (CIS) (e.g. Fulton 1967, 1969, Shaw 1977, Shaw and Archer 1979, Eyles et al. 1987, Sawicki and Smith 1992, Huntley and Broster 1994, Ward and Rutter, 2000). This inequality has somewhat biased our general perception of deglacial lakes in Canada.

Within the southern Interior Plateau of British Columbia, many deglacial lakes developed during the decay of the Cordilleran Ice Sheet (CIS) ~10,000 to 12,000 $^{14}$C yr BP (Fulton 1969, Ryder et al. 1991, Clague and James 2002). These lakes were ribbon-shaped and developed within large valleys that dissect the plateau. They acted as major sinks for sediments derived from plateau areas. A rich geomorphic and sedimentary record of these lakes remains today for study. Their landforms and sediments provide a detailed record of ribbon lake environments associated with the decay of the CIS. This paper applies recent technologies and recent advances in glacial geomorphology and sedimentology to explore (1) the paleogeography and evolution of deglacial lakes in the Thompson Basin, and (2) regional glacio-isostatic effects and causes. Today, the availability of digital elevation models (DEMs), differential global positioning systems (DGPS), ground penetrating radar (GPR) and geographic information systems (GIS) make more detailed explorations of glacial lakes possible. This study is the first integrated application of these technologies for glacial lake reconstruction in the Cordillera.
2.2. Study area

The study area is situated within the southern Canadian Cordillera, in the southern Interior Plateau of British Columbia. It occupies a majority of the Thompson Basin and includes the Thompson, South Thompson, and the lower North Thompson river valleys (Fig. 2.1). It is bordered by the communities of Chase to the east, Lytton to the southwest, and McClure in the north - altogether this area includes 220 km of valleys. Within this area research was concentrated along the Spences Bridge-Ashcroft-Kamloops corridor (Fig. 2.1). Not included in this study, and located within the Thompson Basin, are the Nicola River valley and the Merritt Basin (Fig. 2.1b), previously studied by Anderton (1970) and Fulton and Walcott (1975), respectively.

The study area lies within the Thompson Plateau physiographic region and within the rain shadow of the Coast Mountains to the west. It neighbours the Fraser Plateau to the northwest, and the Shuswap Highlands to the east (Holland 1964). The Thompson Plateau is underlain by volcanic, plutonic, sedimentary and metamorphic rocks (Cockfield 1948, Duffel and McTaggart 1952). Fluvial and glacial erosion have dissected the Thompson Plateau resulting in deep valleys (1400 to 1600 m relief), which are today filled with thick (100 to >150 m) Quaternary deposits.

At the city of Kamloops, the North and South Thompson rivers meet to form the Thompson River that in turn flows west into Kamloops Lake. From the outlet of Kamloops Lake at Savona, the Thompson River continues to flow west and is met by the Deadman River near the outlet of Kamloops Lake, and the Bonaparte River near Ashcroft. The Thompson River turns ninety degrees near Ashcroft and narrows as it to flows south and past Spences Bridge and the Nicola River. South of Spences Bridge the Thompson Valley narrows further and deepens to eventually join the Fraser River at Lytton (Appendix 6).

High (>100 m) cliffs of glaciolacustrine silt dominate the South Thompson and Thompson valleys. West of Kamloops Lake, numerous fluvial terraces dominate the valley-sides up to 150 m above present river level. Trees are sparse in the arid valley-bottom allowing for relatively easy identification of glaciolacustrine landforms.

2.3. Previous Research

2.3.1. Deglacial style

The development of deglacial lakes in the Cordillera was closely tied to the style of deglaciation for the CIS. During the last glacial maximum, the CIS covered most of British Columbia, southern Yukon Territory and southern Alaska (Clague and James 2002). Decay of the CIS in the Interior of British Columbia was by regional stagnation of
Figure 2.1: (a) The study area, Spences Bridge-Ashcroft-Kamloops corridor, showing locations of water plane indicators (Table 2.2) and Figures. (b) The regional context of the study area showing locations of water plane indicator 18 (Table 2.1) and dated material (Table 1.1). (c) Location of study area in British Columbia, Canada. Li = Lillooet, Ly = Lytton, SB = Spences Bridge, M = Merritt, Ash = Ashcroft, CC = Cache Creek, K = Kamloops, Mc = McClure, Ch = Chase, SA = Salmon Arm, Ar = Armstrong, and V = Vernon. Th.R = Thompson River, N.Th.R = North Thompson River, and S.Th.R = South Thompson River.
the CIS rather than by active retreat (Fulton 1967, 1991). This inference is based on (1) a lack of recessional moraines recording CIS retreat from the valleys to the mountains, (2) reconstruction of the horizontal and altitudinal pattern of ice retreat in four small study areas from meltwater channels and ice-marginal landforms (Fulton 1967), and (3) conceptual modeling (Fulton 1991). Areas of thinner ice (plateaus) became ice-free sooner than areas of thicker ice (valleys; Fulton 1967, 1991). This assumes the valleys were not water-filled. Thus, two patterns of recession of the CIS margin have been inferred: (1) regionally, the ice margin retreated northwest toward the ice divide on the Fraser Plateau (Ryder et al. 1991); and (2) more locally, the plateau areas became ice-free before the valleys. Remnant ice masses within the valleys dammed meltwater and created many deglacial lakes (Fig. 2.2). In addition to being dammed by ice, the drainage pathways of these lakes were often in contact with ice (Fig. 2.2; Fulton 1969, 1991). Lake basins lengthened coeval with recession of the margins of these valley ice masses. The patchy distribution of remnant ice masses suggested that break-up of the CIS was by downwasting rather than by active frontal retreat, further supporting a deglacial model of regional stagnation (Fulton 1991).

2.3.2. Lake evolution

Lake evolution is defined here as encompassing changes in lake level, volume and extent, dam type and position, and drainage style. It is inferred from: (1) the elevations of wave-cut benches, deltas, ice-marginal terraces and drainage bedforms, (2) the distribution of lake bottom sediments, (3) topography, (4) the inferred position of ablating and stagnant ice-tongues, (5) glacio-isostatic rebound, and (6) regional base level changes (Fulton 1967, 1969).

A lake may have a range of possible water levels. Water levels that can be defined by correlating water plane indicators are named lake stages. The relative elevation of well-defined water planes (i.e. stages) within these ranges directs the naming of stages. No temporal duration to these stages is implied in this paper, due to the lack of chronological control in the study area. My definition is hydrological and does not necessarily imply a persistent stillstand. This is preferred, as water plane indicators may have formed under favourable energy and sediment supply conditions, not necessarily over a relatively long period of time. A new lake name is used when (1) the water bodies are in separate locations, or (2) the lake outlet position changed significantly.

Previous research presents a generalized picture of lake evolution for the southern interior of British Columbia. Four stages of two eastward draining late glacial lakes,
Figure 2.2: Evolution of late glacial lakes in the southern interior of British Columbia as proposed by Fulton (1969) and Ryder (1981). Regional stagnation of the Cordilleran Ice Sheet led to the development of glacial lakes impounded by valley ice masses. Lake basins lengthened coeval with recession of the fronts of these ice masses. Two glacial lakes and four glacial lake stages are identified in the study area (a-d). Present drainage is shown in (e). Not shown are earlier stages of Glacial Lake Merritt and Glacial Lake Penticton, and early Holocene Kamloops Lake - Cherry Creek stage (Fulton 1969).
Glacial Lake Thompson and Glacial Lake Deadman, are defined in the study area (Fig. 2.2; Fulton 1969, Ryder 1976, 1981).

The lakes and stages are characterized by progressive westward lake extension and relocation, and lake level lowering. The South Thompson Valley between Chase and Kamloops became ice-free first and was occupied by Glacial Lake Thompson – South Thompson stage dammed at either end (Fig. 2.2a). This lake lengthened and lowered forming Glacial Lake Thompson – Niskonlith stage (Fig. 2.2b) as ice tongues at its western and eastern ends and in the lower North Thompson Valley receded. With continued westward ice recession and further lake level lowering, high elevation lake bottom sediments in the South Thompson Valley became exposed, separating lakes in the Shuswap Basin from those in the Thompson Basin by an eastward draining river. This new outlet, east of Kamloops, defines Glacial Lake Deadman – Tranquille stage (Fig. 2.2c) and with further lake lowering and southward growth, Glacial Lake Deadman – Durand stage (Fig. 2.2d). Following lake drainage, the Thompson River reversed course and flowed south (Fig. 2.2e). In the early Holocene, a high stage of Kamloops Lake called Cherry Creek stage (not shown in Fig. 2.2) developed behind raised delta sediments (7, Fig. 2.1).

Glacial lake research in the eastern portion of the study area has focussed on the glaciolacustrine silts of the South Thompson Valley. Sediment descriptions (lithofacies, architecture), spatial sediment trends (e.g. varve analysis), kettled topography, and mapping of ice-marginal landforms have been combined to develop a picture of lake evolution in the South Thompson Valley (Fulton 1965, 1967, 1969). Landforms recording paleo-water plane(s) within the basin (i.e. deltas, wave-cut benches, etc.) have not been reported. Instead a kame terrace at 530 m asl and the highest occurrence of silt within the basin at 500 m asl bracket the South Thompson stage of Glacial Lake Thompson (Fulton 1965).

Glacial lake research within the western portion of the study area is limited (Fulton 1969; Anderton 1970; Ryder 1970, 1976, 1981). South of Ashcroft, Ryder (1970) described one exposure of glaciolacustrine sediments. In the lower Thompson and Nicola valleys, Anderton (1970) examined stratigraphic relationships and sediments to infer ice-tongue positions, lake paleogeography and damming mechanisms. Building on this work and extrapolating lake stages from Fulton (1969), Ryder (1976, 1981) developed a picture of orderly retreat of valley ice masses and concomitant extension and lowering of Glacial Lake Deadman punctuated by the failure of an ice dam south of Spences Bridge. This ice dam was inferred from the striking absence of lacustrine sediments between Spences Bridge and Lytton (Anderton 1970; Ryder 1970, 1976). However, as lake bottom
sediments exist at similar elevations in both the Thompson and neighbouring Fraser valleys (Fig. 2.1b), Ryder (1981) speculated that there may not have been a dam near Spences Bridge but instead the lake continued into the Fraser system. This inference is weak as lake bottom sediments do not indicate a former synchronous lake level, and the absence of lake sediments along 30 km of valley south of Spences Bridge remains unexplained.

2.3.3. Stratigraphy and timing of events

The relative timing of past lakes in the study area has been determined by stratigraphic reconstruction (Fulton and Armstrong 1965, Ryder 1976, 1981, Fulton and Smith 1978, Clague 2000; Fig. 2.3). West of Savona, pre-Fraser Glaciation (pre-oxygen isotope stage 2) sediments are exposed in a 15 km stretch of the Thompson Valley from ~3 km north of Ashcroft to ~12 km south of Ashcroft (Clague and Evans 2003). These sediments include clay-rich lacustrine deposits that form the failure plane for numerous landslides. Late glacial (Fraser Glaciation) lake sediments overlie Fraser till (Kamloops Lake drift, Fulton and Smith 1978) and are commonly overlain by Holocene terrace gravel and loess. These late glacial lacustrine sediments are the dominant Quaternary deposit in the valley. Cliff exposures and well logs indicate that late glacial lake sediments exceed 150 m in thickness.

East of Savona, pre-Fraser sediments have been mapped in a few locations (Fulton 1975). Exposure descriptions have included tephras and bones of several extinct mammals of pre-Fraser Glaciation age (Fulton and Smith 1978, Fulton 2000). The most dominant deposit within the eastern portion of the study area is Fraser Glaciation late glacial lacustrine sediment. In the South Thompson Valley, cliffs of glaciolacustrine silt tower over 100 m above the present river level and dominate the landscape. Cliff exposures and well logs indicate that late glacial lake sediments exceed 140 m in thickness (Fulton 1965).

Absolute timing of late glacial lake initiation and duration are uncertain. Only minimum ages are available for deglaciation and late glacial lakes in the region as datable organics from this period are scarce (Table 2.1). Deglaciation of plateau areas was likely well advanced by 10,000 ¹⁴C yr BP (Fulton 1971, Clague 1981). Radiocarbon dates are not available for specific lake stages, as organics have not been found in deltaic or glaciolacustrine sediments during this or any previous investigations (Fulton 1969). All late glacial lakes drained no later than 8,900 ±150 ¹⁴C yr BP (2, Fig. 2.1b, Table 2.1; Dyck et al. 1965; Fulton 1969). An age determined from thermoluminescence dating of
Figure 2.3: The stratigraphic context of lacustrine sediments (unit 6) examined in this study (Bonaparte Section near, see Fig. 2.1 for location). Stratigraphic position implies that these lacustrine sediments are late glacial in origin (Clague 2000, Clague and Evans 2003) following the Fraser Glaciation (oxygen isotope stage 2). This section records an inset valley-fill stratigraphy. Elsewhere, unit 6 sediments typically occur in units much thicker (>80 m) than at this section, and form the dominant valley fill.
### Table 2.1: Dates relevant to deglaciation and late glacial lakes of the region

<table>
<thead>
<tr>
<th>Site</th>
<th>Lab no.</th>
<th>Age ($^{14}C$ yr BP)</th>
<th>Sample description; context; and inference</th>
<th>Lat.(N)</th>
<th>Long.(W)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Deglaciation lakes (minimum ages)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>S-1737</td>
<td>8,250 ±115</td>
<td>Human skeleton (collagen); overlying glaciolacustrine sediments in South Thompson Valley; minimum age for human occupation</td>
<td>50° 41.8'</td>
<td>119° 49.5'</td>
<td>Cybulski et al. 1981; Chrisholm and Nelson 1983; archaeological site Eeqw48;</td>
</tr>
<tr>
<td></td>
<td>GSC-1857</td>
<td>8410 ±100</td>
<td>Wood; core from Kamloops Bog, Okanagan Valley; minimum age for all glacial lakes in Okanagan Valley</td>
<td>49° 56' 119° 23'</td>
<td>Alley 1976</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>GSC-193</td>
<td>8,900 ±150</td>
<td>Organic-rich silt; spillway for Glacial Lake Shuswap in north Okanagan Valley; minimum age for all Fraser Glaciation late glacial lakes</td>
<td>50° 24' 119° 15'</td>
<td>Dyck et al. 1965, Fulton 1969</td>
<td></td>
</tr>
<tr>
<td></td>
<td>S-113</td>
<td>9,000 ±150</td>
<td>Charcoal; river terrace of the Fraser River; minimum age for human occupation, deglaciation and possible late glacial lakes for the lower Fraser Canyon</td>
<td>49° 33' 121° 24'</td>
<td>Borden 1965, 1968; archaeological site DjiRi-3</td>
<td></td>
</tr>
<tr>
<td></td>
<td>GSC-256</td>
<td>9,320 ±160</td>
<td>Mucky peat; from base of bog in spillway from glacial lake in Merritt basin; minimum age for a high lake in the Merritt basin, Glacial Lake Quilchena</td>
<td>49° 53.0' 120° 37.5'</td>
<td>Dyck et al. 1965, Fulton 1969</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>GSC-526</td>
<td>9,750 ±170</td>
<td>Fibrous organics mixed with clay; spillway from small unnamed glacial lake on plateau; minimum age for lake and deglaciation of plateau</td>
<td>50° 32.5' 119° 45.2'</td>
<td>Lowdon et al. 1967</td>
<td></td>
</tr>
<tr>
<td>Deglaciation (minimum ages)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>GSC-1524-2</td>
<td>10,500 ±370$^2$</td>
<td>Plant organic fraction of marl-peat; core near Harper Lake; minimum age for deglaciation of plateau</td>
<td>50° 44.1'</td>
<td>119° 43.5'</td>
<td>Lowdon and Blake 1973</td>
</tr>
<tr>
<td></td>
<td>I-6058</td>
<td>11,140 ±260$^2$</td>
<td>Clay-gyttja; core of lake situated 175 m above present river level, located in lower Fraser Canyon 80 km south of Lytton; minimum age for deglaciation but not for possible late glacial lakes or sea incursion</td>
<td>49° 29.0' 121° 24.3'</td>
<td>Mathewes et al. 1972</td>
<td></td>
</tr>
<tr>
<td></td>
<td>I-6057</td>
<td>11,430 ±150$^2$</td>
<td>same as above (I-6058)</td>
<td>49° 29.5' 121° 26.0'</td>
<td>Mathewes et al. 1972</td>
<td></td>
</tr>
<tr>
<td>Deglaciation lakes (ambiguous ages)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>SFU-TL-578</td>
<td>14,200 ±2,300$^3$</td>
<td>Clay-rich layer in glaciolacustrine sediments in the South Thompson Valley; as this date is for a single sample and has a large error, it should be treated with caution; this date could correspond to Fulton's (1969, fig. 3), Glacial Lake Thompson - South Thompson stage, or Niskonlith stage, or to an early unnamed stage of Glacial Lake Shuswap</td>
<td>50° 41.8'</td>
<td>119° 49.5'</td>
<td>Berger 1986</td>
</tr>
</tbody>
</table>

$^1$ Site numbers (1-5) refer to sites of dated material located in Fig. 2.1b  
$^2$ These ages likely suffer old carbon effects  
$^3$ Thermoluminescence date
glaciolacustrine silt is enigmatic (Table 2.1). Perhaps future optical dating may improve lake chronology (e.g. Lian and Hicock 2001).

Varve studies completed in the glaciolacustrine sediments of the South Thompson Valley suggest that lakes within this valley were short-lived (Fulton 1965). Up to eighty thinning-upward (6 - 0.03 m thick) varves are described in a single exposure, suggesting that it took ~80 years to deposit tens of metres of sediment (Fulton 2000). This rate of sedimentation is much higher than those calculated for areas of lower relief and for most modern glacial lakes (Smith and Ashley 1985).

Glacial lakes in the Thompson Valley formed after a high glacial lake in the Merrit Basin disappeared (Glacial Lake Quilchena, Fulton 1969). Since the minimum age for a spillway for Glacial Lake Quilchena is 9,750 ±170 14C yr BP and all late glacial lakes drained prior to 8,900 ±150 14C yr BP (Table 2.1), that leaves only ~850 14C years for the lifespan of all glacial lakes in the Thompson Valley. However this is a rough estimate as these are minimum ages.

2.3.4. Glacio-isostasy in the Cordilleran Interior

Marine shoreline indicators provide a rich source of data for the reconstruction of glacio-isostatic rebound along the coastal margin of the Cordilleran Ice Sheet (e.g. James et al. 2002). In contrast, our understanding of glacio-isostatic rebound in the Cordilleran Interior is limited to paleo-lake shoreline indicators. The only robust study of such features in the Cordilleran Interior was completed around Merritt (Fig. 2.1b; Fulton and Walcott 1975). The elevations of water plane indicators (i.e. wave-cut benches and deltas) were measured for glacial lakes in the Merritt Basin using an altimeter tied to benchmarks. Geomorphic pattern (paired wave-cut benches) rather than surface dating was employed for correlation. Linear regression gave a bearing of 350° (upslope direction) and reliable tilts of 1.6 - 1.9 m km⁻¹ (over <35 km baseline). A northward pattern of ice retreat was inferred from these isobases. Estimates of glacio-isostatic rebound in the Thompson Basin, using water plane indicators and a simple three-point geometric method for Glacial Lake Thompson – Niskonlith stage (R.J. Fulton personal communication 2001), suggest glacio-isostatic deformation follows an upslope bearing of 300° and dip of 1.2 m km⁻¹ (Fulton 1969, 2000).

A number of gaps exist in our understanding of glacial lakes in the Thompson Basin. In this chapter, I (1) investigate water plane indicators, (2) reconstruct lake paleogeography, (3) infer lake evolution, and (4) discuss implications for glacio-isostatic rebound. Recent technologies (DGPS, GPR), digital data and analysis (DEM’s, GIS) and geomorphologic and sedimentologic analyses are applied.
2.4. Approach and Methods

Reconstruction of the paleogeographic evolution of glacial lakes and glacio-isostasy in the Cordilleran Interior requires analysis of water plane indicators (Table 2.2). My analysis involves: (1) inventory, evaluation, elevation determination and correlation of water plane indicators, (2) inference of lake extents and dams, and (3) integration of lake extents and DEMs of water planes, topography, and inferred lake bottom to determine lake paleogeography and geometry (Fig. 2.4). Points 2 and 3 are addressed in later sections (sections 2.6.2, 2.6.3).

2.4.1. Inventory of water plane indicators

Water plane indicators are landforms and/or sediment that (1) approximate a paleo-water plane (primary indicators), or (2) over-estimate or under-estimate a paleo-water plane (secondary indicators; Table 2.2).

Potential water plane indicators were identified using aerial photographs, a digital elevation model, an evaluation of previous research (Fulton 1965, 1967, 1969, Anderton 1970, Ryder 1970, 1976, 1981) and field identification. Black and white, and colour aerial photographs of multiple scales (1:20,000 to 1:5,000) and dates and exhibiting a range of sun angles (Appendix 1) were used to identify subtle wave-cut benches and other water plane indicators (Table 2.2). Experience demonstrated that wave-cut benches were not visible in some aerial photograph series, but were visible in others. Smaller area aerial photographs (e.g. 1:15,000 and 1:20,000) proved best for wave-cut bench identification. Hillsides were observed over a long field season and under a variety of lighting conditions to recognize wave-cut benches. Primary water plane indicators mapped by previous workers were also identified in the field.

2.4.2. Evaluation of water plane indicators

Field data were gathered to confirm the identification of potential water plane indicators (Tables 2.2 and 2.3). Where exposures were available the sedimentology of water plane indicators was recorded using standard field observations (texture, structure, bed contact relationships, photographs, etc.).

GPR was employed to determine the sub-surface sedimentary architecture of landforms where exposures were absent. These data assisted in distinguishing deltas from fans and river terraces. A Sensors and Software pulseEKKO™ IV GPR system was employed with a 400 V transmitter and 50 MHz antennae. Antennae separation and length were 2.0 m, trace sampling was 0.5 m, sampling rate was 1600 ps, and traces
Figure 2.4: Flowchart of paleo-lake reconstruction methodology. See Table 2.2 for definitions of primary and secondary water plane indicators.
### Table 2.2: Paleo-water plane indicators

<table>
<thead>
<tr>
<th>Indicator</th>
<th>Appearance</th>
<th>Application</th>
<th>Explanation and limitations</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Primary indicators:</strong> approximation of paleo-water plane</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Wave-cut benches</td>
<td>Distinct horizontal benches along hillsides (100 - 2000 m long), consistent bench and riser slope morphology (Table 2.3), cross-cutting relationships with gullies (Fig. 2.5)</td>
<td>Best approximation of water plane</td>
<td>Elevation records wave base; Elevation may over-estimate average wave base if bench forms during high wind events (e.g. storms); Formation requires sufficient fetch, lake level stability and/or time. The lowest slope-break of a bench was consistently used to estimate the paleo-water plane as it (1) was the most consistently distinguishable morphological element, (2) agrees with modern environments (e.g. Allan et al. 2002), and (3) is an appropriate method for regional correlations (Teller, personal communication 2001). The topset-foreset contact is used for approximation; water plane may lie 1 to 4 m above this contact (Gustavson et al. 1975, Thorson 1989); elevation may be seasonally biased (lake levels may have been highly variable as even under the current regime the unmanaged level of Kamloops Lake varies up to 7 m (Pharo and Carmack 1979); upper surface is longitudinally sloped and can produce significant error; distal truncation of delta leads to over-estimation of water plane. The topset thickness was subtracted from the elevation of the delta surface to derive an estimate of lake level. The topset thickness was determined from either sedimentary exposures, GPR profiles, or estimated from other delta surfaces in the study.</td>
</tr>
<tr>
<td>Delta</td>
<td>Large (2 - 13 km²) bodies of gravel located where tributaries meet main valleys; generally fan-shaped and gently sloping from tributary mouth; often composed of multiple inset surfaces</td>
<td>Approximation of water plane</td>
<td></td>
</tr>
<tr>
<td><strong>Secondary indicators:</strong> over-estimation or under-estimation of paleo-water plane</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Drainage bedforms</td>
<td>Large bedforms (~100 m wavelength, ~6 m amplitude), steeper lee-side slopes; GPR shows erosional origin; found on Deadman delta (within the lake basin; Fig. 2.14)</td>
<td>Under-estimation of water plane</td>
<td>Produced during final drainage of the last lake to occupy the valley; location influenced by confinement in flow (i.e. distal portion of sloped delta surface); associated with erosional surfaces on Deadman delta.</td>
</tr>
<tr>
<td>Lake bottom sediments</td>
<td>Large (50-100 m tall) cliffs of pale-yellow, fine-grained, water-lain sediment; often capped by imbricate gravel and mapped as river terraces (Ryder 1976, 1981)</td>
<td>Under-estimation of water plane, approximation of lake bottom</td>
<td>Elevation of lake bottom sediments is an under-estimation of the water level of the lake in which the sediments were deposited; at their highest elevation there is no capping gravel and thus they record lacustrine benches that approximate the elevation of the former lake bottom. In this study, all large, continuous bodies of lake bottom sediment were below the lowest paleo-water plane.</td>
</tr>
<tr>
<td>Subaqueous fans</td>
<td>Large (10's m tall), wedge-shaped, body of medium to coarse-grained, water-lain sediment; located where tributaries meet main valleys</td>
<td>Under-estimation of water plane</td>
<td>Subaqueous fans are, by definition, deposited into a lake below the water surface (Rust 1977, Cheel and Rust 1980, Bell et al. 2001).</td>
</tr>
<tr>
<td>Paleo-channels</td>
<td>Perched, dry, incised, river channel, gravel lined</td>
<td>Over-estimation or under-estimation of water plane</td>
<td>Represents either former river flow (1) into a lake (over-estimation), or (2) post-lake fluvial incision (over- or under-estimation - see River terraces below).</td>
</tr>
<tr>
<td>River terraces</td>
<td>Broad, near-flat, imbricate gravel surfaces frequently capping cliffs of glaciolacustrine sediment</td>
<td>Under-estimation of water plane, or over-estimation of water plane</td>
<td>After the lake drains, a fluvial system develops. The highest fluvial terraces remaining after incision record an under-estimate of a paleo-lake level. However, if post-lake aggradation exceeded the paleo-lake level, this principle cannot be applied. In this study, all terraces were below the lowest paleo-water plane. Related to this indicator are ice-marginal terraces that form from the deposits of lake-terminating ice marginal streams that run along valley-occupying ice tongues. Ice-marginal terraces over-estimate a paleo-lake water plane.</td>
</tr>
</tbody>
</table>

*Indicators are listed in decreasing order of certainty for estimation of a paleo-water plane.*
### Table 2.3: Inventory of water plane indicators

#### Primary water plane indicators

<table>
<thead>
<tr>
<th>Site no.</th>
<th>Indicator type</th>
<th>Northing</th>
<th>Easting</th>
<th>Elev (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Wave-cut bench</td>
<td>5593908</td>
<td>621706</td>
<td>443.8</td>
</tr>
<tr>
<td>2</td>
<td>Wave-cut bench</td>
<td>5601171</td>
<td>620417</td>
<td>455.7</td>
</tr>
<tr>
<td>3a</td>
<td>Wave-cut bench</td>
<td>5606109</td>
<td>620947</td>
<td>500.1</td>
</tr>
<tr>
<td>3b</td>
<td>Wave-cut bench</td>
<td>5606108</td>
<td>620931</td>
<td>553.6</td>
</tr>
<tr>
<td>4</td>
<td>Wave-cut bench</td>
<td>5625913</td>
<td>632047</td>
<td>570.4</td>
</tr>
<tr>
<td>5a</td>
<td>Wave-cut bench</td>
<td>5626235</td>
<td>632474</td>
<td>469.3</td>
</tr>
<tr>
<td>5b</td>
<td>Wave-cut bench</td>
<td>5626256</td>
<td>632487</td>
<td>480.6</td>
</tr>
<tr>
<td>5c</td>
<td>Wave-cut bench</td>
<td>5626286</td>
<td>632505</td>
<td>470.6</td>
</tr>
<tr>
<td>6a</td>
<td>Wave-cut bench</td>
<td>5622889</td>
<td>649245</td>
<td>526.8</td>
</tr>
<tr>
<td>6b</td>
<td>Wave-cut bench</td>
<td>5622912</td>
<td>649239</td>
<td>516.7</td>
</tr>
<tr>
<td>7a</td>
<td>Delta level</td>
<td>5626039</td>
<td>647867</td>
<td>540.6</td>
</tr>
<tr>
<td>7b</td>
<td>Delta level</td>
<td>5626153</td>
<td>648181</td>
<td>529.0</td>
</tr>
<tr>
<td>7c</td>
<td>Delta level</td>
<td>5625911</td>
<td>645923</td>
<td>479.5</td>
</tr>
<tr>
<td>7d</td>
<td>Delta level</td>
<td>5625966</td>
<td>646060</td>
<td>462.3</td>
</tr>
<tr>
<td>7e</td>
<td>Delta level</td>
<td>5625250</td>
<td>646752</td>
<td>436.7</td>
</tr>
<tr>
<td>8a</td>
<td>Delta level</td>
<td>5624894</td>
<td>654519</td>
<td>421.8</td>
</tr>
<tr>
<td>8b</td>
<td>Delta level</td>
<td>5625215</td>
<td>654626</td>
<td>403.4</td>
</tr>
<tr>
<td>9</td>
<td>Delta level</td>
<td>5623676</td>
<td>655715</td>
<td>527.5</td>
</tr>
<tr>
<td>10</td>
<td>Wave-cut bench</td>
<td>5623261</td>
<td>660816</td>
<td>535.6</td>
</tr>
<tr>
<td>11</td>
<td>Wave-cut bench</td>
<td>5621208</td>
<td>668516</td>
<td>523.7</td>
</tr>
<tr>
<td>12</td>
<td>Delta level</td>
<td>5618433</td>
<td>670162</td>
<td>516.4</td>
</tr>
<tr>
<td>13</td>
<td>Wave-cut bench</td>
<td>5619545</td>
<td>673031</td>
<td>520.1</td>
</tr>
<tr>
<td>14a</td>
<td>Delta level</td>
<td>5622922</td>
<td>675325</td>
<td>422.2</td>
</tr>
<tr>
<td>14b</td>
<td>Delta level</td>
<td>5622548</td>
<td>674572</td>
<td>389.1</td>
</tr>
<tr>
<td>15</td>
<td>Wave-cut bench</td>
<td>5620229</td>
<td>689320</td>
<td>508.7</td>
</tr>
<tr>
<td>16</td>
<td>Delta level</td>
<td>5635879</td>
<td>686630</td>
<td>421.5</td>
</tr>
<tr>
<td>17</td>
<td>Wave-cut bench</td>
<td>5633576</td>
<td>692278</td>
<td>531.8</td>
</tr>
<tr>
<td>18</td>
<td>Wave-cut bench</td>
<td>5628983</td>
<td>728374</td>
<td>492.1</td>
</tr>
</tbody>
</table>

#### Secondary water plane indicators

<table>
<thead>
<tr>
<th>Indicator type</th>
<th>Approx elev (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Drainage bedform</td>
<td>5624082 647943  427</td>
</tr>
<tr>
<td>Fluvial terrace</td>
<td>5615004 618315  455</td>
</tr>
<tr>
<td>Fluvial terrace</td>
<td>5623654 622054  445</td>
</tr>
<tr>
<td>Fluvial terrace</td>
<td>5626595 632600  416</td>
</tr>
<tr>
<td>Fluvial terrace</td>
<td>5627430 687605  360</td>
</tr>
<tr>
<td>Fluvial terrace</td>
<td>5623547 623464  400</td>
</tr>
<tr>
<td>Paleo-channel</td>
<td>5625505 619397  478</td>
</tr>
<tr>
<td>Paleo-channel</td>
<td>5620561 619385  460</td>
</tr>
<tr>
<td>Paleo-channel</td>
<td>5609675 617330  450</td>
</tr>
<tr>
<td>Lake bottom sediments</td>
<td>5625800 640370  400</td>
</tr>
<tr>
<td>Lake bottom sediments</td>
<td>5617747 621432  415</td>
</tr>
<tr>
<td>Lake bottom sediments</td>
<td>5626595 632600  400</td>
</tr>
<tr>
<td>Lake bottom sediments</td>
<td>5625703 629300  410</td>
</tr>
<tr>
<td>Lake bottom sediments</td>
<td>5594100 621642  400</td>
</tr>
<tr>
<td>Lake bottom sediments</td>
<td>5587320 620380  400</td>
</tr>
<tr>
<td>Lake bottom sediments</td>
<td>5618568 694733  460</td>
</tr>
<tr>
<td>Lake bottom sediments</td>
<td>5623572 624758  430</td>
</tr>
<tr>
<td>Lake bottom sediments</td>
<td>5592244 621427  420</td>
</tr>
<tr>
<td>Subaqueous fan</td>
<td>5591220 620230  430</td>
</tr>
</tbody>
</table>

---

1 Refer to locations in Figure 2.1 and 2.10
2 Refer to Table 2.2.
3 Coordinates are NAD83, UTM Zone 10. Elevation is above sea level, CVD28 datum. The elevation error is estimated conservatively at ±20 cm at ±2σ. See text for explanation.
were stacked either 64 or 128 times. Velocity sounding or common midpoint gathers were conducted along GPR profiles. Velocities of 0.10 and 0.13 m ns\(^{-1}\) were measured and generally agree with published data for dry sand and gravel (Jol and Smith 1992, Sensors and Software 1996). GPR data were processed to improve the quality of radar profiles for visualization and interpretation. Processing included: (1) time-zero adjustment, (2) topographic correction where applicable, (3) dewow, (4) 5-point down-trace average, and (5) automatic gain control with maximum cut-off of 500.

The genesis of water plane indicators was determined by retroductive inference (Baker 1996). Results of the inventory and verification of these indicators are presented in section 2.5.

2.4.3. Elevation of water plane indicators

A Leica\textsuperscript{TM} SR530, dual frequency, 24-channel, real-time kinematic (RTK), differential phase global positioning system (DGPS) was used to survey the location of primary water plane indicators (wave-cut benches and deltas, Table 2.2). Stations were occupied more than once to evaluate potential phase ambiguity errors (Higgins 1999); none were found. Survey networks were tied to first-order federal government geodetic control markers. The GPS data were converted to orthometric heights using a geoid model (HT97). The elevation of each water plane indicator was corrected by applying the difference between the GPS estimated orthometric height of a geodetic marker and the published marker height.

The measurement error associated with determining the elevation of water plane indicators by DGPS, as determined by the firmware algorithms (Leica Geosystems 2000a), was less than ±8 cm at ±2\(\sigma\). This error agreed with expected performance results for the survey equipment (Higgins 1999, Leica Geosystems 2000b). Combined with survey errors from control markers and the geoid model, the overall error is conservatively ±20 cm at ±2\(\sigma\). This measurement error is insignificant when compared to (1) complexities in landform morphology (e.g. slopes, irregularities), (2) uncertainties in the application of landforms to infer paleo-water planes (Table 2.2), and (3) the regional scale of the study.

2.4.4. Correlation of water plane indicators

Paleo-water planes are discovered by correlating primary water plane indicators using (1) iterative visual 3-D graphical assessment of water plane indicator data, and (2) trend surface analysis. Chronologic correlation and correlation on the basis of geomorphic pattern (e.g. paired cut-benches) were not possible.
The water planes were modelled using regression analysis (Burrough 1986). This approach was also employed by Fulton and Walcott (1975) and is an appropriate trend surfacing technique (Gray 1983) as glacio-isostatically deformed water planes are typically planar or slightly curvi-planar depending on the scale of study (e.g. Fulton and Walcott 1975, Teller and Thorleifson 1983, Parent and Occhietti 1999). In this study, planar (i.e. first-order) modelling of tilted water planes proved adequate (section 2.6.1). The regression procedure provides a planar surface described by:

\[ Z = a + bX + cY \]

where \( Z, X \) and \( Y \) are the elevation (CVD28 datum), easting and northing respectively in UTM, Zone 10 coordinates (NAD83 datum), and \( a, b \) and \( c \) are parameters determined by the least-squares procedure using SYSTAT™ v10 statistical software.

The upslope direction of the plane of best fit (\( W \)) in degrees azimuth from grid north is derived using:

\[ W = \tan^{-1}(b/c) \]

and converted to true north by correcting for grid north (adding 1.75°). The slope of the plane (\( S \)) in m m\(^{-1}\) is derived using:

\[ S = \tan^{-1}(b/\sin(W)) \]

The error associated with \( W \) and \( S \) was also calculated in SYSTAT™ v10.

### 2.5. Interpretation of water plane indicators

The correct identification of water plane indicators is pivotal to the reconstruction of lake paleogeography; incorrect interpretations of individual landforms can corrupt paleogeographic reconstructions. This section presents the results of a thorough inventory, description and verification of water plane indicators. Their application to the problem of reconstructing past lake stages is discussed.

#### 2.5.1. Wave-cut benches

Wave-cut benches develop in areas of adequate fetch by the action of waves eroding and transporting sediments in the nearshore (Fulton and Walcott 1975, Allan et
al. 2002). They are identified on the basis of their appearance (Table 2.2, Fig. 2.5), within-group consistent geometry (Table 2.4) and regional correlation (section 2.6.1). In the study area, sixteen wave-cut benches were used in regional water plane reconstructions (Fig. 2.1, Table 2.3). All are located well above the elevation of the highest fluvial terraces and appear to be cut into till-covered slopes, similar to those near Merritt (Fulton and Walcott 1975). Wave-cut benches were discovered only at mid to low elevations along valley sides either because at higher elevations (1) vegetation masked the benches, (2) the benches did not develop because of physical controls (e.g. the water plane intersected bedrock), or (3) the lake water level did not exist at high elevations.

Table 2.4: Wave-cut bench morphology

<table>
<thead>
<tr>
<th>Site no.</th>
<th>Bench slope (°)</th>
<th>Riser slope (°)</th>
<th>~Width (m)</th>
<th>Shore aspect</th>
</tr>
</thead>
<tbody>
<tr>
<td>18</td>
<td>11</td>
<td>21</td>
<td>17</td>
<td>SE</td>
</tr>
<tr>
<td>6b</td>
<td>14</td>
<td>28</td>
<td>8</td>
<td>N</td>
</tr>
<tr>
<td>15</td>
<td>12</td>
<td>30</td>
<td>21</td>
<td>NW</td>
</tr>
<tr>
<td>13</td>
<td>14</td>
<td>28</td>
<td>7</td>
<td>N</td>
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<tr>
<td>10</td>
<td>12</td>
<td>18</td>
<td>15</td>
<td>E</td>
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<td>17</td>
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<td>7</td>
<td>N</td>
</tr>
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<td>2a</td>
<td>18</td>
<td>33</td>
<td>12</td>
<td>E</td>
</tr>
<tr>
<td>1b</td>
<td>15</td>
<td>30</td>
<td>17</td>
<td>N</td>
</tr>
</tbody>
</table>

Average 14 26 13
Standard deviation 2 5 5

1 Morphologies were not determined for some wave-cut benches: 3a-b, 4, 5a-c, and 6a (Table 2.3)
2 See Figure 2.1 and Table 2.3 for locations
3 Benches correspond with the Lower water plane, Glacial Lake Deadman - Lowest stage; remainder correspond to the Upper water plane, Glacial Lake Thompson - High stage

2.5.2. Deltas

Deltas developed where rivers delivered large quantities of sediment to the lake (Fig. 2.6). They have upper surfaces that gently slope into the main valleys and often contain several inset surfaces. Post-lake incision has truncated and deeply dissected these landforms. The location and elevation of the topset-foreset contact of twelve delta surfaces (Fig. 2.1, Tables 2.2 and 2.3) are used to reconstruct paleo-water planes in the study area. They are located between the McClure and the Deadman–Thompson River confluence (Fig. 2.1).
Figure 2.5: (a) Aerial view (Aerial photograph BCC132: 123, 1976, copyright British Columbia Government, by permission) and (b) Oblique view of wave-cut bench (solid arrows; 6, Fig. 2.1). The wave-cut bench extends beyond the photographs for a total length of ~2 km. The lateral extent of the wave-cut bench is interrupted by gullies. Paired portion not shown. Bedrock lineaments indicated by dashed arrows.
Figure 2.6: (a) Perspective (5x vertical exaggeration) and (b) plan view hillshade DEMs of the Deadman River delta showing GPR survey locations (G1-G4; Figs. 2.1, 2.7, 2.14, Appendix 2). Note multiple inset delta surfaces, (c, dashed lines 1,2,3a,3b), delta dissection, kettle holes and drainage bedforms. Delta surface 3b and the southern portion of 3a have been heavily modified by catastrophic lake drainage and the melting of buried ice blocks (Fig. 2.14). Lake stages are projected across this DEM. Present river flow direction, blue arrows. (DEM data, British Columbia Government 1996)
Delta identification is confirmed by geomorphology (Table 2.2), sedimentology and/or GPR surveys; one example is discussed. The largest (13 km²) delta in the region is the Deadman delta (7, Fig. 2.1, 2.6; Fulton, 1987). Three large delta surfaces (~540, 480 and 450 m asl topset-foreset contact, Fig. 2.6c) and several smaller delta surfaces are visible. Sediment exposures show inclined beds of sand and imbricate gravel (foresets) underlying horizontally stratified imbricate gravel (topsets; Chapter 3). GPR profiles record dipping reflections (foresets; Fig. 2.7, Appendix 2). Both GPR profiles and sediment exposures confirm the deltaic origin of this landform. As former late glacial lake levels formed the base level for delta progradation, lake levels are inferred to have lowered progressively. GPR information also indicates that the Deadman delta likely built out in lobes. The implications of the Deadman delta for lake evolution are discussed in section 2.7.1 and Chapter 3.

2.5.3. Secondary water plane indicators

Drainage bedforms can be produced during catastrophic lake drainage (Clague and Rampton 1982, Baker et al. 1987). In this study drainage bedforms were observed towards of the toe of Deadman delta (Fig 2.8). They are large wavelength (100 m) forms of erosional origin (Fig 2.8). They under-estimate the former lake level (Table 2.2).

Late glacial (Fraser Glaciation) lake bottom sediments (dominantly fine sand and silt) are often exposed in large cliffs (50-100 m tall) composed of pale-yellow, dominantly fine-grained, water-lain sediment (Table 2.2). They are usually capped by imbricate gravel (river terraces) and under-estimate the former lake-bottom. In some cases the highest occurrences of lake bottom sediments are not capped by river gravel and thus approximate the former lake bottom. In the western portion of the study area large cliffs of lake bottom sediments are found in a continuous swath (~75 km long) along the valley from the outlet of Kamloops Lake to a few kilometres south of Spences Bridge near Skoonka Creek (Fig. 2.1). The highest lake bottom sediments are ~400 m asl throughout this area. In the eastern portion of the study area, impressive glaciolacustrine cliffs towering >100 m above present river level (~460 m asl) are found along the full length of the South Thompson Valley (~55 km long), whereas in the adjacent North Thompson Valley they only extend ~10 m above the river (~360 m asl).

Subaqueous fans are point-source accumulations of sediment that develop below lake level (Rust 1977, Cheel and Rust 1980, Bell et al. 2001). Thus, their surface elevation underestimates a paleo-lake water level. I identify subaqueous fans on the basis of their geomorphology (Table 2.2) and lithofacies associations (e.g. Rust 1977, Cheel and Rust 1980, Bell et al. 2001, Chapter 3). Some of these landforms were
Figure 2.7: 350 m long GPR profile on the Deadman delta (7, Fig. 2.1). The survey was completed along a road cut into the terrace riser between the middle and lower terraces on the western half of the delta (G1, Fig. 2.6b). Dipping reflections (32° dip at 67 m on profile) are delta foresets. The parabola at 120 - 200 m was produced by interference from an overhead powerline.
Figure 2.8: Undulating terrain on Deadman delta interpreted as erosional bedforms produced during drainage of Glacial Lake Deadman - Lowest stage (see Fig. 2.6a for location). (a) View of bedforms looking northwest from wave-cut bench 6 (Fig. 2.1). Kettles, K; inset delta levels, L; G2, GPR location. Drainage direction to west (large arrow) (b) GPR profile (G2, Fig. 2.6b, Appendix 2.1) across an erosional bedform. Inclined radar reflections (dip is 28° at 100 m on profile) record delta foresets truncated at land surface.
previously interpreted as deltas (Ryder 1970). In the study area, they formed where rivers delivered large quantities of sediment to the lake and their deposits did not aggrade to the lake surface. They are a common deposit in the western portion of the study area (Chapter 3). For example, a pair of subaqueous fans, over 150 m thick, issue from Twaal and Pimainus creeks, and a well-exposed subaqueous fan, over 80 m thick, is located near the confluence of the Bonaparte and Thompson rivers (Fig. 2.1; Appendix 5.2a).

Paleo-channels are perched dry channels eroded by former rivers. The rivers either flowed into the late glacial lakes or were post-lake rivers. The former gives an over-estimation of the former lake level while the later gives an over- or under-estimation depending on river incision/aggradation history (Table 2.2). A few channels are near the highest fluvial terraces just west of Ashcroft. They are part of a larger channel complex, the paleo-Bonaparte River system, discussed in section 2.5.4.3.

River terraces are very common throughout the western portion of the study area and are up to 150 m above present river level. Using the elevation of these landforms as indicators of paleo-lake water planes can be difficult (Table 2.2). Related to river terraces are ice-marginal terraces (Table 2.2). This landform has been mapped near Kamloops (Fulton 1967). Similar features have not been found elsewhere in the study area.

The elevation of most secondary water plane indicators was estimated from (1) geodetic control markers on the landforms, (2) altimetry to nearby benchmarks, (3) DEMs, or (4) a combination of these techniques.

2.5.4. Reinterpreted water plane indicators

This investigation has resulted in a reinterpretation of several landforms previously studied in the Thompson Basin (Ryder 1976, 1981): Brassy Creek terrace (Fig. 2.9), Brassy Creek bench and the paleo-Bonaparte River system (Fig. 2.10).

2.5.4.1. Brassy Creek terrace

West of the Deadman–Thompson River confluence is a prominent terrace, referred to here as Brassy Creek terrace, that has been previously interpreted as a delta that prograded from Brassy Creek during the Glacial Lake Deadman – Tranquille stage (Figs. 2.1, 2.9; Ryder 1976). A re-examination of the landform based on aerial photographs, GPR surveys and field observations indicates that it is likely a fluvial terrace.
Figure 2.9: Brassy Creek river terrace (Fig. 2.1). (a) Geomorphic context and GPR survey locations. GPR profiles (b) G5 and (c) G6. Radar facies 1 and 2 are interpreted as river terrace gravel and flood eddy deposits, respectively. See text for discussion. (Aerial photograph BCR6038: 146, 1986, copyright British Columbia Government, by permission)
Figure 2.10: (a) Oblique and (b) plan view hillshade DEMs of the paleo-Bonaparte River system near Ashcroft (figure located in Fig. 2.1). A system of paleo-channels and high terraces form the system located adjacent to Elephant Hill (EH), Coyote Hill (CH) and Red Hill (RH). Sedimentary exposures, GPR survey locations and paleo-water plane projections from this study are also shown. See text for explanation. (DEM data, British Columbia Government, 1996)
Brassy Creek terrace (Fig. 2.9) is re-interpreted as a fluvial terrace as: (1) the upper surface of the landform is not sloped down toward the valley axis like other deltas or alluvial fans in the region but is near-flat like fluvial terraces (at 420 m asl), (2) the gullies above the landform are too small to have transported the cobble-sized gravel that compose the landform (Fig. 2.9 and Appendix 6.2), (3) there is no source of gravel within the catchment area of these gullies, (4) nearby glaciolacustrine sediments are close to the elevation of the landform which indicates that the basin was too shallow for a delta of this thickness (~60 m) to develop, and (5) fluvial terraces nearby are at a similar elevation. Topographic position and bedrock exposure (Fig. 2.9) in the terrace riser facilitated its preservation during Holocene incision.

GPR surveys reveal two radar facies separated by an angular unconformity (Fig. 2.9; Appendix 2.3c). Radar facies 1 is composed of near-horizontal continuous reflections onlapping from the north (Fig. 2.9c) and northeast (Fig. 2.9b), i.e. accreting onto the valleyward sloping upper surface of radar facies 2. An exposure of imbricate gravel correlative with the position of radar facies 1 indicates westward paleoflow, consistent with a floodplain gravel interpretation for radar facies 1. Onlapping argues against delta topsets building to the north and northeast as inferred by Ryder (1976), and is consistent with a Thompson River terrace interpretation.

Radar facies 2 is likely composed of cobbles that are exposed along the terrace riser (Fig. 2.9). The upper surface dips gently down toward the valley axis and is an unconformity. Internal reflections are variable – they dip gently toward the valley axis (Fig. 2.9b) and are concave and inclined toward the valley wall (Fig. 2.9c). Fluvial terraces in the study area are consistently made of floodplain gravel overlying lacustrine sediments. Unlike all other fluvial terraces observed in the study area that are underlain by lacustrine sediments, this terrace overlies cobble gravel. Sediments of radar facies 2 may have been created by preferential deposition of sediments that were eroded from the nearby Deadman delta (~5 km to the east) during catastrophic westward drainage of the last lake to occupy the valley (section 2.7.1). This is supported by (1) the proximity of this landform to Deadman delta (~5 km), (2) the position of this landform east of a bedrock protuberance, which may have aided in the preferential deposition of sediments transported along the valley to the west (Fig. 2.9), (3) the lower elevation of this landform compared to eroded portions of Deadman delta, (4) the similarity in grain size (cobble-sized) of both the eroded topsets of the Deadman delta and the sediments of radar facies 2, and (5) the possibility that taken together radar reflections of facies 2 may represent gravel bedforms or barforms (Beres et al. 1999) tentatively interpreted as eddy deposits (cf. Baker et al. 1987). The unconformity between radar facies 1 and 2 may
have been created by river erosion proceeding aggradation at this location or may be inherent in the creation of eddy bars.

2.5.4.2. Brassy Creek bench

Brassy Creek bench is a subtle, laterally discontinuous bench located 1 km east of Brassy Creek terrace and at the same elevation (Fig. 2.1). Ryder (1976) interprets it as a wave-cut bench recording the Tranquille stage of Glacial Lake Deadman. My re-examination suggests that Brassy Creek bench is a fluvial terrace remnant because: (1) it is at the same elevation as the Brassy Creek terrace to the east and a dominant fluvial terrace to the west, and (2) the bench occurs and is broader where it intersects gullies unlike other wave cut benches in the region that do not occur in gullies (Fig. 2.5). The bench is a product of gully infilling during aggradation of the Thompson River followed by incision and preservation.

2.5.4.3. Paleo-Bonaparte River system

Gravel benches between three valley-parallel hills west of Ashcroft (EH, CH, RH, Fig. 2.10) have been interpreted as possible delta surfaces because they are at a similar elevation to Fulton’s (1969) projected Glacial Lake Deadman – Tranquille stage water plane (Ryder 1976). Leading to these benches are small, abandoned, gravel lined, river channels (i.e. paleo-channels; A, Fig. 2.10, Appendix 5.1b) and altogether the benches and paleo-channels comprise the paleo-Bonaparte River system (Fig. 2.10). I interpret the benches as shared high fluvial terraces of the paleo-Bonaparte and paleo-Thompson river systems as: (1) the benches are not sloped like other delta surfaces in the study area but are near-flat like fluvial terraces, (2) GPR surveys of limited length and penetration did not find dipping reflections characteristic of Gilbert-type deltas (G8, G9, Fig. 2.10, Appendix 2), (3) the benches lie below the lake water plane reconstructions in this study, (4) the paleo-channels indicate that the Bonaparte River once had three alternative routes adjacent to and between the three valley-parallel hills and thus a fluvial aggradation origin for the between-hill terraces is compatible with multiple routes of the paleo-Bonaparte River, and (5) this area is favourable for significant post-lake fluvial aggradation as the Thompson Valley narrows to the south and the Bonaparte River tributary enters the valley from the north. In addition, (6) the benches did not form as ice marginal terraces as (a) there is a lack of a pitted surface recording kettle holes, (b) there is a lack of ice collapse facies locally in the sediments of the main valley, and (c) the presence of an ice mass in this area of the Thompson Valley, as shown later, is inconsistent with the uninterrupted paleogeography of late glacial lakes determined for
this area (section 2.6.1). These fluvial terraces are the highest in the area and have thicker gravel caps than lower fluvial terraces as they represent the maxima of fluvial aggradation (~455 m asl).

Nearby, high terraces at the mouth of the Bonaparte River are at a similar elevation. Exposures (B, C, Fig. 2.10) show the terraces are composed of imbricate, horizontally stratified gravel. Consequently, these terraces are likely fluvial rather than deltaic in origin and likely developed around the same time as the paleo-Bonaparte river system.

2.6. Lake reconstruction

The paleogeography and quantitative properties of late glacial lakes are reconstructed by correlating water plane indicators, determining lake extents and dams, and integrating this information with DEMs in a GIS.

2.6.1. Correlation of water plane indicators

Primary water plane indicators define two distinct lake levels in the study area: (1) the upper water plane (Fig. 2.11a) and (2) the lower water plane (Fig. 2.11b).

The upper water plane is defined by a tight grouping of wave-cut benches along a first-order surface with an upslope bearing of 332° ±7.7° and slope of 1.8 ±0.6 m km⁻¹ (number of observations (n) = 8, coefficient of determination (r²) = 0.97, at 99.99% confidence; error of bearing and slope at 2 times the standard error; Fig. 2.11a). The standard deviation of the residuals is 4.2 m over a ~45 km baseline (Fig. 2.11c). The excellent fit of these water plane indicators along the plane of best fit, as well as an examination of residuals confirm that a higher order trend surface is not warranted. The elevation of this water plane at selected locations is 516 m asl at Spences Bridge, 570 m asl at Ashcroft, 501 m asl at Kamloops, 566 m asl at McClure, and 489 m asl near Chase. The upper water plane is described by:

\[ Z = -7528.9 - 0.00087855 \cdot X + 0.0015381 \cdot Y \]

The upper water plane is best defined by completely excluding delta surfaces and only correlating wave-cut benches. This is expected given seasonal and inferred water plane elevation biases in the formation of deltas (Table 2.2). Other researchers have also found that deltas are unreliable for accurate water plane reconstructions compared to wave-cut benches (J.A. Rayburn 1997 personal communication 2001; R.J. Fulton personal communication 2001).
Figure 2.11: Best-fit (first order trend surface) definable water plane projections for (a) an upper water plane (Glacial Lake Thompson - High stage), and (b) the lowest water plane (Glacial Lake Deadman - Lowest stage), in the Thompson Basin. The shaded bars indicate the range of water levels for each lake. See text for discussion. (c) Plot of residuals from modeling results for the upper water plane indicate that a first order trend surface is adequate. Water plane indicators located in Figure 2.1 and listed in Table 2.3.
The lower plane is defined by a first-order surface with an upslope bearing of \(321° ± 3.8°\) and slope of \(1.7 ± 0.2\) m km\(^{-1}\) \((n = 6, r^2 = 0.99, \text{ at 99.85% confidence}; \text{ error of bearing and slope at 2 times the standard error, Fig. 2.11})\). The standard deviation of the residuals is 2.4 m over a ~30 km baseline. The elevation of this water plane at selected locations is 440 m asl at Spences Bridge, 479 m asl at Ashcroft, 401 m asl at Kamloops, 448 m asl at McClure. The lower water plane is defined by:

\[
Z = -5923.59 - 0.0011026 \cdot X + 0.001261 \cdot Y
\]

Three wave-cut benches and three delta surfaces were used in this correlation. As deltas are less reliable indicators of lake level, less confidence is given to the accuracy of this correlation than for the upper water plane. Compared to the upper water plane the lower water plane has a similar orientation and a lesser slope. A decrease in slope is expected because the younger lower water plane would have experienced less differential rebound than the older upper water plane (Walcott 1970, Peltier 1994).

2.6.2. Independent support for correlations

Regional relationships, independent studies and the statistical strength of correlations argue for a time-synchronous correlation and development of paleo-water planes. In the absence of dating control, the strength of the correlation of water plane indicators must depend on unique indicators and the statistical significance of trend surfacing models. The statistical significance of both water planes is high. In the nearby Merritt Basin (Fig. 2.1b), the correlations of water plane indicators (Fulton and Walcott 1975) are considered strong as (1) unique paired wave-cut benches that are found throughout the basin aided in correlation, and (2) multiple paleo-water planes have overlapping paleogeographies and similar slopes and orientations (the most reliable results gave a bearing of the upslope direction of 350° and slopes of 1.6 - 1.9 m km\(^{-1}\)). The correlation of water plane indicators in the Thompson Basin could not depend on geomorphic pattern (e.g. paired wave-cut benches) to support correlation. However, as glacio-isostatic tilt is a regional phenomenon, the orientations of the slopes of paleo-water planes located near one another should be similar. This is the case in the Merritt and Thompson basins, lending credence to the water plane correlations presented here. This independent evidence argues for a time-synchronous lake level fall with isostatic tilt.
2.6.3. Lake and stage naming

Projections of the upper and lower water planes defined in my research confirm the existence of two late glacial lakes with separate outlets: (1) Glacial Lake Thompson (upper water plane) had its outlet near Chase, and (2) Glacial Lake Deadman (lower water plane) had its outlet through high elevation lake bottom sediments east of Kamloops. As the orientation and tilt of water planes differ substantially from those previously proposed by Fulton (1969; based on the three-point geometric method) for the same area (Thompson Basin) and my reconstructions correlate more water plane indicators, I do not apply Fulton's stage names to my results.

There are ranges of possible water levels for each lake (Fig. 2.11). The position of well-defined water planes (i.e. stages) within these ranges directs the naming of stages. The upper water plane defines a high stage of Glacial Lake Thompson and the lower water plane defines the lowest stage of Glacial Lake Deadman in the study area (Figs. 2.11, 2.12).

The South Thompson Valley silt (s*, Fig. 2.11) lies between these two lake stages. This indicates that Glacial Lake Deadman – Lowest stage intersected these deposits and therefore this stage did not extend into the South Thompson Valley (section 2.7). The position of the South Thompson Valley silt marks the lowest range of possible water levels for Glacial Lake Thompson and the highest range for Glacial Lake Deadman (Fulton 1969).

Some primary water plane indicators located above, between and below paleo-water planes are not correlated. Only indicator 3a is above the Glacial Lake Thompson, upper water plane. Not mapped are isolated deposits of lake bottom sediments of likely late glacial origin found at various elevations above this water plane throughout the study area. These features indicate that higher undefinable water planes of Glacial Lake Thompson existed. Therefore, the upper water plane is called Glacial Lake Thompson – High stage.

There are insufficient lake level indicators between the Glacial Lake Thompson and Glacial Lake Deadman water planes (6, 7, 9, Fig. 2.11) to correlate intermediate stages. However, their existence suggests progressive or step-wise lowering of the water surface between defined stages.

Primary water plane indicators below Glacial Lake Deadman – Lowest stage; 8a, 8b, 14b, Fig. 2.11) are only located east of Deadman delta (7, Fig. 2.1). They were produced in more extensive early-Holocene stages of Kamloops Lake, dammed behind deposits of Deadman delta after the drainage of the last glacial lake in the basin (section 2.7, Fig. 2.12). These high stages of Kamloops Lake are not late glacial because their
Figure 2.12: Late glacial lake evolution in the southern interior of British Columbia based on this investigation. Two glacial lakes with three glacial lake stages (a-c) are identified in the study area. Glacial lakes and naming outside the Thompson Basin applied from Fulton (1969). Earlier stages of Glacial Lake Merritt and Glacial Lake Penticton are not shown. Also, an early Holocene Kamloops Lake - Cherry Creek stage is not shown as its paleogeography at this scale is quite similar to the earlier occurring Kamloops Lake - High stage (Fulton 1969).
delta surfaces are lower than lake bottom sediments west of Deadman delta and drainage bedforms on Deadman delta (Fig. 2.8, 2.12). Glacial Lake Deadman - Lowest stage was the lowest water level as (1) primary water plane indicators below this stage are completely absent west of Deadman delta, and (2) bedforms (s**, Fig. 2.11) produced during lake drainage also lie below this water plane and above primary water plane indicators east of Deadman delta. The fact that lake bottom sediments, subaqueous fans and river terraces (s, Fig. 2.11) lie below Glacial Lake Deadman - Lowest stage, is consistent with the Glacial Lake Deadman - Lowest stage reconstruction. Fluvial aggradation did not exceed the elevation of the Glacial Lake Deadman - Lowest stage.

There are insufficient lake level indicators between the Glacial Lake Thompson and Glacial Lake Deadman water planes (6, 7, 9, Fig. 2.11) to correlate intermediate stages. However, their existence suggests progressive or step-wise lowering of the water surface between defined stages.

Primary water plane indicators below Glacial Lake Deadman - Lowest stage; 8a, 8b, 14b, Fig. 2.11) are only located east of Deadman delta (7, Fig. 2.1). They were produced in more extensive early-Holocene stages of Kamloops Lake, dammed behind deposits of Deadman delta after the drainage of the last glacial lake in the basin (section 2.7, Fig. 2.12). These high stages of Kamloops Lake are not late glacial because their delta surfaces are lower than lake bottom sediments west of Deadman delta and drainage bedforms on Deadman delta (Fig. 2.8, 2.12). Glacial Lake Deadman - Lowest stage was the lowest water level as (1) primary water plane indicators below this stage are completely absent west of Deadman delta, and (2) bedforms (s**, Fig. 2.11) produced during lake drainage also lie below this water plane and above primary water plane indicators east of Deadman delta. The fact that lake bottom sediments, subaqueous fans and river terraces (s, Fig. 2.11) lie below Glacial Lake Deadman - Lowest stage, is consistent with the Glacial Lake Deadman - Lowest stage reconstruction. Fluvial aggradation did not exceed the elevation of the Glacial Lake Deadman - Lowest stage.

2.6.4. Lake extents and dams

Lake extents are defined by the intersection of water planes, topography and paleo dams (ice and/or sediment). The extents of late glacial lake stages discovered in this study are presented in Figure 2.12. The extent of Glacial Lake Thompson - South Thompson stage (renamed in this study as Highest stage) is after Fulton (1965, 1969; Fig. 2.12a); this study found no primary water plane indicators to evaluate this reconstruction.
Regional topography causes the modern South Thompson and Thompson rivers to flow west and then south into the Fraser River (Fig. 2.1). Thus, a dam in the lower Thompson or Fraser valleys was required for late glacial lakes to have existed within the Thompson Basin. DEM-constrained extrapolation of both defined lake stages place the southern extent of Glacial Lake Thompson – High stage and Glacial Lake Deadman – Lowest stage lakes near Skoonka Creek, 7 km south of Spences Bridge (Fig. 2.1). At Skoonka Creek, lake bottom sediments terminate abruptly and are completely absent in the lower Thompson Valley between Skoonka Creek and Lytton (Fig. 2.1b). These observations suggest that late glacial lakes were dammed in this section of valley by (1) glacial or pre-glacial fill, (2) a large landslide, or (3) an ice mass. A downwasting ice remnant seems the most likely dam because: (1) Fraser Glaciation till overlies pre-Fraser sediments (Anderton 1970, Ryder 1981) and is considerably lower than the defined lake stages, (2) major landslide scars or landslide deposits higher than the elevation of the defined lake stages do not exist in this section of the valley (Ryder 1981), (3) thick, downwasting ice once existed in the valley, according to existing Fraser deglaciation models, and the presence of kettled topography in other sections of the Thompson Valley (Fulton 1969, 1991; Chapters 3, 4), (4) this section of valley is narrow and with steep mountainsides that would have provided shade and possibly colluvium, both facilitating ice preservation, and (5) tributaries to the lower Thompson Valley contain elevated lake bottom sediments (~525 m asl; Ryder 1981) necessitating an ice dam in the Thompson Valley. Previous researchers have also suggested that an ice mass dammed this section of valley (Ryder 1970, Anderton 1970).

Previous research proposed that the head of Glacial Lake Thompson – Niskonlith stage (approximately equivalent to High stage) was at Deadman delta (7, Fig. 2.1; Fig. 2.2) based on the absence of wave-cut benches (3, 4, Fig. 2.1, 10a) to the west of the delta (Fulton 1969). However, the discovery of wave-cut benches in the Thompson Valley at elevations correlative to Glacial Lake Thompson – High stage suggests that this lake extended south of Ashcroft at least 15 km, and possibly to the ice dam south of Spences Bridge (Fig. 2.12b).

Previous research proposed that Glacial Lake Deadman – Tranquille stage (approximately equivalent to Lowest stage) was dammed by ice ~15 km south of Ashcroft (Fig. 2.2c; Ryder 1981). However, the discovery of wave-cut benches in the west portion of the study area corresponding to the elevation of Glacial Lake Deadman – Lowest stage (1, 2, 5c, Figs. 2.1, 2.11b) suggest that this lake extended south of Ashcroft at least 25 km, and possibly to the ice dam south of Spences Bridge (Fig. 2.12c). As high cliffs of lake bottom sediment in the South Thompson Valley (s*, Fig.
extend to a higher elevation than the projected Glacial Lake Deadman – Lowest stage, the eastern extent and outlet for this lake was just east of Kamloops (Fulton 1969).

I agree with Fulton (1969) that there was an ice tongue at the eastern end of the South Thompson Valley and within the Shuswap Basin that provided the eastern margin for stages of Glacial Lake Thompson (Fig. 2.12a,b). Abundant kettled lake bottom sediments near Chase and the inference that the long and deep Shuswap Lake was occupied by an ice mass while lake bottom sediments were accumulating in the adjacent South Thompson Valley both support this inference (Fulton 1965, 1969). Lake bottom sediments in the South Thompson Valley tower >100 m above of the present valley floor.

The western margin of Glacial Lake Thompson – Highest stage was near Kamloops based on (1) an abrupt drop (~100 m drop) in the elevation of lake bottom sediments, (2) tilted lake bottom sediments, and (3) ice marginal terraces. This suggests that an ice dam was present near Kamloops during the time lake bottom sediments were accumulating in Glacial Lake Thompson – Highest stage (Fulton 1965, 1969; Fig. 2.12a).

Previous research identified Glacial Lake Deadman – Durand stage as the lowest late glacial lake level in the study area (Fig. 2.2d, Fulton 1969, Ryder 1976, 1981). The Durand stage had no measurable isostatic tilt (Fulton 1969) and therefore it was inferred to have developed after the period of major glacio-isostatic adjustment following unloading of the CIS (Fulton 1969). A projection of this water plane (375 m asl, Fulton 1969) through the western portion of the study area intersects continuous deposits of lake bottom sediments (tops ~400 m asl) and lies well below the elevation of lake drainage bedforms on Deadman delta (427 m asl, s**, Fig. 2.11). If the water level in Glacial Lake Deadman fell to this stage an eastward flowing river would have formed in the western portion of the study area, not a lake. No gravel terraces record eastward paleoflows. These inconsistencies suggest that the Durand stage of Glacial Lake Deadman is erroneous.

Rather, the landforms attributed to the Durand stage record water levels in a smaller post-glacial lake. These early lake levels of Kamloops Lake formed in the early-Holocene after the drainage of Glacial Lake Deadman and the resulting drainage reversal (Fig. 2.12d). Therefore, Glacial Lake Deadman – Durand stage is renamed Kamloops Lake – High stage; its paleogeography is mapped using no isostatic tilt and elevation data from Fulton (1969; Fig. 2.12d). Detailed surveys of water plane indicators of these early-Holocene lakes were not completed. Deadman delta (Fig. 2.6) obstructed western outflow from Kamloops Lake and controlled the local base level of these high stages of Kamloops
Lake (Fig. 2.12d). Deadman delta continues to control the water level in Kamloops Lake today (Figs. 2.6, 2.12e).

An ice mass in the basin now occupied by Kamloops Lake was probably important to the environment and early development of lakes in the Thompson Valley. Abrupt changes in the elevation of adjacent lake bottom sediments can indicate separate lake basins, as interpreted just east of Kamloops. The only other large contrast in lake bottom sediments in the Thompson Basin is found at the western end of Kamloops Lake. The over-deepening of Kamloops Lake (max. 143 m, Pharo and Carmack 1979) suggests that ice was occupying this section of valley while neighbouring areas accumulated sediment. This may explain the abundance of kettle holes in the neighbouring Deadman delta (Fig. 2.6). Thus, this ice mass may have controlled the extents of early late glacial lakes. However it did not control the extents of lakes by the time Glacial Lake Thompson – High stage developed because well-developed wave-cut benches found in the Kamloops Lake area (10, 11, 13, Fig. 2.1) corresponding to the High stage required sufficient fetch for their creation. Thus, during this time this ice mass was mostly, if not completely, below water level.

2.6.5. Lake paleogeography and quantitative parameters

The paleogeography of Glacial Lake Thompson – High stage and of Glacial Lake Deadman – Lowest stage was mapped by integrating inferred lake extents with DEMs of their stages (from equations 4, 5) and the land surface (TRIM I data, British Columbia Government 1996) in a GIS (e.g. Fig. 2.13). A schematic down-valley cross-section through these lakes is presented in Figure 2.14. Quantitative parameters for these late glacial lakes and their subsequent incision were determined by differencing DEMs of paleolake stages, present topography and estimated lake bottom (Fig. 2.14). Lake bottom was estimated from the highest occurrences of lake bottom sediments. The results are presented in Table 2.5.

2.7. Evolution of late glacial lakes in the Thompson Basin

The history of late glacial lakes in the Thompson Basin has been inferred from a synthesis of information (Fig. 2.12). No dates are available for individual lakes, or their stages. However, higher stages were followed by lower stages because (1) lake base levels were controlled by stagnant, downwasting ice masses and readily eroded lake bottom sediment (Fulton 1967, 1991) (Fig. 2.12), (2) higher delta surfaces developed prior to inset, lower delta surfaces (Fig. 2.6), and (3) lake drainage bedforms that may be associated with the lowest defined paleo-water plane lie just below the elevation of
Figure 2.13: (a) Glacio-isostatically tilted upper water plane (light blue and dark blue; Glacial Lake Thompson - High Stage) and lower water plane (light blue only; Glacial Lake Deadman - Lowest Stage) projected onto a hillshade DEM of the western portion of the study area. (b) Inset showing detail around Ashcroft. This technique allows for accurate mapping of paleo-shorelines (exaggerated at tributaries from Holocene gully ing and incision). Location of ice dam, deltas (green triangles) and wave-cut benches (pink bars) also shown (Fig. 2.1). (DEM data, British Columbia Government 1996)
this paleo-water plane (section 2.7.1).

The South Thompson Valley became ice-free first and was occupied by Glacial Lake Thompson with an eastern outlet along ice margins and topographic divides. The highest stage of this lake was restricted to the South Thompson Valley, impounded by ice on its eastern and western ends (Fig. 2.12a; Fulton 1969). This lake extended westward and lowered as ice backwasted into the Thompson Valley. Lake stabilization and/or high-energy conditions allowed for the development of wave-cut benches that define the later High stage of this lake (Figs. 2.11a, 2.12b). Glacial Lake Thompson - High stage contained ~84 km³ of water (Table 2.5). Continued downwasting and backwasting of ice resulted in continued westward lake growth and lake level lowering. Once the lake level lowered to expose the South Thompson silt, lakes in the Thompson Basin were separated from those in the Shuswap Basin. Glacial Lake Deadman formed with a new eastern outlet near Kamloops - a river incised into South Thompson silt (Fig. 2.12c). The lowest stage of Glacial Lake Deadman occurred after substantial incision (~65 m) into the South Thompson silt (s*, Fig. 2.11, 2.12c). Glacial Lake Deadman - Lowest stage contained ~24 km³ of water (Table 2.5). Lake level lowering between Glacial Lake Thompson - High stage and Glacial Lake Deadman - Lowest stage was likely gradual rather than abrupt because (1) primary water plane features and inset delta surfaces are found between these two water planes (Figs. 2.6, 2.11), and (2) base level was controlled by readily-eroded South Thompson silt.

2.7.1. Catastrophic lake drainage

Glacial Lake Deadman - Lowest stage likely drained westward catastrophically as the ice dam south of Spences Bridge failed. Catastrophic drainage is supported by (1) the absence of glacial lake water plane indicators below the lowest stage (Fig. 2.11), (2) the presence of erosional bedforms on the distal portion of Deadman delta (Fig. 2.8), and (3) theory on the drainage of ice-dammed lakes.

Drainage bedforms may be produced during catastrophic lake drainage (Clague and Rampton 1982, Baker et al. 1987). Asymmetrical bedforms (steeper up-flow slope as inferred from westward lake drainage direction) on the distal portion of Deadman delta are interpreted as high-energy erosional bedforms because: (1) they have a long wavelength (~100 m); (2) delta foresets (inclined radar reflections) are truncated by the land surface (Fig. 2.8b), and (3) topsets (subparallel horizontal radar reflections) are absent (Fig. 2.8b). A minimum water depth of ~16 m is required to make depositional bedforms of the same wavelength as those on Deadman delta (Allen 1982; equation 12,
Carling and Shvidchenko 2002; similar equations are not available for erosional bedforms). Glacial Lake Deadman - Lowest stage was ~26 m above the bedforms.

Table 2.5: Quantitative properties of late glacial lakes and incision in the Thompson Basin

<table>
<thead>
<tr>
<th>Property</th>
<th>Glacial Lake Thompson</th>
<th>Glacial Lake Deadman</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>- High stage (Upper water plane)</td>
<td>- Lowest stage (Lower water plane)</td>
</tr>
<tr>
<td><strong>Water plane</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>orientation (up-plane) (°)</td>
<td>332</td>
<td>321</td>
</tr>
<tr>
<td>tilt (m km(^{-1}))</td>
<td>1.8</td>
<td>1.7</td>
</tr>
<tr>
<td><strong>Volume (km(^3))</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>of lake</td>
<td>83.6</td>
<td>23.6</td>
</tr>
<tr>
<td>of drainage(^2)</td>
<td>--</td>
<td>20.0</td>
</tr>
<tr>
<td><strong>Surface area (km(^2))</strong></td>
<td>801</td>
<td>414</td>
</tr>
<tr>
<td><strong>Length (km)</strong></td>
<td>1.5 - 6.7</td>
<td>1.5 - 4.0</td>
</tr>
<tr>
<td><strong>Max depth (m)</strong></td>
<td>198</td>
<td>80</td>
</tr>
<tr>
<td><strong>Typical depth (m)</strong></td>
<td>~140</td>
<td>~50</td>
</tr>
<tr>
<td><strong>Incision volume(^1) (km(^3))</strong></td>
<td>--</td>
<td>14.1</td>
</tr>
</tbody>
</table>

\(^1\) All calculations account for the isostatic deformation of representative water planes. The Kamloops Lake area may have been occupied by ice during the time of these lakes and therefore the volume (3.7 km\(^3\)) and depth (71 m mean depth, 143 m maximum depth) of Kamloops Lake were not used in the calculations.

\(^2\) Significant incision occurred following drainage of Glacial Lake Deadman during the Holocene. Ignoring suspended sediment loads passing through Kamloops Lake, only valley-fill west of Kamloops Lake would have been transported out of the region because Kamloops Lake is a sink for the North and South Thompson rivers, and Glacial Lake Shuswap was a sirk for these rivers before the drainage reversal to present. This gives ~8 km\(^3\) of valley-fill that was eroded and transported out of the section of valley west of Kamloops Lake to Skoonka Creek. The remaining 6.1 km\(^3\), was deposited in Kamloops Lake and delivered to Glacial Lake Shuswap and is a low estimate as not all of the North Thompson Valley was included in the calculations.

Various draining mechanisms have been proposed for ice-dammed lakes, including overspill, tunnel enlargement, ice-dam flotation and seismic events (Walder and Costa 1996, Tweed and Russell 1999, and references therein). It is difficult to discern which of these mechanisms resulted in the failure of the ice dam south of Spences Bridge. Failure by dam flotation may not have been possible, as colluvium from the adjacent, steep slopes may have caused the weight of the dam to exceed the buoyant force. Only a small amount of rock debris (as little as 4% by volume) is needed to prevent ice flotation (Tweed 2000, Chapter 4). Failure of the ice dam by overspill may have occurred by lowering of the ice dam by ongoing ablation, or during a short-lived increase in lake level.
Figure 2.14: Cartoon of down-valley projections of water planes and inferred lake bottom elevation from Spences Bridge to Chase (Fig. 2.1). Changes in slopes to the water plane result from changes in valley orientation. The North Thompson Valley is not shown. ~20.0 km$^3$ of water (Table 2.5) in the lake forming the lowest stage of Glacial Lake Deadman drained catastrophically once the ice dam south of Spences Bridge failed (bold arrow). An additional ~3.6 km$^3$ of water dammed behind the Deadman delta (a) was likely not part of this lake drainage.

When the ice dam in the lower Thompson Valley failed sometime before 8,900 ±150 $^{14}$C yr BP (GSC-193, Dyck et al. 1965, Fulton 1969, Table 2.1), an estimated 20 km$^3$ (Table 2.5) of water drained catastrophically into the Fraser River system. It is possible that this event may have triggered the failure of other glacial lakes downstream or upstream in the Fraser River system. Eventually the floodwaters would have reached the coast and flowed into the Strait of Georgia, a total distance of ~250 km (Fig. 2.15a). This event resulted in the regional drainage reversal of the Thompson and South Thompson rivers, and the capture of the Thompson Basin (presently about 54,000 km$^2$ in area, Environment Canada 1989) by the Fraser River system.

2.7.2. Offshore outburst flood deposits

Exotic muds containing reworked Tertiary microfossils have been identified in drill core from the Strait of Georgia and Saanich Inlet (Blais-Stevens et al. 2001, 2003, Conway et al. 2001). These muds are inferred to record a major flood event or events from the British Columbia mainland that occurred between ~9,200 $^{14}$C yr BP and 10,800 $^{14}$C yr BP (Fig. 2.15b; Blais-Stevens et al. 2001, 2003, Conway et al. 2001) because altogether: (1) they contain abundant reworked Tertiary microfossils of mainland provenance, (2) texturally and structurally the muds are anomalous in core, are normally graded, and have a sharp, non-erosional basal contact and gradational upper contact, (3) they lack bioturbation structures except at the base of one deposit (possible invertebrate escape structures), (4) they have a well-developed brackish-water signal in the diatom assemblage, and (5) they were deposited during the same period late glacial lakes on the mainland disappeared (Blais-Stevens et al. 2001, 2003, Conway et al. 2001). Cores in the Saanich Inlet show there are two exotic mud deposits separated by tens of
Figure 2.15: (a) Flood routes during the drainage of Glacial Lake Deadman (D) and Glacial Lake Fraser (F). Location of marine cores recording outburst flood deposits in Strait of Georgia (G, Conway et al. 2001) and Saanich Inlet (S, Blais-Stevens et al. 2001). (b) Marine flood deposits (exotic mud) emplaced between 9,200 $^{14}$C yr BP and 10,800 $^{14}$C yr BP. (DEM imagery, British Columbia Government 1996)
centimetres of sediment (Blais-Stevens et al. 2001, 2003). The upper deposit has a
gradational upper and lower contact. Cores in the Strait of Georgia show only one exotic
mud deposit. Some variation in the dates between the Saanich Inlet and Strait of Georgia
deposits (Blais-Stevens et al. 2001, Conway et al. 2001; Fig. 2.15b) may indicate that
these deposits record separate flood events. These deposits may be linked to the
catastrophic drainage of Glacial Lake Deadman – Lowest stage (G, S, Fig. 2.13). Glacial
Lake Fraser (Clague 1987, Huntley and Broster 1997) was of sufficient volume and may
have drained catastrophically to produce the Strait of Georgia deposit. The
paleogeography and drainage style and timing of Glacial Lake Fraser are poorly
constrained. It is possible that both lakes played a role in creating these deposits.

2.7.3. Post-glacial lake events

Following the drainage of Glacial Lake Deadman and the regional drainage
reversal, the high stages of Kamloops Lake developed behind the raised Deadman delta
(Fig. 2.12d). The ancestral Thompson River incised down through ~150 m of valley-fill to
bedrock, deposited gravel and developed numerous fluvial terraces. Incision to within a
few metres of present river level was achieved by the mid-Holocene (Ryder 1981; 6730
14C yr BP, Hallett et al. 1997). As the maxima of fluvial aggradation was somewhat close
to Glacial Lake Deadman – Lowest stage, the highest fluvial terraces have sometimes
been mistaken for former deltas (section 2.5.4). During and since the Glacial Lake
Deadman outburst flood event, >14.1 km³ of valley-fill has been eroded from the study
area (Table 2.5). Most of this eroded fill was glaciolacustrine silt and fine sand (Chapter
3).

2.8. Glacio-isostasy

To the best of my knowledge the isostatic tilt of glacial lakes (Table 2.5) recorded
within this region is among the highest of any glacial lakes in the world. For example,
lakes associated with the Laurentide Ice Sheet typically have their shorelines tilted 0.5 to
1.0 m km⁻¹ (Teller and Thorleifson 1983). The high isostatic tilt and direction of tilt of
these lakes was likely influenced by (1) the mantle and lithosphere properties, (2) paleo-
topography of the CIS, (3) rapid deglaciation, (4) the age of the lakes, (5) the size of the
lakes, and (6) tectonic activity.

2.8.1. Mantle and lithosphere properties

Geophysical studies have revealed that the lithosphere over much of the southern
Cordillera is thin (~33 km, Clowes et al. 1995). In addition, modelling of late glacial
isostatic and eustatic changes along the southern coast of British Columbia indicate that the mantle is of low viscosity (Clague and James 2002). This inference is also supported by the high heat flow within the crust, which acts to lower viscosity (Hyndman and Lewis 1995). These properties are related to the fact that the Cordillera region is host to a major subduction zone just offshore. These unique conditions describe a crust that would have been very responsive to loading by the CIS. Isostatic depression along the southern coast of the Cordillera is up to 300 m (Clague and James 2002). As the indicators used to derive this value developed after the CIS had decreased in size, and isostatic uplift was extremely rapid (Muhs et al. 1987), isostatic depression may have been as high as 400-500 m at glacial maximum (Clague and James 2002). It is likely that glacio-isostatic depression was also high in the southern interior of British Columbia. This can be estimated by scaling ice thickness differences between the coast (1930 m max thick) and the southern interior (1630 m max thick; Wilson et al. 1958, James et al. 2000). A scaling factor of 0.84 gives ~420 m for maximum glacio-isostatic depression in the southern interior. It is possible that this may have been greater as the southern interior was once the centre of the CIS (J.T. Teller personal communication 2003). Such high values for glacio-isostatic depression could easily account for the extreme water plane deformation modeled for the Thompson Basin.

2.8.2. CIS paleo-topography

The paleo-topography of the CIS caused differential loading and rebound of the crust. Areas that experienced greater crustal load generally record the greatest rebound, assuming crustal homogeneity. The tilt directions of water planes in the Thompson and Merritt basins suggest that crustal load was greatest to the north-northwest (330° - 350° azimuth) and thinner to the south-southeast. This direction corresponds with a previously inferred ice divide for the CIS to the northwest (Prest et al. 1968, Clague 1989, Ryder et al. 1991). It is interesting to note that the largest deltas in the Thompson Basin are found along the northern side of the valley in the Kamloops Lake area and the western side of the North Thompson Valley and all share the same general headwaters. The large size of some of these deltas (e.g. Deadman delta is 13 km²), in addition to their biased distribution, suggests the possibility that ice was within their headwaters, generally to the north, supplying large amounts of meltwater and sediment to these tributaries and late glacial lakes within the Thompson Basin. As areas of thinner ice deglaciate sooner than areas of thicker ice, it is expected that the last area to deglaciate on the plateau would be generally near the ice divide. In addition, the orientation and topographic setting of wave-cut benches within the Thompson Basin indicate that the winds that created these
landforms were generally from the north, likely produced by cold-air drainage from ice masses situated to the north. Finally, a general northward retreat due to downwasting of the CIS has been inferred mostly from the pattern of meltwater channels on the plateau (Fulton 1967). Thus, these four separate lines of evidence suggest the CIS was thicker north and northwest of the Thompson Basin.

2.8.3. Rapid deglaciation and relative timing of lake development

The CIS decayed rapidly following the Fraser Glaciation maximum (Porter and Swanson 1998, Clague 1981, Clague and James 2002). The sooner a late glacial lake develops following the glacial maximum (i.e. maximum loading), the longer the period of isostatic adjustment the paleo-shorelines will record. As rebound declines exponentially (Muhs et al. 1987, Peltier 1994), early developing lakes could have much higher isostatic tilts than those developed later. Thus, rapid deglaciation of the CIS coupled with a possible early development of these late glacial lakes may also account for their high isostatic tilt.

2.8.4. Size of the lakes

The rate of rebound of the crust is also related to the size of the lakes that developed as the ice sheet decayed. Large lakes load the crust more. If a large lake developed then the crust would not rebound as quickly and its shorelines would be less tilted. As the CIS late glacial lakes were very small compared to the former ice mass, rebound rates were faster than if the lakes were larger.

2.8.5. Tectonic activity

Finally, tectonic activity may have affected the rate, amount and pattern of rebound. A major subduction zone lies just offshore resulting in considerable tectonic activity concentrated along the coast of the Cordillera. Although present earthquake activity is relatively low in the Interior compared to the coast, it was likely significantly higher during deglaciation considering that there was hundreds of metres of glacio-isostatic rebound, and that fault instability is promoted during deglaciation (Lagerbäck and Witschard 1983, Johnston 1989, Muir-Wood 1989, Wu and Hasegawa 1996, Stewart et al. 2000). Extensive sub-aqueous failure deposits from glacial lakes in the Thompson Valley may owe their genesis to late glacial earthquake activity, although this remains speculative (Chapter 3). Water plane reconstructions have been used elsewhere to detect post-shoreline fault movement (e.g. Firth and Stewart 2000 and references therein).
There are too few water plane indicators in the Thompson Basin to detect fault movements using this technique.

2.9. Conclusions

Detailed study of late glacial lakes in the Thompson Basin reveal a number of significant findings. These results greatly improve our understanding of deglacial style, and lake stages, and extend the database for Cordilleran rebound:

1. Cordilleran late glacial lakes, in contrast to Laurentide late glacial lakes, are ribbon-shaped. These ribbon lakes develop in response to topography and deglacial style. As the unique paleogeographic properties of ribbon lakes would affect the style of sediment delivery and dispersal, and energy within these lakes, it is likely that their sedimentary environments are equally as unique. The findings of this study are used as the basis for interpreting their detailed sedimentary record (Chapter 3).

2. During the decay of the CIS, two late glacial lakes developed within the Thompson Basin: Glacial Lake Thompson and Glacial Lake Deadman. The highest stage of Glacial Lake Thompson was confined to the South Thompson Valley. The paleo-water plane of this lake is poorly defined. Ice recession and lake lowering led to the development of the well-defined High stage of Glacial Lake Thompson. Continued lake lowering below the level of lake bottom sediments in the South Thompson Valley isolated the lakes in the Thompson Basin from those in the Shuswap Basin; Glacial Lake Deadman developed in the Thompson Basin. Water level indicators in this lake mainly record its lowest stage. This lake drained catastrophically into the Fraser River system. A lower Durand stage (Fulton 1969) of Glacial Lake Deadman did not exist. Rather, a Holocene High stage of Kamloops Lake is identified and confined east of Deadman delta. These stages need not record standstills, but rather water plane indicators may have formed under favourable energy and sediment supply conditions. Higher lake levels may have existed within the Thompson Basin but left little record of their existence. Intermediate lake levels are recorded by a few indicators.

3. Late glacial lakes within the Thompson Basin developed in contact with receding valley ice masses. An ice dam south of Spences Bridge dammed the waters in the Thompson Basin and was responsible for the development of lakes and drainage reversals. This dam eventually failed causing catastrophic drainage of Glacial Lake Deadman – Lowest stage. Catastrophic flow within the basin eroded the distal portions of the valley-confining
Deadman delta and developed drainage bedforms. ~20 km$^3$ of floodwaters travelled ~250 km into the Fraser River system and terminated in the marine environment, depositing exotic mud sometime between ~9,200 $^{14}$C yr BP and 10,800 $^{14}$C yr BP.

4. Glacio-isostatic deformation of the paleo-water planes of these lakes is among the highest in the world (1.7 – 1.8 m km$^{-1}$) and reflects the thin lithosphere, the low viscosity mantle, the paleo-topography of the CIS, rapid deglaciation, possible early development of these lakes, the size of these lakes and possible tectonic activity. The orientation of their tilt (332° and 321°) supports the notion of greatest crustal load and consequentially greatest rebound, to the north to northwest.

5. Ice was likely present within the plateau to the north delivering water and sediment to these lakes. This indicates that, at least locally, the plateau did not become completely ice-free prior to the valley.
References


Clague, J.J. and James, T.S. 2002 History and isostatic effects of the last ice sheet in southern British Columbia. Quaternary Science Reviews 21: 71-87.


Paleoenvironment
Chapter 3
The environment in and around late glacial ribbon lakes, Thompson Valley, British Columbia

3.1. Introduction
During the decay of the Cordilleran Ice Sheet (CIS), approximately 10 to 12 ka
$^{14}$C BP, numerous ribbon lakes developed within the valleys incised into the Interior
Plateau of British Columbia (Chapter 2, Fulton 1969). Sediments from these lakes form
the dominant valley fill in the region. Throughout the interior these sediments are
generally referred to as the "White Silts" (Dawson 1879), and at more local scales they
are named after the valleys (e.g. "South Thompson silt", Fulton 1965). These lake
sediments are well exposed and are important records of environments during
deglaciation. However, our understanding of these lakes remains limited as previous
studies of deglacial environments have been (1) limited in scope (e.g. stratigraphic rather
than sedimentologic, reconnaissance mapping; excepting Fulton 1965, Shaw 1977, Shaw
and Archer 1979), (2) limited in detail (e.g. few exposures mapped or reported), or (3)
limited by an incomplete understanding of lake paleogeography.

I build on previous studies by exploring the deglacial environment in and around
ribbon lakes in the Spences Bridge-Ashcroft-Savona corridor (hereafter referred to as the
Thompson Valley or study area). I focus on four questions. (1) What types of ribbon lakes
existed (e.g. ice-contact vs. non-ice-contact)? (2) What were the dominant modes of
sediment delivery and deposition in these lakes? (3) What were the sedimentary
environments of these lakes? (4) What deglacial style was associated with these lakes?
The Thompson Valley offers an opportunity to address these questions as sedimentary
exposures are numerous, deglacial lake sediments are clearly identified in the
stratigraphic sequence (Clague 2000), and the paleogeographic evolution of these lakes
is now well understood (Chapter 2).

3.2. Study area and previous research
The study area is situated within the southern Canadian Cordillera, in the
southern Interior Plateau of British Columbia. It is a 75 km corridor of the Thompson
Valley from the outlet of present day Kamloops Lake at Savona to Spences Bridge (Fig. 3.1). The study area lies on the western portion of the Thompson Plateau physiographic region, within the rain shadow of the Coast Mountains to the west, and neighbours the Fraser Plateau to the northwest and the Cascade Range to the south (Holland 1964). It is an area of moderately high relief.

The Thompson Plateau is underlain by volcanic, plutonic, sedimentary and metamorphic rocks (Cockfield 1948, Duffel and McTaggert 1952). Fluvial and glacial erosion have dissected the Thompson Plateau resulting in deep valleys (1400 to 1600 m relief), today partially filled with thick (100 to >150 m) Quaternary deposits. There are numerous tributaries to the Thompson River. The largest of these are the Deadman, Bonaparte and Nicola rivers (Fig. 3.1). Cliffs of late glacial lake bottom sediments occur in a continuous swath (~75 km long) along the valley from the outlet of Kamloops Lake to a few kilometres south of Spences Bridge near Skoonka Creek (Fig. 3.1). These late glacial lake sediments form the dominant valley fill. Lake bottom sediments extend from ~400 m asl to below river level. Well-log records indicate they extend >50 m below present river level in some locations. Numerous river terraces traverse the valley-sides greater than 100 m above present river level. The arid climate has resulted in minimal vegetation. As a result, there are abundant sediment exposures and landforms are relatively easy to identify.

Surficial geology mapping (Ryder 1976, 1981), and stratigraphic interpretation (Fulton and Armstrong 1965, Ryder 1976, 1981, Clague 2000) have been the thrust of Quaternary research in the area. Late glacial lake sediments are easily identified as they frequently underlie Holocene river terraces (gravel) and overlie till of the Fraser Glaciation (oxygen isotope stage 2; Ryder 1976, Clague 2000, Fig. 3.2).

Deglaciation is inferred to have proceeded by regional downwasting and stagnation of the CIS accompanied by frontal retreat resulting in the formation of deglacial lakes dammed by remnant ice masses within the valleys of the Cordilleran Interior (Fulton 1967, 1969, 1991). This model requires further testing.

Deposits within the study area formed in at least two ribbon-shaped deglacial lakes that occupied, and extended beyond the study area: Glacial Lake Thompson and Glacial Lake Deadman (Fig. 3.1a; Chapter 2). Glacial Lake Thompson was deeper (~140 m) than Glacial Lake Deadman (~50 m deep) and is older. Both lakes were dammed by ice south of Spences Bridge and drained eastward from outlets located in the Shuswap Basin and at Kamloops (Fig. 3.1c), respectively (Fulton 1969, Chapter 2). Glacial Lake Deadman (the last deglacial lake in the area) drained catastrophically when the ice dam failed (Chapter 2). Drainage then reversed, flowing west and south as it does today. The
Figure 3.1: (a) Hillshade DEM showing locations of sedimentary sections and paleoshorelines (Chapter 2). (b) Location of study area in British Columbia, Canada. (c) Regional physiographic context and locations mentioned in text. Map abbreviations: Li = Lillooet, Ly = Lytton, SB = Spences Bridge, M = Merritt, Ash = Ashcroft, CC = Cache Creek, Sv = Savona, K = Kamloops, Mc = McClure, Ch = Chase, SA = Salmon Arm, Ar = Armstrong, and V = Vernon. Th.R = Thompson River, N.Th.R = North Thompson River, and S.Th.R = South Thompson River. (DEM data, British Columbia Government 1996)
Figure 3.2: The stratigraphic context of lacustrine sediments (unit 6) examined in this study (Bonaparte Section, near 13, Fig. 3.1). Stratigraphic position implies that these lacustrine sediments are deglacial in origin (Clague 2000, Clague and Evans 2003) following the Fraser Glaciation age (oxygen isotope stage 2). This section records an inset valley-fill stratigraphy. Elsewhere, unit 6 sediments typically occur in units much thicker (>80 m) than at this section, and form the dominant valley fill. Lithofacies variability in unit 6 is shown in Appendix 3.3.
former shorelines of both lakes have been tilted (up to 1.8 m km\(^{-1}\)) by differential glacio-isostatic uplift resulting today in sloped projections of these former lake surfaces (Chapter 2).

Some paleoenvironmental interpretations have been made from limited study of lake sediments. In the southern portion of the study area (Anderton 1970, Ryder 1970) it has been inferred that lakes lengthened in contact with receding valley ice tongues and that fining-upward trends in sedimentation reflect deposition at ever increasing distances from retreating ice. Ryder (1970) described one deglacial lacustrine section and found that some sediments were emplaced by powerful density currents, others by slumps and mudflows. Sediments are not readily interpreted as varved. East of Kamloops (Fig. 3.1c), paleoenvironmental research on Glacial Lake Thompson (South Thompson stage – somewhat equivalent to Highest stage, Chapter 2) indicates that the lake lengthened as ice tongues receded from both ends of the basin, sediments were derived from plateau areas, and rates of sedimentation were high (Fulton 1969).

Detailed sedimentological studies critical for paleoenvironmental reconstruction have not been completed for the Thompson Valley. In this paper, I (1) infer lake type, (2) identify glaciolacustrine lithofacies, (3) discuss the sedimentary environment of ribbon lakes, and (4) explore the implications of these conclusions for the local style of deglaciation of the CIS.

3.3. Methods

Sedimentary exposures were selected to best explore spatial patterns of sedimentation related to tributaries (stream inputs via deltas and subaqueous fans) and tributary-free sections of the valley (potential inputs from valley ice tongues). A total of twenty-four sediment sections were examined. Ten of these exposures (>20 m in height) were logged at the centimetre to decimetre scale. The remaining fourteen exposures were examined in a stratigraphic manner as they were either too high or unsafe (Appendix 3.7); some portions of these exposures were examined in detail. Sediment texture and structure, paleoflow indicators, bed contact relationships, and lateral continuity and thickness of units were recorded. Clay fractions were quantified by sedigraph analysis.

Lithofacies are defined as the smallest homogenous sedimentary unit representing a discrete sedimentary process (e.g. Reading 1986). Lithofacies are identified and interpreted based on their characteristics and associations, and their geomorphic (landform) and paleogeographic (e.g. location with respect to possible sediment sources) context. Interpretation of lithofacies and landforms follow sedimentary principles (e.g.
Middleton and Hampton 1976, Allen 1982, Boggs 1987) and examples in the literature (Table 3.1). Landform genesis and deglacial style are inferred from landform-sediment relationships.

3.4. Lake type

The paleogeographic requirement for ice dams (Chapter 2) and the stratigraphic relationships support an ice-contact origin for the late glacial ribbon lakes in the Thompson Valley. The lakes were dammed by ice and glacial ice was buried within their sediments. However, these lakes received much of their sediment supply from outwash streams based on their geomorphic and sedimentologic relationships. Such lakes have also been inferred elsewhere in the southern interior of British Columbia (Lesemann and Brennand 2003) and in the Yukon Territory (Ward and Rutter 2000).

Evidence for the presence of large valley ice masses and buried ice within the Thompson and Shuswap basins during the time of Glacial Lake Thompson, Glacial Lake Deadman and Glacial Lake Shuswap includes: (1) the depth (max. 150 m) of Kamloops Lake and Shuswap Lake (Fig. 3.1c) today, (2) the presence of kettle holes adjacent to and within lake boundaries (Chapter 4), (3) abrupt, large changes in the thickness of neighbouring lake bottom sediments, and (4) the presence of ice-marginal terraces near Kamloops (Fulton 1965).

Stratigraphic relationships between glacial and late glacial sediments (Figs. 3.2, 3.3) indicate that the first late glacial lake in the Thompson Valley (Glacial Lake Thompson) likely developed in contact with ice. Typically, late glacial lacustrine sediments overlie discontinuous Fraser Till that in turn overlies bedrock or older
Table 3.1: Lithofacies descriptions, occurrence and interpretation for late glacial ribbon lakes in the Ashcroft area, British Columbia

<table>
<thead>
<tr>
<th>Code</th>
<th>Facies</th>
<th>Lithofacies name and log symbol</th>
<th>Characteristics</th>
<th>Occurrence and distribution</th>
<th>Process</th>
<th>Interpretation</th>
<th>Environment(s)</th>
<th>Refs</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>CI</td>
<td>Laminated clayey-silt</td>
<td>Laminated clayey-silt (mm to cm scale thickness) and fine sand (mm to cm scale thickness)</td>
<td>Very rare; most common near Deadman delta</td>
<td>Deposition from suspension; likely derived from underflow events (see text for explanation)</td>
<td>Lake bottom</td>
<td>3, 15, 31, 37</td>
<td></td>
</tr>
<tr>
<td></td>
<td>C1n</td>
<td>Clayey-silt with inclusions</td>
<td>Clayey-silt (convoluted) including irregular clasts of silt to medium sand; beds 5-25 cm thick</td>
<td>Very rare; most common near Deadman delta</td>
<td>Resedimentation by loading and subaqueous flow or slump</td>
<td>Lake bottom</td>
<td>1, 30</td>
<td></td>
</tr>
<tr>
<td>Zl</td>
<td></td>
<td>Laminated silt</td>
<td>Laminated (planar to undulatory; grading sometimes visible) silt with minor fine sand separations</td>
<td>Very common; found throughout area</td>
<td>Deposition from suspension</td>
<td>Lake bottom; occasionally in subaqueous fans</td>
<td>3, 12, 15, 31</td>
<td></td>
</tr>
<tr>
<td>Zn</td>
<td></td>
<td>Silt with sand inclusions</td>
<td>Massive silt2 to highly contorted laminated silt including folded and rolled coarser-grained saucer to irregular inclusions; cosets 0.5-55 m thick; inclinations (0.05 m - 10 m apparent long-axis) are composed of contorted laminated silt and diffusely graded sand; mega-inclusions sometimes include contorted ripple cross-laminated sand; this lithofacies is sometimes associated with tilted blocks of lacustrine sediment</td>
<td>Common; found throughout area</td>
<td>Subaqueous slumping and soft sediment loading</td>
<td>Lake bottom</td>
<td>1, 8, 9, 16, 17, 19</td>
<td></td>
</tr>
<tr>
<td>Zc</td>
<td></td>
<td>Convolute silt</td>
<td>Silt with convolutions, load, fold and drag structures; sometimes massive2, sometimes coarser grains incorporated; cosets 0.1-15 m thick</td>
<td>Somewhat common throughout area</td>
<td>Subaqueous slumping and soft sediment deformation</td>
<td>Lake bottom</td>
<td>1, 9, 19, 29</td>
<td></td>
</tr>
<tr>
<td>Srs</td>
<td>type S</td>
<td>Ripple cross-lamination: type S</td>
<td>Sinusoidally laminated very fine sand and coarse silt, near vertical angle of climb; cosets 0.5-20 cm thick; commonly drapes type B ripple cross-laminated sand</td>
<td>Somewhat common; found near Deadman delta and tributaries</td>
<td>Mainly suspension deposition from underflows forming ripples</td>
<td>Lake bottom; subaqueous fan</td>
<td>1, 4, 5, 16, 24, 35, 37</td>
<td></td>
</tr>
<tr>
<td>Srb</td>
<td>type B</td>
<td>Ripple cross-lamination: type B</td>
<td>Sloos-side and lee-side cross-laminae preserved in fine sand; moderate angle of climb, variable amplitude; variable vertical associations - type B overlying type A most common; cosets 0.1-1 m thick</td>
<td>Somewhat common; found near tributaries</td>
<td>Deposition from suspension and traction from underflows forming ripples</td>
<td>Lake bottom; subaqueous fan; delta (foresets)</td>
<td>1, 4, 16, 24, 27, 35, 37</td>
<td></td>
</tr>
<tr>
<td>Sra</td>
<td>type A</td>
<td>Ripple cross-lamination: type A</td>
<td>Lee-side cross-laminae preserved in fine to medium sand; low angle of climb, variable amplitude; lower contact erosional and sometimes loaded; most common ripple type; cosets 0.05 - 1 m thick</td>
<td>Somewhat common; found near tributaries</td>
<td>Deposition mainly from traction from underflows forming ripples</td>
<td>Lake bottom; subaqueous fan; delta (foresets)</td>
<td>1, 4, 16, 24, 27, 35, 37</td>
<td></td>
</tr>
<tr>
<td>Sp</td>
<td></td>
<td>Planar stratified sand</td>
<td>Planar stratified fine to coarse sand with minor granule fraction; sometimes contains irregular-shaped clasts of silt; cosets 0.5 - 20 m thick</td>
<td>Somewhat common; found near tributaries</td>
<td>Mostly deposition from traction from underflows</td>
<td>Subaqueous fan; delta (foresets); lake bottom</td>
<td>1, 2, 5, 7, 8, 11, 14, 14, 22, 24, 25, 27, 28, 32, 35</td>
<td></td>
</tr>
<tr>
<td>Facies Code</td>
<td>Lithofacies name and log symbol</td>
<td>Characteristics</td>
<td>Occurrence and distribution</td>
<td>Process</td>
<td>Interpretation</td>
<td>Environment(s)</td>
<td>Refs[^1]</td>
<td></td>
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<tr>
<td>Sdg</td>
<td>Diffusely graded sand</td>
<td>Diffusely graded (sometimes normal) medium and coarse sand sometimes with occasional granules; lower contact sometimes loaded; occasional dish structures; cosets 0.1 - 20 m thick.</td>
<td>Somewhat common; found near tributaries</td>
<td>Mostly rapid suspension deposition from hyper-concentrated and turbidity flows; fluidization occasional</td>
<td>Lake bottom; subaqueous fan</td>
<td>1,2,5,7,11,13,14,21,22,24,25,26,27,28,31,33</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Srg</td>
<td>Reverse graded sand</td>
<td>Reverse graded medium and coarse sand sometimes with occasional granules; lower contact sometimes loaded; occasional dish structures; cosets 0.1 - 0.5 m thick</td>
<td>Uncommon; found near tributaries</td>
<td>Mostly deposition from grain flow; fluidization occasional</td>
<td>Lake bottom; subaqueous fan</td>
<td>1,14,30,33</td>
<td></td>
<td></td>
</tr>
<tr>
<td>St</td>
<td>Trough cross-bedded sand</td>
<td>Trough cross-bedded sand; cosets 0.5 - 2 m thick</td>
<td>Uncommon; found near Ashcroft and Deadman delta</td>
<td>Deposition from traction from underflows forming dunes</td>
<td>Subaqueous fan; delta (foreset)</td>
<td>1,5,11,27,31,33,34</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Gus</td>
<td>Undulatory stratified gravel</td>
<td>Undulatory, stratified, sub-rounded to well-rounded clasts with small cobbles dominant; lenticular beds; cosets 3 - 15 m thick</td>
<td>Uncommon; found near Ashcroft</td>
<td>Deposition from traction and suspension from underflows forming in-phase waves</td>
<td>Subaqueous fan</td>
<td>1,6,7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>GI</td>
<td>Trough cross-bedded gravel</td>
<td>Trough cross-bedded gravel; cosets 2 - 15 m thick</td>
<td>Uncommon; found near Ashcroft</td>
<td>Deposition from traction and suspension from underflows forming dunes</td>
<td>Subaqueous fan</td>
<td>1,5,11,27,31,33,34</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Gi</td>
<td>Imbricate gravel</td>
<td>Stratified (sometimes weakly stratified), imbricate, matrix supported, subrounded to rounded pebble to large cobble; beds 2 - 5 m thick</td>
<td>Uncommon; found near Ashcroft, Deadman delta and Section 20</td>
<td>Deposition from traction from underflows and avalanching</td>
<td>Delta (foreset); subaqueous fan</td>
<td>13,25,27,31,33,37</td>
<td></td>
<td></td>
</tr>
<tr>
<td>zD</td>
<td>Stony silty diamicton</td>
<td>Matrix supported sub-angular to sub-rounded clasts (granule to small cobble); &lt;40% clasts; silt to fine sand matrix; beds &lt;30cm thick</td>
<td>Uncommon; near lake edges and tributaries</td>
<td>Subaqueous debris flow</td>
<td>Lake bottom; subaqueous fans</td>
<td>10,11,20,22,24,26,34</td>
<td></td>
<td></td>
</tr>
<tr>
<td>bD</td>
<td>Boulder diamicton</td>
<td>Matrix or clast supported sub-angular to angular boulders (up to 2 m size); silt to fine sand matrix; &gt;70% clasts; clast lithology traced to neighbouring outcrop; beds 0.5 - 15 m thick</td>
<td>Uncommon; found near Ashcroft</td>
<td>Subaqueous rockfall</td>
<td>Lake bottom; proximal to bedrock cliff</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>


[^2]: Hard, dry silt appears massive. Thin-sections and micromorphological analysis not attempted.
Quaternary sediments; contacts between stratigraphic units are sharp (Fig. 3.3). In some exposures, discontinuous lenses (<3 m thick) of fluvially sorted gravel separate till and lacustrine sediments (Fig. 3.3). The lowest 5 m of lacustrine sediments may contain thin (<10 cm) interbeds of diamicton (Fig. 3.3). These stratigraphic relationships suggest that as soon as ice melted from a location within the Thompson Valley, a lake developed in which sediments accumulated. There is no evidence of subaerial exposure or the presence of an outwash plain following till deposition within the reconstructed lake boundaries. Gravel lenses and diamicton beds record proximal deposition in an ice-contact lake. As the fluvially sorted gravel forms thin lenses it does not represent a persistent outwash plain.

3.5. Glaciolacustrine lithofacies

Seventeen glaciolacustrine lithofacies were identified within the late glacial ribbon lakes of the study area (Table 3.1). Examples of some lithofacies are presented in Figure 3.4. They range in grain sizes from clayey-silt to boulder. Laminated silt (ZI, Table 3.1; Fig. 3.4a, b) is the most common lithofacies and is typically associated with low-energy lake bottom environments that are generally (1) sheltered from the direct influence of tributaries, or (2) present at a relatively high elevation within the lacustrine fill and thus record the latest lacustrine deposition. Most other lithofacies (Srs, Sra, Srb, Sp, Sdg, Srg, St, Gus, Gt, Gi, zD, Table 3.1) are associated with delta, subaqueous fan and moderate to high-energy lake bottom environments (Table 3.1). Silt with sand inclusions (Zin), convoluted silt (Zc) and boulder diamicton (bD) occur in low, moderate and high-energy lake bottom environments (Table 3.1). Some lithofacies (e.g. laminated clayey-silt, Cl and clayey silt with inclusions, Cin) are very rare. Only sedimentary sections (Sections 1a, 2, 3, 10, 11, 13, 20) that illustrate major findings are presented in this Chapter. Additional sedimentary logs are presented in Appendix 3 (Sections 12, 14, 16; part details of Sections 11, 13). Additional photographs of lithofacies are presented in Appendix 4.

3.6. Sediment delivery and deposition

The variety of lithofacies identified in Table 3.1 reflects the range of processes in which sediments were delivered to, dispersed through and deposited in the late glacial ribbon lakes of the Thompson Valley. Three landforms provide the framework for the following exploration of the sedimentary environment of these lakes: deltas, subaqueous fans and low-, moderate- and high-energy lake bottom.
Figure 3.4: Lithofacies examples. (a) Dropstone (very rare) in laminated silt. (b) Convoluted silt underlying (i), loaded laminated silt (flames at top) (ii), and a thin bed of stony silty diamict (iii). (c) Stony silty diamict beds (bracket). (d) Ripple cross-laminated sand (type B and S). (e) Thick beds of diffusely graded sand. (f) Silt with sand inclusions (saucer-shaped pillows). (g) Undulatory stratified gravel (i), and overlying laminated silt (ii). (h) Angular boulder bed with silt matrix. Scale card is 8 cm wide. Stick is 1 m long with 10 cm increments. Sedimentary section number indicated (see Fig. 3.1 for location). See Table 3.1 for interpretations.
Paleoflow measurements (Fig. 3.1) and trends in lithofacies away from tributary inputs (discussed below) indicate that tributaries were the dominant source of sediment, not valley ice. Taken together, lithofacies, lithofacies associations and landform-sediment relationships record high rates of sedimentation and a dynamic and high-energy lake environment. Where tributaries delivered large quantities of sediment to the lake they formed subaqueous fans and deltas.

Most sedimentary exposures are the erosional remnants of Holocene incision by the Thompson River and so are capped by fluvial gravel. In many exposures late glacial lacustrine sediments extend below present river level. Consequently, most sections only provide a partial record of lacustrine sedimentation.

3.6.1. Deltaic sedimentation

Three paleo-deltas are identified in the study area; one is located where the Deadman River tributary meets the Thompson Valley and the other two are located on Durand Creek east of Deadman River (Fig. 3.1). Additional deltas are located east of the study area (Chapter 2, Fig. 2.1). Of all the deltas in the region the Deadman delta is the largest (~13 km²) and contains an important geomorphic and sedimentologic record of lake paleoenvironment. The paleogeographic significance of Deadman delta to late glacial lake evolution in the Thompson Valley is explored in Chapter 2 based on geomorphic and geophysical data. The paleoenvironmental significance of Deadman delta is elucidated here from sedimentological evidence.

Deadman delta formed at the Deadman-Thompson River confluence, near Savona and the outlet of present day Kamloops Lake (Fig. 3.1). It is composed of three large and several smaller delta surfaces (Fig. 2.6). As lake levels progressively lowered within the Thompson Valley during deglaciation the delta surface was concomitantly incised and lowered (Chapter 2). GPR profiles completed in close proximity to one another show large contrasts in the orientation of foreset beds (compare Appendix 2.2 and Fig. 2.8) and suggest that the delta likely prograded in lobes as distributaries switched positions and the delta was incised.

Deadman delta is a coarse-grained, steep-faced, deep-water (~50 to 100 m deep depending on lake stage), Gilbert-type delta (after Postma 1995). Topsets (Section 2, Fig. 3.1 and 3.5) are composed of imbricate cobble gravel recording the bed of abandoned braided channels on the delta surface. Delta progradation is recorded by foresets composed of inclined, alternating gravel and sand lithofacies (Fig. 3.5). Gravel lithofacies include stratified and weakly stratified imbricate gravel (Table 3.1). Sand lithofacies include planar stratified, trough cross-beds and type A ripple cross-lamination.
Figure 3.5: (a) Sedimentary Section 2 (location in Fig. 3.1) below a terrace in Deadman delta. Person (circle) for scale. Topset-foreset contact indicated by dashed line. (b) Close-up of foresets showing inclined, alternating imbricate gravel and plane-bedded sand lithofacies. The range of sediment delivery mechanisms recorded in foreset lithofacies suggests that discharge and sediment supply was highly variable (see text for discussion). Paleoflow measurements (Sections 1a, 1b, 2, 3, 4, 5, Fig. 3.1) indicate that a large proportion of lake basin sediments were derived from flows building the delta.
(Table 3.1). Some sandy lithofacies include lignite clasts. Foreset lithofacies suggest that sediments delivered to the delta edge avalanched down the delta front as grain flows, rolled down the delta front under the influence of gravity, or flowed down the delta front as river-generated underflows (hyperpycnal flows, Table 3.1). This range of sediment delivery mechanisms suggests that discharge and sediment supply was highly variable. Inflows from Deadman delta dominated the sedimentary environment of the lake in their vicinity. Paleoflow measurements (Sections 1a, 1b, 2, 3, 4, 5, Fig. 3.1) indicate that a large proportion of lake basin sediments were derived from the delta. Lignite clasts are also found in lake bottom sediments near the delta (Section 1a) indicating their Deadman delta provenance.

Like Deadman delta, the two small deltas on Durand Creek have exposures showing gravel topsets and foresets. Other deltas in the region also showed this Gilbert-type sedimentary architecture.

The anomalously large size of Deadman delta and the variability in flow regime and sediment supply implied by the range of lithofacies contained within it, suggest that remnant ice from the CIS was likely within the headwaters of Deadman River during progradation of Deadman delta into the late glacial lakes of the Thompson Valley. Deadman River is misfit within Deadman Valley and has been mapped as a major meltwater channel (Fulton 1975). Thus, the paleo-Deadman River was an outwash system that deposited a braided outwash plain and delivered large amounts of sediment to paleo-lakes in the Thompson Valley.

Paleotopographic reconstructions of the CIS from glacio-isostatically tilted lake shorelines suggest that crustal load was greater to the north and northwest (Chapter 2). Consequently, northern ice may have persisted longer than ice to the south. Deltas along the northern side of the Thompson Valley and the western side of the North Thompson Valley are larger than those on the opposite side of the valleys (Fig. 2.1). The biased distribution of large deltas supports the notion of remnant ice within tributary headwaters on the Fraser and northern Thompson Plateaus during delta progradation.

### 3.6.2. Subaqueous fan sedimentation

Where inflowing rivers contained sediment loads of sufficient concentration to produce frequent underflows (hyperpycnal flows), and where sediments did not aggrade to the lake surface, subaqueous fans evolved (e.g. Rust 1977, Cheel and Rust 1980, Bell et al. 2001, Russell and Arnott 2003 and references therein). There are nine subaqueous fans in the study area (Fig. 3.1). All are located where tributaries meet the main valley. Most occur in the higher relief southern portion of the study area where there are a
greater number of steeper tributaries than in the north. Below I describe the sedimentary environment associated with several subaqueous fans in the study area (Sections 11, 20, 21 and 22, Fig. 3.1), propose a valley-fan model of subaqueous fan formation and discuss the effects of large subaqueous fans on lake drainage.

3.6.2.1. Paleo-Bonaparte subaqueous fan

The paleo-Bonaparte subaqueous fan is exposed in an extensive river bluff (100 m high, 2 km long) ~4 km east of Ashcroft along the Thompson River (Section 11, Figs. 3.1, 3.6). Given the size and limited safe access to this section, it has been divided into five macro-sedimentary units; units 1 to 3 accumulated in late glacial ribbon lakes and units 4 and 5 are post-lake, Holocene river and alluvial fan deposits, respectively. Lithofacies associated with each lacustrine unit are listed in Figure 3.6c. Unit 1 is dominated by coarse-grained lithofacies characteristic of high-energy sedimentation in subaqueous fans (e.g. Rust 1977, Cheel and Rust 1980, Burbidge and Rust 1988, Rust 1988, Sharpe 1988, Gorrell and Shaw 1991, Bell et al. 2001, Russell and Arnott 2003; Table 3.1). Units 2 and 3 are dominated by laminated silt and sand characteristic of quiet-water sedimentation on the lake bottom. Coarse-grained lithofacies within unit 1 include planar stratified sand, trough cross-bedded sand and granules, diffusely graded sand (Fig. 3.4e), reverse graded sand, undulatory stratified gravel (Fig. 3.4g, 3.7), trough cross-bedded gravel, imbricate gravel, stony silty diamicton and boulder diamicton (Table 3.1). Undulatory stratified gravel, trough cross-bedded gravel and diffusely graded sand lithofacies suggest very high rates of deposition primarily from suspension and traction and possibly from hyperconcentrated hyperpycnal flows (e.g. Cheel 1989, Carling 1999, Russell and Arnott 2003). Planar stratified sand, trough cross-bedded sand and imbricate gravel lithofacies suggest deposition from traction transport in high-density turbidity currents (e.g. Eyles et al. 1987, Rust 1988, Sharpe 1988, Cheel 1989, Russell and Arnott 2003). Diamicton lithofacies record subaqueous debris flows (e.g. Costa 1988, Levson and Rutter 1988). Cosets (0.1 – 10 m thick) of laminated silt punctuate unit 1 (Fig. 3.6). They record deposition from suspension and may represent quiet-water sedimentation during pauses in inflow. Gravel and sand lithofacies overlying laminated silt cosets often contain large laminated silt rip-ups (up to 3 m width) and exhibit sharp, irregular or loaded (>5 m thickness: anvil shapes, stringers and mega-flames) lower contacts (Fig. 3.7), evidence of the erosive power of, and rapid sedimentation from, hyperconcentrated and high-density hyperpycnal flows. Clastic dikes cut through diffusely graded sand lithofacies record rapid loading and dewatering of deeper sediments. Collectively the lithofacies within unit 1 are consistent with descriptions of subaqueous
Heavily slumped area (detail in Chapter 4)

Location of detailed sedimentary Section 11 log (Appendix 3.1)

Lacustrine Fades (Table 3.1): Sra, Srb, zin, Sp, Sdg, Srg, St, Gus, Gt, Gi, Gos.

Alluvial fan containing tephra (likely Mazama)

Holocene terrace gravels

High energy lake bottom sediments - hyperpycnal underflows

Low energy lake bottom sediments with occasional high energy hyperpycnal underflows

Subaqueous fan sediments - coarse sand and gravel with occasional silt interbeds; underlain by boulders

Figure 3.6: (a) Photographic panorama and interpreted sedimentary sequence of Section 11 (2 km long, 110 m high; location in Fig. 3.1). Colluviated portions of section inferred. (b) Perspective hillshade DEM showing geomorphic context of sections (white dots and black arrows), paleoflow directions (red arrows) and present river direction (blue arrows). (DEM data, British Columbia Government 1996). (c) Generalized sedimentary sequence of Section 11. Lithofacies variability in a portion of unit 1 (location in (a)) illustrated in Appendix 3.1.
Figure 3.7: Close-up of a portion of subaqueous fan sediments in Sedimentary Section 11 (location in Fig. 3.6a). Large blocks of silt were ripped up and deposited within the overlying gravel bed. Note the undulatory stratification of the gravel bed (white dashed lines). Partial location of Appendix 4.3a indicated by dashed box.

fans and the sediment delivery processes associated with them elsewhere (e.g. Rust 1977, Cheel and Rust 1980, Burbidge and Rust 1988, Rust 1988, Sharpe 1988, Gorrell and Shaw 1991, Bell et al. 2001, Russell and Arnott 2003 and references therein). Faceted and rounded boulders (>2 m long, a-axis) at the bottom of the section (unit 1, Fig. 3.6) are likely equivalent to proximal subaqueous fan lithofacies (boulder gravel) described and interpreted in other studies of subaqueous fans (e.g. Rust 1988, Sharpe 1988). A jökulhlaup origin for the sediments of the paleo-Bonaparte subaqueous fan is proposed based on its large size and inferred high sedimentation rates.

The upper surface of the subaqueous fan (contact between unit 1 and 2) on the eastern portion of the section is sloped ~5° (apparent dip) down to the southeast (Fig. 3.6, Appendix 4.2a) and eventually disappears beneath river level. This slope is conformable to the sedimentary architecture of the fan. This southeastward-sloping architecture and local topography (Fig. 3.1) suggest that sediments were delivered from the northwest and arrested in a fan-shaped deposit. Paleoflows measured in the overlying
lake bottom sediments (unit 3, Fig. 3.6) also indicate that sediments were derived from the northwest. These paleoflows along with those measured at neighbouring sections collectively indicate the dominance of sediment delivered from the Bonaparte River Valley (Fig. 3.6b).

Following lake drainage, fluvial gravels were deposited (unit 4), blocks of buried ice melted causing collapse (slumping and faulting on west portion of section; detail in Chapter 4) of units 1 through 4 and creating a kettle hole that was later filled by a fine-grained alluvial fan (unit 5) containing tephra (likely Mazama Ash, J. Clague personal communication 2000). The bottom of this kettle hole is largely covered by modern colluvium that likely hides unit 4 gravel. A small unit of gravel was observed in the bottom of the kettle hole but its exposure was too small to allow confident interpretation.

3.6.2.2. Paleo-Venables subaqueous fan

The lower ~30 m of Section 20 also records a subaqueous fan (Figs. 3.1, 3.8). This subaqueous fan is easily accessed. Large fluctuations in energy and sediment supply are recorded by vertical contrasts in the grain size and sedimentary structure of lithofacies (e.g. diffusely graded sand, planar stratified sand, laminated silt and beds of imbricate gravel). Numerous erosional surfaces truncate underlying beds. The position of these sediments at a tributary mouth and fabric measurements in the lowest imbricate gravel bed (at ~10 m, Fig. 3.8) suggest sediments were derived from Venables Creek tributary (Fig. 3.1) rather than from a valley ice mass (Ryder 1970).

3.6.2.3. Summary of subaqueous fan characteristics in the Thompson Valley

All subaqueous fans in the study area are located at tributary confluences. They also lie below the elevation of Glacial Lake Deadman lake level (Chapter 2). Sediments within these landforms are typically composed of interbedded coarse (gravel and sand) and fine-grained (silt) lacustrine sediments, and where well-exposed the overall form is fan-shaped and the sedimentary architecture is sloped (~5-25°) downwards from the tributary into the main valley. Paleoflow measurements indicate sediment delivery from tributaries. Stratigraphically they lie toward the base of the lacustrine fill and thus were deposited largely during the early stages of basin infilling (the largest subaqueous fans dominate the lacustrine fill – e.g. Sections 11, 21 and 22, Fig. 3.1, Appendix 5.2; see section 3.6.2.4). Fans frequently are overlain by fine-grained lake bottom sediments, and are not overlain by deltaic topset beds. Subaqueous fans were important sites of sediment deposition and underscore the importance of tributaries in delivering sediments to the lake.
Figure 3.8: Sedimentary Section 20 log (location in Fig. 3.1). Subaqueous fan sediments (7–32 m),
low energy lake bottom sediments (32–68) and deeply loaded low energy lake bottom sediments (68–118).
Deep loading is indicated by megascale sand inclusions (dashed outline) in contorted to massive silt (68–118 m).
Overlain by Holocene fluvial gravel (118–120 m). Refer to Figure 3.9 for legend.
3.6.2.4. A valley-fan model of subaqueous fan development

Subaqueous fans are created by the transport in, and deposition from, hyperpycnal turbidity currents and hyperconcentrated flows. Study of subaqueous fans has focussed on low relief areas (e.g. Cheel and Rust 1980, Bell et al. 2001, Russell and Arnott 2003 and references therein). In these areas subaqueous fans form in contact with ice as part of conduit-fan (esker-fan) systems. Glacier melt produces high discharges and sediment loads that are maintained by confinement in an ice conduit. Rapid deposition at the submerged conduit mouth produces subaqueous fans.

Throughout the study area the preferred presence of subaqueous fans at tributary mouths and at the base of lake deposits, as well as sedimentary architecture and paleoflow evidence, indicate that sediment was derived from tributary systems early in deglaciation, not from ice occupying the Thompson Valley. This means that subaqueous fans developed from hyperpycnal flows that either (1) developed from tributary river effluent, or (2) issued from submerged ice conduits. In moderately high relief settings like the Thompson Valley, high sediment load could have been maintained in steep, confined rivers draining plateau-remnant ice and surrounded by erodable, unvegetated slopes. Such valley-fan systems are characterized by (1) discharge and sediment delivery driven by melt cycles, (2) an abundant supply of coarse-grained sediment, and (3) a basin deep enough to prevent delta formation by sediment aggradation to the lake surface.

The large paleo-Bonaparte subaqueous fan (Section 11) does not occur at the end of a steep, narrow tributary, but rather it lies at the terminus of the wide, north-south trending Bonaparte Valley (Fig. 3.1c). In this case, the discharge and sediment supply required to produce the hyperpycnal flows necessary for subaqueous fan formation may have been achieved by either (1) conduit flow in Bonaparte Valley ice that terminated in the Thompson Valley, or (2) extreme river discharge event(s) such as jökulhlaups. It is possible that, at least early on, some subaqueous fans (e.g. paleo-Bonaparte subaqueous fan) formed at the end of conduits in an ice-contact lake (see section 3.4). The Chasm, at the head of the Bonaparte Valley, ~60 km north of Ashcroft, may be evidence of an extreme discharge event(s) related to the paleo-Bonaparte subaqueous fan. The Chasm is a dry falls canyon, ~300 m high and ~8 km long, cut into basalt at the edge of the Fraser Plateau. The Chasm is v-shaped in plan view and a ~35 km long esker terminates at its apex. Similar, smaller v-shaped dry falls canyons neighbour the Chasm. The Chasm dry falls is similar to those in the Channelled Scablands of Washington State that were eroded into basalt by jökulhlaups (Baker et al. 1987).
3.6.2.5. Effects of subaqueous fans on lake drainage

A pair of subaqueous fans developed on opposite sides of the Thompson Valley from Twaal and Pimainus Creeks (21, 22, Fig. 3.1). These large landforms are >200 m thick (Appendix 5.2a) and the resulting incision of the Thompson River has left their upper surfaces ~200 m above present river level. During their formation these fans coalesced and built close to the level of Glacial Lake Deadman – Lowest stage. Lake bottom sediments are found on either side of this area of the valley for many kilometres. This subaqueous fan complex would have significantly altered the longitudinal bathymetry of the lake basin such that sedimentation processes north and south of the Twaal-Pimainus fans were separated during the later stages of Glacial Lake Deadman infilling.

During catastrophic drainage of Glacial Lake Deadman (Chapter 2), the fan complex likely had a significant effect on the hydrodynamics of drainage. The fan complex may have temporarily dammed the northern lake causing two large peaks in the flood hydrograph. These two pulses may be recorded in offshore flood deposits (Blais-Stevens et al. 2001, 2003) that may correlate to the drainage of Glacial Lake Deadman (Chapter 2). The marine core sediments show a pattern of sedimentation that may indicate two closely occurring outburst flood events (Blais-Stevens et al. 2001, 2003).

3.6.3. High- to medium-energy lake bottom sedimentation

Downflow trends in lithofacies from tributaries record the transition from deposition in deltas and subaqueous fans, to deposition in high-, medium- and low-energy lake bottom environments. Lake bottom environments are identified based on their near-horizontal sedimentary architecture and lithofacies associations. This section explores the sedimentary record of the high- to medium-energy lake bottom environment.

Given the moderately high relief and numerous tributaries in the study area the lake bottom environment is dominated by high- and medium-energy lithofacies recording a range of sediment delivery mechanisms downflow from subaqueous fans and deltas. The most common lithofacies include type S, A and B ripple cross-laminated sand and diffusely graded sand (Table 3.1). These lithofacies record sand transport and deposition by turbidity currents (Table 3.1; after Middleton and Hampton 1976, Costa 1988). In high-energy lake bottom environments, sandy lithofacies often alternate with thinner laminated, infrequently convoluted, silt lithofacies and rarely with clayey-silt lithofacies (Fig. 3.9). Silt lithofacies record quiet-water deposition from suspension (Table 3.1) and are more common in medium-energy than in high-energy lake bottom environments.
Infrequent beds of stony silty diamicton record sediment remobilization in subaqueous debris flows (Table 3.1).

The lake bottom sedimentary environments proximal to a major tributary inflow, Deadman delta, are discussed below. Additional sections recording high- to medium-energy lake bottom environments are included in Appendix 3.

3.6.3.1. Lake bottom sediments downflow from Deadman delta

Paleoflow measurements (1-5, Fig. 3.1), downflow trends in lithofacies (recording energy reduction) and clast provenance data (presence of lignite clasts from Deadman Valley (Cockfield 1948) in Sections 2 and 1a, Figs. 3.1, 3.10) suggest that hyperpycnal flows from the paleo-Deadman River deposited turbidites and grain flows south and east of the delta and at least 10 km west of the delta.

In Section 3 (Fig. 3.9), stacked turbidites dominate the proximal lake bottom sedimentary sequence and indicate rapid sedimentation. Climbing-ripple sequences are decimetres to >1 m thick (Fig. 3.9). Paleoflows are consistently westward away from Deadman delta (Figs. 3.1, 3.10). Turbidity currents may have been generated by (1) failure of the delta foresets or delta-proximal lake bottom sediments, or (2) the plunging of denser sediment-laden river water into the lake (hyperpycnal flows; Gilbert 1975, Smith and Ashley 1985). Vertical structural successions (type A, B and S ripple cross-laminated sand) are highly variable, reflecting the variable flow conditions (traction versus suspension deposition, Ashley et al. 1982) of hyperpycnal (river-derived) flows. Hyperpycnal flows are confirmed by the presence of sand lithofacies within the foresets of Deadman delta (see section 3.6.1). Turbidites alternate with laminated silt and laminated or convoluted clayey-silt. Silt lithofacies record quiet-water deposition from suspension (Table 3.1). Turbidity flow and suspension sedimentation were periodically punctuated by deposition from grain flows and slumps recorded by lithofacies of sand, convoluted silt and clayey-silt with inclusions (Table 3.1).

Sediments of Section 1a record a medium-energy lake bottom environment (1a, Fig. 3.1 and 3.10). Compared to a high-energy lake bottom environment, more quiet-water conditions are recorded by a more frequent occurrence of laminated silt and a less frequent occurrence of climbing-ripple sand. Consequently, deposition by suspension is more common than in high-energy lake bottom environments. Cosets of diffusely and reverse graded sand punctuate laminated silt cosets above 11 m in the section (Fig. 3.10). Some sand cosets contain clasts of low-density lignite and dish structures, and their lower contacts are often loaded and contain flame structures and non-directional load structures. Together, these features within the sand cosets are indicative of rapid
Figure 3.9: Sedimentary Section 3 records a high-energy lake bottom environment (location in Fig. 3.1). Stacked turbidites dominate this section and indicate rapid sedimentation. Paleoflows are consistently westward from the nearby Deadman delta. Holocene floodplain gravel at 17.7-18.5 m.
Figure 3.10: Sedimentary Section 1a records a medium-energy lake bottom environment (location in Fig. 3.1). Compared to high-energy lake bottom sedimentary environments, here deposition by suspension is more common as recorded by greater occurrence of laminated silt and lesser occurrence of ripple cross-laminated sand. Above 11 m in section, alternating beds of laminated silt and diffusely graded and reverse graded sand record rhythms in sedimentation related to river inflow and/or deltaic processes. Refer to Figure 3.9 for legend.
deposition from turbidity (possibly hyperconcentrated) and grain flows issuing from Deadman delta. Intervening, relatively quiescent periods allowed deposition of silt from suspension. The strong rhythmic pattern of these beds may be related to daily, seasonal or annual changes in sediment flux to the basin and may be sourced to changes in meltwater discharge and/or avalanching of foreset beds (possibly triggered by wave action).

The location of Section 1a (recording medium-energy lake bottom conditions) closer to Deadman delta than Section 3 (recording high-energy lake bottom conditions) requires comment. Paleoflows at Section 1a record mainly eastward flow, those at Section 3 record westward flow (Fig. 3.1). This implies that Section 1a was distal to the mainly westward flowing paleo-Deadman River inflow (Fig. 3.1). Alternatively, sediments at Section 1a were deposited at a different time from those at Section 3 and when lake conditions were quieter.

3.6.4. Low-energy lake bottom sedimentation

Low-energy lake bottom sediments are dominated by laminated silt lithofacies (Table 3.1). Laminated silt records deposition from suspension, the sediment mainly derived from overflows, interflows and underflows issuing from tributary rivers. Low-energy lacustrine lithofacies occur throughout the valley (e.g. Figs. 3.3, 3.4a, 3.4b, 3.8, 3.9, 3.10, 3.11, Appendix 3), but given the dominance of tributary inflows to the late glacial ribbon lakes of the Thompson Valley, thick sequences of undisturbed low-energy lake bottom sediments are uncommon (Fig. 3.11). Section 10 is dominated by low-energy lake bottom lithofacies (Fig. 3.11). This section is located in an area relatively sheltered from the direct effects of tributary inputs. Cosets of massive and convoluted silt with irregular-shaped sand and silt inclusions punctuate thick cosets of laminated silt and record subaqueous slumping. Steep lake bathymetry (recorded in the bed architecture that dips 5-10° towards the valley centreline) and sediment loading from infrequent turbidity currents, grain flows and rockfalls nearby (Fig. 3.4h, Appendix 4.1a) likely facilitated subaqueous slumping. Foundering of sand into underlying silts is recorded in saucer-shaped pillows (Fig. 3.4f; at 8.2 m in section, Fig. 3.11).

3.6.4.1. The rarity of clay in late glacial ribbon lakes

An inspection of all sediment section logs reveals that lithofacies dominated by clay and clayey-silt are rare in exposed lake bottom sediments from the late glacial ribbon lakes of the Thompson Valley. Some of the few finest grained lithofacies found were examined using sieves and a sedigraph. Samples were derived from very thin (<1
Figure 3.11: Sedimentary Section 10 records a low-energy lake bottom environment (location in Fig. 3.1). Thick beds of laminated silt dominate recording quite-water sedimentary conditions. Occasional subaqueous slumps produced beds of massive and convoluted silt with irregular shaped sand inclusions. These events were likely facilitated by a steep lake bathymetry (beds dip 5-10° toward valley centreline) and sediment loading from infrequent turbidity currents, grain flows and rock falls nearby (Fig. 3.4h, Appendix 4.2a). Large faults extend through the whole exposure and intersect the logged section at 8, 13.5, and 44.5 m. They likely formed as Holocene river incision removed lateral support. Refer to Figure 3.9 for legend.
cm) clayey laminae and the fine portion of clayey-silt with sand inclusions (Fig. 3.12). Consequently, it was difficult to collect samples that were not contaminated by very thin (<1 mm thick) sand partings and so these have skewed the sedigraph results toward sandy-silt. Despite this, clay is present in the samples and it likely forms a somewhat higher percentage than the grain size curves indicate (Fig. 3.12). The strong inflection point at the fine sand-silt boundary (Fig. 3.12) is a function of the lab techniques employed (i.e. dry sieving for fine sand and sedigraph analysis for silt and clay). In contrast, clay is relatively abundant in the sedimentary record of aerially expansive deglacial lakes (Antevs 1951, Teller 1976, Fenton et al. 1983, Smith and Ashley 1985, Parent and Occhietti 1999).

**Figure 3.12:** Cumulative percent grain size curves for selected fine-grained samples. Samples likely contained some sand partings that have skewed the sedigraph results toward sandy-silt 1) grey laminated clayey-silt with sand partings at 5.3 m in Section 3 (Fig. 3.9); 2) grey clayey-silt with sand inclusions (fine portion) at 1.75 m in Section 3; 3) grey laminated clayey-silt with sand partings at 2.3 m in Section 3; 4) reddish clayey-silt at 4.25 m in Section 12 (Appendix 3.2).

Bedrock lithologies around Ashcroft indicate that clay would have been delivered to the lakes. So why was it not deposited? The lack of clay can be attributed to (1) sediment stratification within the lakes preventing settling and lake overturning (Smith and Ashley 1985), (2) lake currents (encouraged by the narrow, ribbon-like geometry of the lakes, numerous tributary inputs and katabatic winds) preventing settling, (3) a deep
basin requiring too long of a time period for the settling of clay (i.e. longer than the winter period), (4) clay flushing during high discharge events (e.g. jökulhlaups), (5) a climate warm enough to prevent lake freezing in winter, or (6) a combination of these factors.

Of all twenty-four sections examined in the Thompson Valley, Section 3 (Fig. 3.9) contains the most clayey-silt, in laminated and convoluted clayey-silt cosets recording both settling from suspension and lake bottom instability. Laminated clayey-silt cosets are generally ~0.5-15 cm thick. The deposition of clay requires quiet-water conditions. However, Section 3 mainly records high-energy conditions dominated by hyperpycnal turbidity currents off Deadman delta (see section 3.6.3.1). At this site, clay may have been deposited on the lake bottom either (1) when inflows temporarily switched away from the site (i.e. delivered effluent toward Kamloops Lake), or (2) when turbidity currents drew down clay through the sediment stratified water column. The latter mechanism may explain the common association of clayey-silt laminae overlying ripple cross-laminated sand (Fig. 3.9). Convoluted clayey-silt with sand inclusions record periodic slumping of clayey-silt perhaps from quieter, shallow inshore areas.

3.6.5. Large-scale subaqueous deformation

Soft-sediment deformation processes were common in the ribbon lakes of the Thompson Valley. Many of these processes are recorded by millimetre to centimetre-scale flames, convolutions and ball-and-pillow structures. In many cases, engulfed/loaded clasts are elongate, with their long-axis parallel to the bed (Fig. 3.4f), and the cosets they occur in are <6 m thick (Appendix 3.4). However, very thick cosets (10's m) of massive to highly contorted laminated silt with megascale (up to 10 m long axis) sand inclusions are also common (Table 3.1; Fig. 3.8). To my knowledge such megascale ball-and-pillow structures have not been reported in any study of glacial lakes and are rarely described from the rock record (e.g. Howard and Lohrengel 1969). Independent of scale, ball-and-pillow structures are created by foundering, in place, of coarser sediments into underlying finer sediments (Howard and Lohrengel 1969, Allen 1982 and references therein). Subaqueous slumping or sliding may compliment this process (Howard and Lohrengel 1969, Hubert et al. 1972). The higher the rate of deposition of either the fine-grained or coarse-grained units, the more the system is unstable. This condition was frequently met in the late glacial ribbon lakes of the Thompson Valley as hyperpycnal turbidity currents were common and rates of sedimentation were high (see section 3.7.4). At Section 20 a 50 m thick unit (at 68-118 m, Fig. 3.8) of contorted laminated to massive silt contains very large sand inclusions (ball-and-pillow structures). Curvilinear
shear planes (<1 cm thick) are also observed (at ~70 m, Fig. 3.8). The presence of convolutions, ball-and-pillow structures and shear planes suggest that this unit was subject to both ductile and brittle deformation. Loading was accompanied by slumping. Slumping is inferred from shear planes and the asymmetric shapes of sand inclusions, implying a lateral flow component (Fig. 3.8).

Large-scale subaqueous failure events may have been triggered by (1) sediment loading, (2) changing lake levels (including lake drainage), (3) lake currents, or (4) seismic events. Presently, earthquakes are relatively infrequent in the interior compared to the coast. Earthquakes were likely significantly more frequent during deglaciation due to hundreds of metres of glacio-isostatic rebound (Chapter 2) and fault instability (Lagerbäck and Witschard 1983, Johnston 1989, Muir-Wood 1989, Wu and Hasegawa 1996, Stewart et al. 2000).

3.7. Summary of the sedimentary environment of ribbon lakes

3.7.1. Model of the sedimentary environment

A conceptual model of the sedimentary environment of ribbon lakes in the Thompson Valley is presented in Figure 3.13. Paleoflow measurements indicate that the majority of sediments were delivered from tributary outwash streams rather than ice within the Thompson Valley. Remnant ice masses within the headwaters of some tributaries delivered large amounts of meltwater and sediment to ribbon lakes.

3.7.2. Sedimentation processes

Sediments were dispersed by turbidity currents (hyperpycnal and slump-generated), hyperconcentrated hyperpycnal flows, grain flows, debris flows, overflows and interflows and were deposited from suspension and traction. Sediment gravity flows frequently eroded and loaded underlying saturated sediments to produce a variety of ductile deformation structures (e.g. flames, convolutions, dish structures, ball-and-pillows, etc.) and brittle deformation structures (e.g. synsedimentary faults). Large-scale subaqueous loading and failure produced megascale sand inclusions (see section 3.6.5). Rockfalls triggered lake bottom instabilities. Finally, the melting of buried ice resulted in sediment collapse (faults and folds) and the creation of kettle holes (Fig. 3.6, Chapter 4).

3.7.3. General spatial and temporal patterns of sedimentation

A strong spatial pattern of sedimentation is related to sediment delivery and dispersal from tributary inflows. Considerations of lithofacies, their associations and
Tributary outwash stream with high suspended sediment load

Buried ice

Low energy lake bottom sediments

Rockfall sediments

Subaqueous fan sediments

Bedrock and pre-glacial fill capped by till

Subaqueous slumping

Suspension settling

Hyperpycnal turbidity currents and hyper-concentrated flows

Overflow interflow

Delta prograding from large tributary river

Tributary ice mass

Turbidity current sediments concentrated in bathymetric lows

Sediment stratified water column

Figure 3.13: Conceptual model of the sedimentary environment in and around deglacial ribbon lakes in the Ashcroft area. Ice dam not shown.
architecture, and their geomorphic context allow the classification of lake sediments into lacustrine landforms: deltas, subaqueous fans and lake bottom. High-energy sand and gravel deposits in deltas and subaqueous fans give way to sandy turbidites in high- and medium-energy lake bottom environments and finally to laminated silt in more distal, lower-energy lake bottom environments. Boulder diamicton, convoluted silt and silt with sand inclusions are not strongly spatially biased with respect to tributaries.

Temporal patterns in sedimentation reflect (1) hourly to seasonal to episodic pulses in meltwater discharge and sediment supply, and (2) an overall reduction in meltwater discharge and sediment supply to the lakes over deglaciation and prior to drainage. Pulsed sedimentation is recorded in laminated silt and stacked turbidites. Classic varves are not present and consequently varve chronology cannot be used to constrain lake duration or inflow periodicity. Lake bottom turbidites become less dominant and laminated silt predominates upsection (e.g. Fig. 3.8).

3.7.4. Rate of sedimentation

High rates of sedimentation are indicated by (1) the presence of large deltas and subaqueous fans at tributary mouths, (2) extensive, thick, lake bottom turbidites recording pulsed rhythms of sedimentation, (3) thick cosets of laminated silt, (4) mega-scale subaqueous deformation, and (5) a rarity of clay. Rapid basin infilling was accomplished by (1) energetic meltwater and sediment delivery via tributaries associated with rapid deglaciation (rapid melting of plateau-remnant ice and abundant sediment supply from steep, unvegetated, unstable slopes; Clague 1981, Porter and Swanson 1998, Clague and James 2002) and (2) sediment focussing from a regional catchment to a relatively small basin (Fig. 3.1c). It is unlikely that large snowpacks could have produced these high rates of sedimentation.

3.7.5. Duration of ribbon lakes

The duration of late glacial ribbon lakes can only be estimated as varve chronology cannot be used. Based on lithofacies analysis, sedimentation rates were likely high. This suggests that the lakes may have existed for a relatively short time. Based on very limited radiocarbon data, the lifespan of all glacial lakes in the Thompson Valley is ~850 14C years (Chapter 2). A short lifespan for lakes in the Thompson Valley agrees with the inference that these lakes were ice-dammed (Chapter 2). East of the study area, in the South Thompson Valley (Fig. 1c), lake bottom sediments interpreted as varved also support the notion of high rates of sedimentation during deglaciation (Fulton 1965). There, single exposures are reported to record up to 80 varves, which suggests it only
took only 80 years to deposit tens of metres of sediment (Fulton 2000). Consequently, the duration of late glacial ribbon lakes in the Thompson Valley was likely a few hundred years, but possibly much shorter.

3.8. Implications for Cordilleran deglaciation in the southern Interior Plateau

The prevailing model of Cordilleran deglaciation calls for regional downwasting of the CIS with the plateau uplands becoming ice-free before the valleys. This model is based on (1) a lack of evidence demonstrating recession of the CIS to the mountains (i.e. pervasive recessional moraines), (2) reconstruction of the horizontal and altitudinal pattern of ice retreat in four small study areas from meltwater channels and ice-marginal landforms, and (3) conceptual logic (Fulton 1967, 1991). It is tested here.

3.8.1. Regional stagnation

The stratigraphic record in the Thompson Valley generally supports regional stagnation rather than active retreat of the CIS. If ice within the valleys actively retreated, fluctuations of the ice margin would have occurred. In the ice-contact lake environment, this process may have caused over-riding of previously deposited lake sediments producing interdigitation of thrust lake sediments and till. Rather, the stratigraphic record shows an abrupt transition from till or older Quaternary sediments to deglacial lake sediments (Fig. 3.2, 3.3). If deglaciation occurred by active calving, more of an abundance of dropstones would be expected. Dropstones are rare within lake bottom sediments. The rarity of dropstones may also indicate that (1) valley ice was too thick compared to the depth of lakes, (2) valley ice contained enough debris to prevent flotation (Chapter 4), or (3) only a small length of the lake was in contact with ice.

3.8.2. Ice-free valleys and plateau-remnant ice

A high rate of sediment delivery, including possible jökulhlaup event(s), to late glacial ribbon lakes from tributaries with their headwaters on the Plateau favours the presence of ice remnant on the Plateau (see section 3.6.1). This indicates that portions of the valley were ice-free and containing lakes while portions of the plateau were not ice-free. This pattern of deglaciation has also been inferred elsewhere in the southern interior of British Columbia (Lesemann and Brennand 2003) and central interior (Eyles et al. 1987).

If at the start of deglaciation the valleys (over 1 km deep) were filled with ice then it seems unlikely that the much thinner ice on the plateau could have persisted long
enough to exist by the time lakes developed in the valleys. In addition it also seems unlikely that plateau ice could have become fully active, and thus persisted, as the accumulation areas for the ice sheet (i.e. mountainous areas) deglaciated first (Fulton 1991, Clague 1989). An absence of recessional moraines on the plateau also argues against active decay of the CIS (Fulton 1967, 1991). Still, some process must account for the inferred pattern of deglaciation.

Deglaciation of the Thompson Valley prior to the plateau may have occurred by preferential thinning of the CIS over the valley prior to and during downwasting. Thinning of valley ice prior to deglaciation may have been accomplished by fast ice flow in the valley and meltwater routing though the valley. Drumlins are thought to record fast ice flow (Boulton 1987, Shaw 1994, 1996, Jorgensen and Piotrowski 2003). Drumlins occur within the valley near Ashcroft. Their orientations conform to the valley orientation (Fig. 3.1) yet contrast with the dominant orientation of drumlins on the surrounding plateau (to SE, fig. 13, Ryder 1976), lending support to the hypothesis that valley ice may have been thin prior to lake development. This hypothesis requires further testing.

3.9. Conclusions

Study of late glacial lakes in Canada has been dominated by investigations of areally extensive proglacial lakes in relatively low relief regions associated with the Laurentide Ice Sheet. This paper broadens and deepens our knowledge of late glacial lakes in Canada by focusing on ribbon lakes in moderately high relief areas of the Cordilleran Ice Sheet.

The sedimentary environment of ribbon lakes was unique in many respects. Although technically ice-contact lakes (they were ice dammed and remnant ice blocks were buried by them), ribbon lakes received most of their water and sediment supply from rivers (Ward and Rutter 2000; Lesemann and Brennand 2003) with their headwaters on the plateau (Fig. 3.13). Early water and sediment delivery may have been through ice conduits. Sedimentation rates were very high and resulted in the formation of numerous deltas, subaqueous fans and extensive high-energy lake bottom sediments (recording turbidity currents, grain flows and debris flows). I propose a valley-fan model for subaqueous fan formation. The high discharges and sediment loads required to form hyperpycnal flows suggest that ice likely remained on the plateau while the Thompson Valley was mostly ice-free and occupied by deglacial lakes. This conclusion suggests that ice in valleys thinned prior to the on-set of deglaciation. Away from inflows thick cosets of low-energy lake sediments (laminated silts; "white silts" Dawson (1879)) settled out of the turbid water column. In contrast to areally extensive deglacial lakes, clay and
dropstones are rare and classic varves were not identified (q.v. Antev 1951, Teller 1976). The absence of classic varves and the rarity of clay are attributed to hindered settling of clay due to sediment stratification in the water column and energetic lake conditions throughout the year. Dropstones were rare likely as stagnant decay of the CIS discouraged calving.
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Buried ice
Chapter 4

Millennia-scale ice preservation and kettle development following deglaciation in the Canadian Cordilleran Interior

4.1. Introduction

During the decay of the Cordilleran Ice Sheet (CIS) a number of deglacial lakes developed within the deep valleys of the Interior Plateau of British Columbia (Fulton 1969, Shaw and Archer 1979, Clague 1987, Sawicki and Smith 1992). Lake sediments within these basins frequently buried ice. As a result, kettled, tilted and faulted lake sediments are commonly observed in many valleys of the Cordilleran Interior (e.g. Eyles et al. 1987, Fulton 1965, Ward and Rutter 2000, Chapter 3). Excellent exposures of such ice-collapse sediments are found in the Thompson Valley and are presented in this paper. This short paper summarizes (1) the landform-sediment relationships associated with buried ice blocks and the size of buried ice blocks, (2) the relative timing of kettle hole formation and (3) the process of ice burial in late glacial ribbon lakes in the Thompson Valley, British Columbia (Figs. 4.1, 4.2). These results may facilitate interpretation of disturbed lacustrine sediments observed in exposures and geophysical surveys elsewhere (e.g. Larocque 1985, Larocque and Shilts 1986, Kaszycki 1987, Eyles et al. 2003). These findings are part of a larger research project undertaken to explore the paleogeography and sedimentary environment of late glacial ribbon lakes in the Interior Plateau (Chapters 2, 3).

Figure 4.1: Study area located within the Canadian Cordillera.

4.2. Landform-sediment relationships and the size of buried ice blocks

Many exposed and buried kettle holes are found in the Thompson Valley. They are ovate, closed depressions that range in diameter from tens to hundreds of meters and are up to 40 m deep (K, Fig. 4.3). A concentric, step-like pattern of ridges flank and
ascend up to 700 m horizontal distance from the centre of some of the larger kettle holes (R, Fig. 4.3). The morphology and orientation of these ridges, with respect to the kettle holes, suggest that these features were produced by the retrogressive slumping of sediments toward the centre of the hole along concave-up, curvi-planar normal faults.

**Figure 4.2:** Hillshade digital elevation model of the study area showing kettle hole locations (Fig. 4.3 - 4.5) discussed. (DEM data, BC Government 1996)

Post-glacial incision by the Thompson River has cut through portions of kettle holes creating sedimentary exposures in them (Figs. 4.4, 4.5). Some exposures are over 100 m high (Fig. 4.4). Down-warping, and normal faults with displacement down toward the centre of the kettle hole, are common features and are indicative of sediment collapse following melting of buried ice (Fig. 4.4, 4.5).

As the volume of the void in the kettle hole approximates the volume of the original ice block, some of the ice blocks that formed these kettles holes were up to ~2 x 10^6 m^3 in volume. Similarly as the depth of the kettle hole approximates the thickness of the original ice block, some ice blocks were up to 40 m thick.

**4.3. Relative timing of kettle hole formation**

Geomorphic and sedimentary evidence suggests that many ice blocks were buried at significant depths (>100 m) within the valley fill, mainly in the early stages of deglacial lake basin infilling. However, the melting of buried ice, in many cases, did not occur until the latest stages of fluvial incision, well after the disappearance of the lakes from the valley in which the ice was buried. This inference is based upon relationships between kettle holes and river terraces, and is best exemplified at the Slough Section (Fig. 4.4; Section 11, Chapter 3) and the Battle Creek Section (Fig. 4.5; Section 8, Chapter 3).
Figure 4.3: Stereogram showing geomorphology of kettle holes (K) and kettle slump-ridges (R). In this area the melting of ice blocks resulted in the collapse of glaciolacustrine sediments (deltaic and lake bottom). See Figure 4.2 for location. (Aerial photographs BCC210: 96-98, 1979, copyright British Columbia Government, by permission)
Figure 4.4: Buried kettle hole at the Slough Section (Section 11, Fig. 3.1). Section is ~600 m long. Numerous normal faults in late glacial lake sediments (unit 1) and overlying Holocene river terrace gravel (unit 2) formed by the melting of a large buried ice block (~60 m tall and 150 m wide). Alluvial fan sediments (unit 3) containing Mazama ash (white line; 6730 \(^{14}C\) yr BP, Hallett et al. 1997) filled the kettle hole. Colluviated portions of the section inferred. See Figure 4.2 for location.
Figure 4.5: (a) Perspective view hillshade digital elevation model of the Battle Creek kettle holes (K1, K2) and surrounding river terraces (T1 - T4). DEM is 3 x 3 km, 3 x vertical exaggeration. The elevation of present river level is ~315 m asl and the elevation of fluvial terraces are ~415 m asl (T1), ~410 m asl (T2), ~365 m asl (T3), and ~350 m asl (T4). (b) Cutbank exposure along the rim of a kettle hole (K1) and (c) Inferred stratigraphy (colluviated portions of the section inferred). See Figure 4.2 for location. (DEM data, BC Government 1996)
At the Slough Section (Fig. 4.4), downwarped and normally-faulted glaciolacustrine silt and river terrace gravel record a buried kettle hole. This kettle hole was filled with alluvial fan sediments containing Mazama ash (J. Clague personal communication 2000; 6730 $^{14}$C aBP, Hallett et al. 1997) within ~10 m of the top of the exposure. Holocene river terrace gravel (unit 2) lies above the elevation of the kettle hole. If the kettle hole existed prior to river incision then the hole would have been filled with river terrace gravel. Instead, river terrace gravel blankets the edges of the hole and is downwarped into the hole indicating that the ice block did not melt until the later stages of fluvial incision, but prior to the start of alluvial fan infill (alluvial fan sediments are not downwarped).

At the Battle Creek Section, kettle holes (K1, K2, Fig. 4.5) retain their surface expression and lie closer to river level than the one at the Slough Section. The rim of K1 is at ~350 m asl (T4, Fig. 4.5), ~35 m above present river level and well below the highest river terrace locally at ~415 m asl (T1, Fig. 4.5). Both kettles holes have well-preserved rims. K2 has been partially filled by alluvium, producing a flat bottom at an elevation of ~325 m asl and ~10 m above present river level. The river has eroded a cutbank exposure along the rim of K1. The exposure reveals Holocene river terrace gravel overlying late glacial (Fraser Glaciation) lake silt and sand. Evidence for the collapse of sediment into a space previously occupied by buried ice includes (1) a semi-circular land surface depression, (2) normally-faulted sediment downthrown under the surface depression, and (3) sedimentary architecture (i.e. bedding planes) in late glacial lake sediments tilted toward the kettle hole (Fig. 4.5; note that north (right) of the photograph the bedding planes become level). The faulting and downwarping of relatively young Holocene river gravel and the land surface expression of the kettle holes indicates that the buried ice melted during the later stages of river incision – after the river had incised below the level of the kettle holes. If the ice block melted and sediment collapsed prior to incision below the kettle rims, the kettle holes would have been buried by alluvium and their land surface expression lost.

Absolute ages are not yet available for these events; bracketing ages can be inferred. Late glacial lakes of the Thompson Valley drained by 8,900 ±150 $^{14}$C yr BP (GSC-193, Dyck et al. 1965; Fulton 1969; Chapter 2). It is possible that the last lake drained between ~9,200 $^{14}$C yr BP and 10,800 $^{14}$C yr BP (Chapter 2). Incision to within a few metres of present river level was achieved by the mid-Holocene, 6730 $^{14}$C yr BP (Ryder 1981, Stryd 1987, Hallett et al. 1997). Thus, buried ice blocks in the Thompson Valley persisted for millennia - through late glacial lake basin infilling (~150 m thick lake sediment package), lake drainage and much of the Holocene river incision (~75 m).
persistance of these ice blocks is explained by their burial and insulation under at least 100 metres of late glacial lake sediment. The melting of ice blocks was triggered by the removal of this insulating sediment during river incision in the early to mid-Holocene.

4.4. Process of ice burial

These ice blocks, along with many others identified in the Cordilleran Interior, were buried in a lake environment. As ice has a lower density than water, how does ice manage to be buried at/in the bottom of a lake when it is expected to float?

During deglaciation of the region, stagnant ice masses were trapped in valleys causing the damming of lakes. Lakes developed in contact with stagnant ice and so ice was largely drowned prior to the burial of valleys by lake bottom sediments (Chapter 3). Many of the ice blocks were buried during the earliest stages of basin infilling. As lakes were deep (>100 m, Chapter 3), clean ice blocks >100 m thick could have floated. However, dropstones are rarely observed in the lake bottom sediments and suggest that icebergs were uncommon (Chapter 3). The rarity of dropstones combined with the fact that ice blocks were buried when the lake was very deep suggest that the ice blocks did not float. Ice flotation may have been prevented by: (1) freezing of ice to its bed, (2) anchoring of ice by lateral or overlying accumulations of sediment, and/or (3) debris loading within ice such that it had a bulk density greater than that of the lake water. The first hypothesis is unlikely as the CIS was warm-based in the valleys (Lian and Hicock 2000), especially during deglaciation when remnant valley ice was in contact with lakes. The second hypothesis is possible, as sediment delivered by tributary outwash streams to trunk valleys may have partially buried valley ice (Chapter 3). Sediments derived from topographically higher valley ice or valley sides may have also helped pin ice blocks (e.g. Kaszycki 1987, fig. 16). The third hypothesis is also possible as it can take as little as 4% volume of sediment to cause an ice block to resist flotation, depending on ice and sediment density (Fig. 4.6).

4.5. Implications of millennia-scale ice preservation for mapping

Delayed melting of buried ice blocks has implications for terrain and geological mapping. Terrain and geological maps generally classify gravely kettled terrain as glaciofluvial sediments formed during deglaciation (e.g. Howes and Kenk 1997). Such mapping of surficial sediment relies heavily on interpretations from aerial photographs. In the example of the Battle Creek kettles, stratigraphic and geomorphic relationships demonstrate that the gravely and kettled sediment is Holocene alluvium that became
kettled in the early- to mid-Holocene when ice buried by underlying late glacial lake sediments melted. Thus, the Battle Creek Section demonstrates that kettled surfaces may not indicate a glacial origin for surficial sediments. Considerations of millennia-scale ice preservation and observations of kettle hole stratigraphy in other studies may help in the accurate mapping of surficial sediments and in reconstructing landscape evolution.

Figure 4.6: Range of the minimum percent volume of sediment and rock needed to sink an ice block calculated for a range of ice densities (830-917 kg m\(^{-3}\)), a range of sediment and rock densities (1000 to 3000 kg m\(^{-3}\)), and a lake water density of 1000 kg m\(^{-3}\) (similar to Tweed 2000).

4.6. Conclusions

This is the first study in the Cordillera to estimate the duration of preservation of buried ice produced from the decay of the Cordilleran Ice Sheet. Buried ice persisted from the early stages of late glacial lake basin infilling to the latest stages of fluvial incision in the early- to mid-Holocene, well after the lakes drained. Future research may better delimit the duration of ice preservation by dating low river terraces using tephra chronology, radiocarbon dating of archaeological remains, or optical dating of post-glacial loess on their surface (O.B. Lian personal communication 2002). As the development of kettle holes in the Thompson Valley was tied to regional deglacial processes, it is expected that other valleys in the Cordilleran Interior experienced similar events and may show evidence for prolonged ice preservation as well.
Our results suggest that caution is required when mapping gravely kettled terrain from aerial photographs in the Cordillera and similar terrains – gravely kettled terrain may record a deglacial glaciofluvial deposit or a Holocene river terrace.


Chapter 5

Conclusions

At the outset of this study a number of research questions were posed to fill gaps in our understanding of late glacial lake paleogeography, evolution and paleoenvironment in the Thompson Basin. Answers to these questions are provided below and arranged to address each of the two research objectives.

Objective: 1: To determine lake paleogeography

1. What was the geometry (i.e. areal extent, elevation, depth and isostatic tilt) of Glacial Lake Thompson and Glacial Lake Deadman?

Two well-defined lake stages were identified: Glacial Lake Thompson – High stage and Glacial Lake Deadman – Lowest stage. Both lakes and their stages were remarkably narrow and long (width to length ratios of ~3:100). This geometry was dominantly controlled by the physiography; namely, the moderately high relief of the Thompson Valleys incised into the Interior Plateau. Glacial Lake Thompson – High stage was ~220 km long, ~140 m deep and ~83 km³ in volume. Isostatic tilt has caused the paleo water plane to have an upslope direction of tilt to 332° ±7.7° with a slope of 1.8 ±0.6 m km⁻¹. Glacial Lake Deadman – Lowest stage was ~160 km long, ~50 m deep and ~24 km³ in volume. Isostatic tilt has caused the paleo water plane to have an upslope direction of tilt to 321° ±3.8° with a slope of 1.7 ±0.2 m km⁻¹. These isostatic tilts are among the highest measured in the world and are likely related to the unusual properties of the Cordilleran lithosphere and mantle (thin crust and low viscosity mantle with high heat flow). The orientation of their tilt supports the notion of greatest crustal load and consequentially greatest rebound, to the north and northwest.

2. How did the lake geometries change through time?

These ribbon lakes developed in response to topography and deglacial style. As valley ice masses passively melted, lakes lengthened, became shallower and narrower. Prior to Glacial Lake Thompson – High stage, higher although poorly defined lakes developed within the Thompson Basin. Glacial Lake Thompson – Highest stage developed
in the South Thompson Valley. Continued ice recession led to the development of the well-defined Glacial Lake Thompson – High stage that occupied the North Thompson, South Thompson and Thompson valleys with its outlet in the Shuswap Basin. Continued lake lowering eventually exposed high elevation lake bottom sediments in the South Thompson Valley and caused the outlet for ponded water in the North Thompson and Thompson Valleys to shift to near Kamloops. This shift in outlet location marked the beginning of Glacial Lake Deadman. The lowest stage of this lake (Glacial Lake Deadman – Lowest stage) occurred prior to final lake drainage. After this drainage higher stages of present day Kamloops Lake developed, dammed behind raised delta sediments at the modern outlet of Kamloops Lake.

3. **How and where were the lakes dammed?**

   Glacial Lake Thompson – Highest stage was dammed by ice at either end of the South Thompson Valley. Glacial Lake Thompson – High stage and Glacial Lake Deadman – Lowest stage were dammed by ice in the lower Thompson Valley, south of Spences Bridge. The lower Thompson Valley is the narrowest, highest relief portion of the Thompson Basin. These conditions likely aided in the delayed melting of valley ice. If the ice dam was buried by sediment (e.g. colluvium), it may have been able to persist for hundreds of years (Chapter 4). In support of an inferred ice dam in this area there is also a complete absence of thick deposits of lake bottom sediments, unlike the continuous 75 km swath of cliffs of lake bottom sediments in the valley immediately to the north.

4. **What was the style of final lake drainage?**

   Glacial Lake Deadman – Lowest stage drained catastrophically. Drainage bedforms within the basin, near the present outlet for Kamloops Lake, and an inferred ice dam support this interpretation. ~20 km$^3$ of floodwaters travelled ~250 km into the Fraser River system and terminated in the Georgia Basin, depositing exotic mud sometime between ~9,200 $^{14}$C yr BP and 10,800 $^{14}$C yr BP.
Objective 2: To investigate the paleoenvironmental controls on sedimentation

5. What were the styles of sediment deposition?

Seventeen glaciolacustrine lithofacies were identified ranging in grain size from clayey-silt to boulder. Sediments were deposited from suspension and traction. In a few instances deposition by falling clasts occurred (i.e. rockfall, dropstones). Deposition by suspension settling was more common for finer-grained lithofacies (e.g. laminated silt).

6. What were the sediment transport dispersal mechanisms?

Sediments were dispersed in the lakes by overflow (hypopycnal flow), interflow and underflow (low to high density hyperpycnal turbidity flows and slump-generated flows). These processes also included debris flow, subaqueous avalanche, subaqueous rockfall, and sediment remobilization by slumping and soft sediment loading.

7. What were the sources of water and sediment?

Paleoflows and progressive changes in lithofacies from tributary inputs indicate that the many tributaries to these lakes were by far the dominant contributors of sediment. Thus, valley ice played a very minor role in contributing sediment to the basin. Ice remnant on the plateau and within the headwaters of some tributaries best explains the sedimentary record and rapid basin infilling.

8. Are there spatial patterns in lake sedimentation?

At tributary mouths, large amounts of sediment accumulated to form large deltas and subaqueous fans. Sediments that travelled further into the basin were deposited in high-, medium- and low-energy lake bottom environments. Lake bottom sediments record increasing sedimentation by suspension (e.g. laminated silt) and decreasing sedimentation by turbidity currents (and other sediment gravity flows; e.g. ripple cross-laminated sand) further from tributary mouths. Underflows travelled along and filled bathymetric lows while overflows and interflows had a more diffuse pattern of dispersal and deposition. Large subaqueous failures were common and occurred in various portions of the basin with no strong spatial relationship to tributaries. Fine-grained lacustrine sediment (clay and clayey-silt) was rarely deposited (i.e. no classic varves). The deposition of clay was likely hindered by energetic lake conditions and a sediment stratified water column.
9. Are there temporal patterns in lake sedimentation?

Temporal changes in lake sedimentation are recorded by vertical changes in sediments. There is a general fining in sediments vertically (i.e. over time). Lake bottom turbidites become less dominant and laminated silt predominates upsection. These patterns are related to an overall reduction in meltwater discharge and sediment supply to the lakes over deglaciation and prior to drainage. Subaqueous fans are in the lowest position within the basin fill. Classic varves are not present and consequently varve chronology cannot be used to constrain lake duration or pulse periodicity.

Rates of deposition varied from relatively low rates producing laminated silt and clayey-silt lithofacies to extremely high rates producing thick beds (up to tens of metres) of coarse-grained (sand to gravel) lithofacies, some possibly of jökulhlaup origin.

10. What was the paleohydrologic regime of the lake?

Pulsed sedimentation reflecting hourly to seasonal to episodic pulses in meltwater discharge and sediment supply is recorded in laminated silt and stacked turbidites. This regime is related to the melting of remnant ice on the plateau, variations in weather and changing climate.

11. Can tributary sedimentation patterns be divorced from ice-front influxes?

Paleoflow data suggest that most sediment was derived from tributary outwash streams. Discontinuous lenses (<3 m thick) of gravel and diamicton were likely derived directly from a valley ice mass. The majority of lake basin infilling occurred after most of the valley ice (except for the ice dam and buried ice) had melted.

12. Do conclusions corroborate and/or extend our understanding of glacial history around Ashcroft?

Stratigraphic and sedimentologic relationships indicate that ice decay was likely by downwasting rather than active retreat (Fulton 1967, 1991). However, during the time of these lakes not all portions of the valley were ice-free, as the lakes were ice dammed. In addition, not all portions of the plateau were ice-free at this time as remnant ice in the headwaters of some tributaries, especially from the north and northwest plateau headwaters, was delivering water and sediment to the lakes.
Future work

During the course of this project a number of potential future research projects were identified.

- Further refine ideas on the causes of high glacio-isostatic tilt.
- Estimate the paleohydraulics of Glacial Lake Deadman catastrophic drainage.
- Reassess paleogeographic reconstructions completed by Fulton and Walcott (1975) for the Merritt Basin. Gain a better understanding of relative lake chronologies and the geometry of glacio-isostatic rebound.
- Investigate the effects of catastrophic lake drainage downflow of Spences Bridge to the Strait of Georgia.
- Assess the relative chronology and significance of catastrophic lake drainage from both Glacial Lake Fraser and Glacial Lake Deadman.
- Examine the record of meltwater events in the headwaters of the Bonaparte River, along the southern portion of the Fraser Plateau.
- Evaluate the hypothesis of fast ice flow in the Thompson Valley. Examine and contrast the genesis of plateau and valley drumlins.
- Investigate relationships between plateaus and valleys. Better understand the pattern and style of deglaciation, and meltwater and sediment contributions to lakes.
- Employ radiocarbon and optical stimulated luminescence dating techniques to better constrain lake chronologies, deglaciation and the duration of buried ice block preservation.
Appendix 1: Vertical aerial photographs used in this study
## Appendix 1.1: Vertical aerial photographs used in this study

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1 Not all photographs used are listed.
Appendix 2: GPR profiles not shown in thesis chapters and stereogram showing GPR survey locations on eastern portion of Deadman delta
Appendix 2.1: Stereogram of the east portion of the Deadman delta between the Thompson River, Deadman River and Hwy 1. Retrogressive failure of sediments into the void previously occupied by a buried ice block (volume \(\sim 2 \times 10^6 \text{ m}^3\), Chapter 4) created a series of kettle slump ridges. This same pattern is found for other large kettles on the delta surface. Lake drainage bedforms located around G2 are described in Chapter 2 (Fig. 2.14). The troughs of the bedforms have been cross-cut by kettle slump ridges. An absence of delta topsets from GPR profiles (G2, G3) implies that this portion of the delta is an erosional surface. The bedforms and erosional surface were formed by westward catastrophic drainage of Glacial Lake Deadman - Lowest stage). Also shown are locations of sedimentary Sections 1a, 1b (Fig. 3.1). (Aerial photo BCC210: 96-98, 1979, copyright British Columbia Government, by permission)
Appendix 2.2: (a) 50 MHz and (b) corresponding 100 MHz GPR profiles of kettle slump ridges on Deadman delta (survey location G3, Appendix 2.1). The resolution of the 50 MHz survey and the penetration of the 100 MHz survey were too poor to allow confident identification of faults. Reflections dipping to the left are delta foresets.
Appendix 2.3: GPR surveys. (a) G4, survey completed down the terrace face of the upper delta level of Deadman delta (Fig. 2.6b). Dipping reflections are delta foresets (apparent dip ~25° for reflection indicated by arrow). (b) Survey located next to the start of line G1 (Fig. 2.6b, 7) on Deadman delta. Horizontal reflections are delta topsets and dipping reflections are delta foresets. (c) G7, located on ridge-line part way down Brassy terrace riser. Dipping reflections are likely correlative with GPR radar facies 2 in lines G5 and G6 (Fig. 2.8). (d) G8, located on terrace of the paleo-Bonaparte River system (Fig. 2.9). (e) G9, located on terrace of the paleo-Bonaparte River system (Fig. 2.9).
Appendix 3: Detailed sedimentary logs not shown in thesis chapters, list of sedimentary section locations and general observations for sedimentary sections described in limited detail
Appendix 3.1: Portion of sedimentary Section 11 log (location in Fig. 3.6a) between ~370 and 390 m asl. This small portion (20 m) of the overall section (~100 m) records abrupt changes in the energy of sediment delivery: undulating stratified gravel (3.7-7 m; in-phase waves) associated with plane-bedded sand (upper flow regime), laminated silt (suspension), stony silty diamicton (debris flow) and convoluted silt (subaqueous failure). Refer to Figure 3.9 for legend.
Appendix 3.2: Sedimentary Section 12 log (location in Fig. 3.1). Proximal to a tributary input (paleo-Bonaparte River), paleoflows record a dominance of sediments derived from the tributary. Ripple cross-laminated sand lithofacies dominate while laminated silt is less common. These lithofacies are punctuated by occasional diffusely graded sand beds. The lowest diamicton unit (0-1.3 m) is Fraser till (Clague, personal communication 2002). Refer to Figure 3.9 for legend.
Appendix 3.3: Sedimentary Section 13 log (location in Fig. 3.1; part of unit 6, Bonaparte Stratigraphy Section, Fig. 3.2). This section is proximal to a tributary input (paleo-Bonaparte) and thus contains lithofacies that record a high energy lake bottom environment: a thick bed (6.3-9.1 m) of normally graded sand containing large silt rip-ups with a loaded lower contact indicating rapid deposition, plane-bedded and diffusely graded sand, ripple cross-laminated sand, and some laminated silt. Refer to Figure 3.9 for legend.
Appendix 3.4: Sedimentary Section 14 log (location in Fig. 3.1). A medium-energy lake bottom environment is recorded by lithofacies of laminated silt (suspension) with some ripple cross-lamination (turbidity currents and possible subaqueous failures), convoluted beds (subaqueous failure), and laminated clayey-silt (suspension). Bedding architecture dips 4° to 8° toward the valley centreline and likely encouraged subaqueous failures and directed flow paths of turbidity currents. Refer to Figure 3.9 for legend.
Appendix 3.5: Sedimentary Section 16 log (location in Fig. 3.1). Silt with sand inclusions dominate indicating that subaqueous slumping and loading were important processes in this area of the basin. Prior to failure it is likely that interbedding of silt and sand lithofacies existed here providing ideal conditions to form such thick beds (up to 6m) of silt with sand inclusions. Large faults (possibly related the melting of buried ice) extend through the entire exposure and cross the logged section at 14 m and 25 m. Significant rotation along the curvi-planar fault at 25 m caused beds above the fault to dip north (right) and those below to dip to the south (left). Lake sediments (0 - 28.8 m) are overlain by Holocene floodplain gravel.
Appendix 3.6: List of sedimentary section locations and names

<table>
<thead>
<tr>
<th>Section no.</th>
<th>Coordinates&lt;sup&gt;1&lt;/sup&gt;</th>
<th>Name&lt;sup&gt;2&lt;/sup&gt;</th>
</tr>
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</table>

<sup>1</sup> UTM, Zone 10, NAD 83
<sup>2</sup> General location shown in Fig. 2.1
**Appendix 3.7: General observations for sections described in limited detail**

<table>
<thead>
<tr>
<th>Section no.</th>
<th>General observations</th>
</tr>
</thead>
<tbody>
<tr>
<td>1b</td>
<td>~25 m high by ~100 m long. Dominated by ripple cross-laminated sand sequences with lesser diffusely graded sand and laminated silt lithofacies. Paleoflows are consistently east (apparent). Numerous truncation surfaces. At and west of this site for ~100 m is a pronounced undulatory contact between deglacial lacustrine sediments and underlying gravel and diamicton. Section stratigraphically underlies Section 4a based on eastward dipping architecture. No access to sediments above ~25 m although sediments appear similar to Section 4a.</td>
</tr>
<tr>
<td>1c</td>
<td>~15 m high by ~50 m long. Composed of two stratigraphic units. Unit 1 (at bottom; ~15 m thick) is composed of filled, convoluted and faulted beds of laminated silt and convoluted silt with gravel inclusions; Unit 2 (~15 m thick) overlaps unit 1, has a steeply-angled lower contact and is dominated by ripple cross-laminated sand. As the contact between unit 1 and unit 2 is steeply angled, both units extend from the bottom of the section to an overlying bed of terraced scoured gravel.</td>
</tr>
<tr>
<td>4</td>
<td>~20 m high by &gt;100 m long. Composed of interbedded ripple cross-laminated sand, diffusely graded sand and laminated silt. A series of large subparallel normal faults cut through the section. Faults dip to the east while the bedded sediments dip to west. Paleoflows of ripples consistently west (apparent).</td>
</tr>
<tr>
<td>5</td>
<td>~20 m high by ~80 m long. Composed of two stratigraphic units: Unit 1 (~5-20 m thick) is composed of convoluted silt, laminated silt and silt with sand inclusions; Unit 2 (~15 m thick) thickness) oversteps unit 1 and has an erosional, loaded, and concave-up lower contact that extends to the top of the section. Unit 2 is dominated by ripple cross-laminated sand (paleoflows consistently west (apparent)) with some laminated silt and trough cross-bedded sand.</td>
</tr>
<tr>
<td>6</td>
<td>~30 m high by &gt;200 m long. Dominated by convoluted silt with mega-inclusions, and some laminated silt and ripple cross-laminated sand. Some tilted blocks (&gt;40 m) of sediment observed.</td>
</tr>
<tr>
<td>7</td>
<td>~10 m high by ~30 m long. Stratigraphic section - three units: Unit 1 (at bottom; ~10 m thick) is a highly compact matrix supported diamicton overlying bedrock. Striated and faceted clasts somewhat common. Clast fabric completed (n=25, trend=303°, plunge=19°, S1=0.80, S2=0.03). Unit 2 (~2.5 m variable thickness) has a undulating lower contact and is a poorly sorted, matrix (silt to gravel) supported, pebble to boulder gravel. Unit 3 (~3 m thick) has an undulatory lower contact and somewhat interfingers with unit 2. Unit 3 is dominantly laminated silt.</td>
</tr>
<tr>
<td>8</td>
<td>Section is partially described in Chapter 4 (Fig. 4.5). ~35 m high by ~100 m long. Section cuts edge of kettle hole (Chapter 4) and is composed of three stratigraphic units. Unit 1 is at the northern end of the section, is most distant from the kettle hole, and dominated by laminated silt that is horizontally bedded to the north but increasingly downwarped moving south along section. Unit 2 is south of but overlies unit 1 and is dominated by convoluted silt with sand inclusions. Unit 3 is south of unit 2, and is separated by an apparent south-dipping normal fault, and is dominated by ripple cross-laminated sand. All units extend from the bottom to the top of section (~35 m) and are overlain by terraced fluvial gravel.</td>
</tr>
<tr>
<td>9</td>
<td>~40 m high by &gt;100 m long. Dominated by convoluted silt with mega-sand inclusions (some inclusions &gt;6 m wide). The finer grain size of inclusions makes them more difficult to detect.</td>
</tr>
<tr>
<td>15</td>
<td>~10 m high by ~15 m long. Composed of interbedded laminated silt, silt with sand inclusions and diffusely graded sand. Dike of diffusely graded sand runs from bottom of section and possibly terminates past very up section. Dike cuts through a few faults that cut through the section. Section is overlain by a a thick (~6 m) loess cover, fluvial gravel is absent.</td>
</tr>
<tr>
<td>17</td>
<td>~3 m high by ~30 m long. Composed of interbedded laminated silt and silt with sand inclusions.</td>
</tr>
<tr>
<td>18</td>
<td>~40 m high by &gt;100 m long. Dominated by convoluted silt with mega-inclusions.</td>
</tr>
<tr>
<td>19</td>
<td>~50 m high by &gt;100 m long. Dominated by convoluted silt with sand mega-inclusions.</td>
</tr>
<tr>
<td>21</td>
<td>~75 m high by ~150 m long. Generally composed of interbedded laminated silt, stony silty diamicton, diffusely graded sand, ripple cross-laminated sand and imbricate and massive gravel. Bedding architecture dips toward valley centreline.</td>
</tr>
<tr>
<td>22</td>
<td>Various small (~2-10 m high) exposures along road-cut and gullies. Generally composed of near-horizontally interbedded, and laminated silt, stony silty diamicton, diffusely graded and plane bedded sand, ripple cross-laminated sand and imbricate and weakly stratified gravel. Bedding architecture dips toward valley centreline. Paleoflows of ripples toward valley centreline.</td>
</tr>
<tr>
<td>23</td>
<td>~10 m high by 5 m long. Composed of interbedded and alternating laminated silt and diffusely graded sand.</td>
</tr>
<tr>
<td>24</td>
<td>~50 m high by &gt;100 m long. Generally composed of stratified compact stony silty diamicton, diffusely graded sand, imbricate gravel and possible undulatory stratified gravel; laminated silt very dominant above section.</td>
</tr>
</tbody>
</table>

1 Observations limited because of access and time restrictions. Refer to Fig. 3.1 and Appendix 3.6 for location.
Appendix 4: Selected photographs of sediments in the study area
Appendix 4.1: (a) Rockfall deposit (within dashed lines; bimodal angular boulder bed) in glaciolacustrine sediments. Section 10 just off left side of photograph. Source of rockfall is rock face above. Person for scale. Figure 3.4h shows close-up of this deposit. (b) Sedimentary Section 11 (eastern portion; location in Fig. 3.1). View to the northwest showing dominance of silt lithofacies (unit 2, Fig. 3.6). Person (circle) for scale.
Appendix 4.2: Sedimentary Section 11 (location in Figs. 3.1, 3.6a). (a) Looking southeast along section. Unit numbers correspond to those in Figure 3.6. Notable is the southeastward dipping surface (unit 1 - unit 2 contact) of the subaqueous fan. (b) Undulatory stratified gravel lithofacies (within dashed lines, Table 3.1) shown in detailed sedimentary Section 11 (Appendix 3.1).
Appendix 4.3: (a) Close-up of a portion of Sedimentary Section 11 showing trough-cross bedded gravel (middle of photograph; location in Figure 3.6a). Note that some gravel is openwork. (b) Fold in lake bottom sediments of Sedimentary Section 24 (location in Fig. 3.1).
Appendix 4.4: (a) Typical fine-grained flame structures and micro-flame structures in glaciolacustrine sediments (Section 11, Fig. 3.1, Appendix 3.1). (b) Load and flame structures produced by rapid loading of coarse-grained sediments (unit 1, Section 11, Fig. 3.1 and 3.6).
Appendix 5: Selected oblique ground photographs of landforms and valley character in the study area
Appendix 5.1: (a) Large kettle hole in Deadman Delta (location in Fig. 2.6 and 3.1). Kamloops Lake in the background. (b) Paleo-channel (arrow) of the Paleo-Bonaparte River (south of Coyote Hill, Fig. 2.9).
Appendix 5.2: (a) The Twaal subaqueous fan (SF) towers \( \sim 200 \) m above present river level (location in Fig. 3.1). View to the north. (b) Ashcroft drumlin field (location in Fig. 2.9). Tarps for Ginseng farming in foreground.
Appendix 5.3: (a) View east of Savona looking east over Kamloops Lake. An ice mass was in the space presently occupied by the lake when adjacent areas (e.g. (b)) were filling with glaciolacustrine sediments (Chapter 2). (b) View to the west showing Ashcroft (A), Coyote Hill (C) and Elephant Hill (E). Sedimentary Section 11 (Fig. 3.6) in the foreground.
Appendix 5.4: (a) View north from 13 km south of Ashcroft. Coyote (C) and Red hills (R) in the background. Well-exposed fluvial terraces (T) overlying glaciolacustrine sediments (Section 18, Fig. 3.1) in the midground. (b) Complex of multiple-leveled fans (F) 16 km south of Ashcroft that are incised into glaciolacustrine sediments (white cliffs) and higher level fluvial terraces (T), and deposited onto lower level fluvial terraces.
Appendix 5.5: (a) View to the north from near Inkikuh Creek, 24 km south of Ashcroft (Fig. 3.1). Section 20 in the background. Note the narrowing of the valley from background to midground and steep valley sides. (b) View to the south in the Lower Thompson Valley, 23 km south of Spences Bridge. The valley is narrowest and glaciolacustrine sediments are absent.
Appendix 6: Selected oblique aerial photographs of the Thompson Valley
Appendix 6.1: Oblique aerial photograph (BC 376-90, copyright British Columbia Government, by permission). View from Deadman River delta (foreground; Fig. 3.1) looking east towards Kamloops Lake and Kamloops (K).
Appendix 6.2: Oblique aerial photograph (BC 359:30, copyright British Columbia Government, by permission). View looking south, Thompson River, Brassy Creek terrace (B; Fig. 1 and X) and Section 3.
Appendix 6.4
Chiefly an aerial photograph (8" x 10") of the British Columbia coast. In permanent view, looking north along Alaska, a typical
fjord (1) and peninsulas (2). Section 20 in middle ground, the greater concentration of tributaries and narrower valley in the north (of Appendix 6.3).
Appendix 6.5: Oblique aerial photograph (BC 653:62, copyright British Columbia Government, by permission). View looking south from Spences Bridge (foreground) along the lower Thompson Valley. The valley is its narrowest south of Spences Bridge. Fraser Valley (F) in the background.
Appendix 6.6: Oblique aerial photograph (BC 498-38, copyright British Columbia Government, by permission). View from the vicinity of Lytton looking northeast up the lower Thompson Valley. (A) Ashcroft, (SB) Spences Bridge, and (K) Kamloops. The valley is very narrow here.