INFORMATION TO USERS

This manuscript has been reproduced from the microfilm master. UMI films the text directly from the original or copy submitted. Thus, some thesis and dissertation copies are in typewriter face, while others may be from any type of computer printer.

The quality of this reproduction is dependent upon the quality of the copy submitted. Broken or indistinct print, colored or poor quality illustrations and photographs, print bleedthrough, substandard margins, and improper alignment can adversely affect reproduction.

In the unlikely event that the author did not send UMI a complete manuscript and there are missing pages, these will be noted. Also, if unauthorized copyright material had to be removed, a note will indicate the deletion.

Oversize materials (e.g., maps, drawings, charts) are reproduced by sectioning the original, beginning at the upper left-hand corner and continuing from left to right in equal sections with small overlaps. Each original is also photographed in one exposure and is included in reduced form at the back of the book.

Photographs included in the original manuscript have been reproduced xerographically in this copy. Higher quality 6" x 9" black and white photographic prints are available for any photographs or illustrations appearing in this copy for an additional charge. Contact UMI directly to order.

UMI

A Bell & Howell Information Company
300 North Zeeb Road, Ann Arbor MI 48106-1346 USA
313/761-4700 800/521-0600

**The sedimentology and paleogeography of Glacial Lake Champagne, southern
Yukon Territory**

Scott Donald Barnes, B.Sc.

**A thesis submitted to the Faculty of Graduate Studies and Research as partial fulfillment
of the degree of**

Master of Arts

**Department of Geography
Carleton University**

Ottawa Ontario

September 19, 1997.

**© copyright 1997
Scott Donald Barnes**

National Library
of Canada

Acquisitions and
Bibliographic Services

395 Wellington Street
Ottawa ON K1A 0N4
Canada

Bibliothèque nationale
du Canada

Acquisitions et
services bibliographiques

395, rue Wellington
Ottawa ON K1A 0N4
Canada

Your file Votre référence

Our file Notre référence

The author has granted a non-exclusive licence allowing the National Library of Canada to reproduce, loan, distribute or sell copies of this thesis in microform, paper or electronic formats.

The author retains ownership of the copyright in this thesis. Neither the thesis nor substantial extracts from it may be printed or otherwise reproduced without the author's permission.

L'auteur a accordé une licence non exclusive permettant à la Bibliothèque nationale du Canada de reproduire, prêter, distribuer ou vendre des copies de cette thèse sous la forme de microfiche/film, de reproduction sur papier ou sur format électronique.

L'auteur conserve la propriété du droit d'auteur qui protège cette thèse. Ni la thèse ni des extraits substantiels de celle-ci ne doivent être imprimés ou autrement reproduits sans son autorisation.

0-612-26901-9

Canada

Carleton University
Ottawa, Canada K1S 5J7

1942-1992

Thesis contains black and white and/or
coloured graphs/tables/photographs which when
microfilmed may lose their significance. The
hardcopy of the thesis is available upon
request from Carleton University Library.

The undersigned recommend to
The Faculty of Graduate Studies and Research
acceptance of the thesis

**The sedimentology and paleogeography of Glacial Lake Chamapgne, southern
Yukon Territory**

submitted by
Scott Donald Barnes
in partial fulfillment of the requirements for
the degree of Master of Arts

Thesis Supervisor

Chair, Department of Geography

Carleton University

September 19, 1997

Abstract

Glacial Lake Champagne was a major feature of the deglacial landscape of southern Yukon Territory. The sedimentology reflects two major depositional environments; the settling of sediment laden plumes of meltwater from the surrounding ice sheets, and deposits of density underflows. Initial stratigraphic and geomorphic evidence suggests that much of the lake was bounded by active or stagnant ice of the Cordilleran ice sheet. There is one well preserved stand of the lake at 765 m, and a less well preserved stand 725 m above sea level. The 765 m stand of the lake covered approximately 2425 km², and had a maximum depth of at least 159 m. Glacial Lake Champagne occupied the valleys of the Yukon and Dezadeash Rivers and tributaries; and the valleys of the Southern Lakes district. Initial investigation suggests that the lake existed sometime between 12 500 and 10 500 BP, for a period of up to 400 years, based on sediment accumulation.

I would like to thank all of the people who helped make this thesis possible. All of the people listed here contributed in their own way to the completion of this thesis, from technical advice, to commiserating over beer. If I have missed anyone, I apologize, as it was not on purpose.

Jeff Hughes, who was a very capable and enthusiastic field assistant, as well as Jim Howes who provided unique insight in the field, and Bill Laberge who visited my field sites with me.

The people of the Carcross-Tagish region who made us feel welcome, and had good fishing tips.

Ken Torrance, whom I consulted about grain size analysis.

My roommates, Kari and Steve for putting up with me during the process, and offering (occasionally) sage advice.

My friends in Ottawa: Brenda, Alice, Paula, Michael, Alette, John and Debbie.

My friends in Calgary: Derek, Rakowski, Popo, Coleman, Heather, Andrew, Dave, Todd, Nancy, Mike, Tannis, Meier, Verot, Jason, and Carol; thanks for keeping me in the loop.

Bob and Sylvia, for letting me go to San Diego with them, among other things.

Lindsay, for all of her help. Thanks.

Joyce Lundberg, who supported this project financially, and intellectually as my supervisor. Without her energies, this would not have been possible.

Last, but not least, my family: Don, Sharon, Wade, Heather, Garrett, and Jack.

Table of contents

1.0 Introduction	1
1.1 Statement of Problem	1
1.2 Previous research	5
1.3 Thesis organization	10
2.0 Regional Setting	12
2.1 Physiography	12
2.2 Quaternary evolution	14
2.3 Permafrost	16
2.4 Eolian activity	18
3.0 Properties of Glacial Lakes	19
3.1 Introduction	19
3.2 Dynamics of glacial lakes	19
3.3 Sedimentology of glacial lakes	21
3.4 Glacial lake reconstruction and nomenclature	31
4.0 Methodology	33
4.1 Introduction	33
4.2 Map and airphoto interpretation	35
4.3 Field work	36
4.4 Lab analysis of sediments	37
4.5 Data analysis	38
5.0 Sedimentology of deposits of Glacial Lake Champagne	39
5.1 Introduction	39
5.2 Location and situation of stratigraphic logs	41
5.3 Stratigraphic logs	42
5.4 Periods of deposition	50
5.5 Quaternary sedimentary facies	51
5.6 Local paleoenvironmental interpretations	55
5.7 Regional depositional environments in Glacial Lake Champagne.	66
6.0 Paleogeography of Glacial Lake Champagne	68
6.1 Introduction	68
6.2 Stratigraphic and geomorphic evidence: Glacial Lake Champagne	69
6.3 Re-evaluation of the previously published paleogeography of Glacial Lake Champagne	78

6.4 Geometry of the ice front-Glacial Lake Champagne interface	80
6.5 Geometry of Glacial Lake Champagne	84
7.0 Holocene geomorphic evolution in Glacial Lake Champagne deposits	89
7.1 Introduction	89
7.2 Eolian features and impact on the landscape	89
7.3 Thermokarst development	91
8.0 Conclusions and recommendations for future research	92
8.1 Sedimentology of the Glacial Lake Champagne basin	92
8.2 A tentative reconstruction of the geomorphic evolution of the Glacial Lake Champagne basin	96
8.3 Recommendations for future research	97
References	99

List of tables

Table 1: Subdivision of Quaternary events and deposits in Yukon Territory.	17
Table 2: Characteristics of ice rafted sediments	30
Table 3: Location and situation of sites of stratigraphic logs.	41
Table 4: Facies associations key for Figures 11-16.	43

List of figures

Figure 1a: Study area, southern Yukon Territory.	3
Figure 1b: Topography of the study area.	4
Figure 2: Limit of Pleistocene glaciations, Yukon Territory.	7
Figure 3: Ice flow direction, southern Yukon Territory.	9
Figure 4: Generalized physiography of southern Yukon Territory.	13
Figure 5: Idealized sedimentary lacustrine features that result from changes in clastic sediment input, and water column stratification in an oligotrophic lake.	24
Figure 6: Factors that lead to the formation, preservation and destruction of rhythmites in lakes.	26
Figure 7: Location of sites where detailed stratigraphic logging took place.	40
Figure 8: Stratigraphic log of the Whitehorse-Yukon River site.	44
Figure 9: Stratigraphic log of the Watson River site.	45
Figure 10: Stratigraphic log of the Tagish Lake site.	46
Figure 11: Stratigraphic log of the Whitehorse-Riverdale site.	47
Figure 12: Stratigraphic log of the Marsh Lake Dam site.	48
Figure 13: Stratigraphic log of the Millhaven Bay site.	49
Figure 14: Example of facies St and Sr in the Whitehorse-Riverdale site.	53
Figure 15: Massive to weakly bedded glaciolacustrine silts in the Whitehorse-Yukon River site.	57
Figure 16: The Whitehorse-Riverdale section.	62
Figure 17: Locations of exposed glaciolacustrine sediments, deltas, and shorelines associated with the 765 m or 725 m stand of Glacial Lake Champagne.	70
Figure 18: Large section of glaciolacustrine sediments exposed by the Takhini River.	71
Figure 19: The delta at Lime Creek.	73
Figure 20: The delta at Watson River.	75
Figure 21: Thermokarst and shorelines in the Takhini Valley.	77
Figure 22: Inferred ice margins in the study area during the time Glacial Lake Champagne existed.	82
Figure 23: Geometry of Glacial Lake Champagne at the 765 m stand.	85
Figure 24: Eolian deposit at the Marsh Lake Dam site.	90

List of appendices

Appendix A: List of airphoto flight lines and photo numbers	108
Appendix B: Results of grain size analysis of samples	109

1.0 Introduction

1.1 Statement of Problem

Southern Yukon is an area of deeply incised valleys, carved into the Yukon Plateau north and east of Carcross, and into the Coast Range south and west of Carcross (Figure 1a, p. 3; Figure 1b, p. 4). Much of the valley bottom environments in this region are blanketed by lacustrine silt and sands, and deltaic gravels. The glacial lake associated with these deposits has been discussed by a number of researchers, but not identified as a contiguous isochronous unit throughout the area. Names given to the local lake stages are Glacial Lake Champagne, near Champagne, Yukon Territory (Kindle, 1953), and Glacial Lake Carcross in the area near Carcross, Yukon Territory (Wheeler, 1961; Hart and Radloff, 1990), but this previous work provided no regional correlation of the deposits. In addition to this, broad areas of lacustrine deposits have not been documented in any way.

Reconnaissance fieldwork indicated extensive and often up to 30 m thick glaciolacustrine deposits occupying all the valleys floors, and reaching a height of 765 m. It appeared that Glacial Lake Champagne might be a far more extensive and important feature than had been previously recognised; that Glacial Lake Champagne might in fact be one of the most important late glacial features in southern Yukon, and that the Holocene geomorphic evolution of southern Yukon might have been greatly influenced by the presence of sediments of Glacial Lake Champagne. Preliminary stratigraphic logging suggested that the sedimentology developed primarily as a result of either suspended sediment deposition, or density underflow deposition. It became obvious that a detailed

study of the sedimentology was required. The research for this thesis was thus planned with the following aims:

- a) to map the distribution of glaciolacustrine sediments and geomorphological features, such as shorelines, in order to establish the paleogeography of Glacial Lake Champagne. The field area, dictated by locations of glaciolacustrine sediments, extended over the Southern Lakes District, the Whitehorse area north to Lake Laberge, and west through the Takhini and Dezadeash River valleys, and tributaries (Fig 1, p. 3);
- b) to produce detailed stratigraphic logging of selected sites chosen as far as possible to adequately represent the former lake basin;
- c) to carry out facies analysis of the sedimentary packages and produce facies models in order to interpret the glaciolacustrine depositional environments in the lake basin.

Field work was undertaken between May 15, and July 31, 1996. Work focussed on field mapping of shoreline and delta features, facies logging and description, and lacustrine sediment sampling. Sites were accessed by four wheel drive vehicle on the numerous roads and tracks in the region, by boat on the Southern Lakes system, and by walking to sites that were otherwise inaccessible.

The available field evidence allowed the immediate aim, to document the sedimentary history of the lake and to reconstruct its shorelines at the maximum stillstand, to be achieved. However, it did not allow the wider-scale aim, to reconstruct the paleogeography of the lake, to be fully realised. Since all the obvious and readily accessible geomorphic features were studied it is clear that further, much more extensive and expensive geomorphological mapping would be required to understand the complete

Figure 1a: Study area, southern Yukon Territory.

history of the lake. This thesis therefore presents (i) the results of the sedimentological analysis; and (ii) maps of the former extent of Glacial Lake Champagne at its maximum stable stand at 765 m; and (iii) some conjectures about the paleogeography based on quite limited evidence.

The research was considered to be important from many points of view:

- a) The documentation of Glacial Lake Champagne would provide important baseline information to help fill in the deglacial history of southern Yukon Territory, and bring up to date, and tie together, the earlier research into Glacial Lake Champagne and Glacial Lake Carcross. Since it developed in an environment that is poorly understood in terms of the retreat of ice lobes, it was hoped that the position of Glacial Lake Champagne would shed light on the retreat of these lobes during deglaciation.
- b) Glacial lakes are common deglacial features in the Canadian Cordillera, and have been documented in terms of sedimentology and paleogeography; the documentation of Glacial Lake Champagne would add to this body of information and enhance our understanding of the behaviour of glacial lakes. However, Glacial Lake Champagne, because of its large size and northern position, might prove to be a unique cordilleran glacial lake and thus the sedimentological study might produce new models.

1.2 Previous research

The majority of the Quaternary research completed in Yukon Territory has been large scale ice reconstructions. These were synthesized by Hughes (1987), as part of the XIIth INQUA Congress excursions, and by Clague (1989), as part of the summary of the

Quaternary evolution, and sediments of the Canadian Cordillera. The accepted chronology of Quaternary evolution for Yukon Territory was developed primarily by Bostock (1966), based on evidence found in the central region of the territory. Bostock described four major glacial events, with each subsequent event being less areally extensive than the prior one (Figure 2, p. 7). The advances are termed, from oldest to youngest Nansen, Klaza, Reid, and McConnell. The advances were limited in areal extent and a large part of central Yukon Territory was not glaciated in the Pleistocene (Clague, 1989). Bostock's chronology was based primarily on the "freshness" of the landforms associated with each advance. Further evidence for the relative age of each advance has been provided by the analysis of the degree of soil development on each deposit (Foscolos *et al.*, 1977; Tarnocai *et al.*, 1985; Smith *et al.*, 1986).

During the last glacial maximum, the St. Elias Mountains gave rise to a piedmont glacier complex that was contiguous with the rest of the Cordilleran ice complex (Hughes, 1987). Rampton (1971) identified two advances in the Snag Klutlan area, which were designated Mirror Creek and MacCauley Glaciations, that are correlated with Bostock's Reid and McConnell advances, based on preservation (Figure 2, p. 7). Three tills occur at the south end of Kluane Lake (Denton and Stuiver, 1967; Table 2). Correlation of the Kluane deposits with the Snag Klutlan deposits remains uncertain (Rampton, 1971). Cordilleran ice in the study area had three major source areas. Most of the ice from the Cassiar Mountains region trended northwest, forming the Cassiar Lobe (Wheeler, 1961). This ice covered the Teslin and Whitehorse area (Figure 3, p. 9). Ice flow directions were strongly influenced by local topography (Jackson *et al.*, 1991). Ice from the St. Elias

Figure 2: Limit of Pleistocene glaciations, Yukon Territory. (Source: Hughes *et al.*, 1969; Duk-Rodkin and Hughes, 1991)

Mountains trended easterly, and was much less extensive than the Cassiar ice complex. This is because St. Elias ice formed in the rain shadow of the St. Elias Range, whereas the Cassiar ice formed on the windward flank of the Cassiar Mountain Range (Hughes *et al.*, 1968). Coast Mountain ice also formed on the leeward side of the Coast Mountain Range, and trended north and east (Hughes *et al.*, 1968). During the climax of the Last Major Glaciation the firn line fell to an elevation of 1500 m (Jackson *et al.*, 1991). In the area in which Glacial Lake Champagne formed, glacial ice was a product of coalescence from these three sources, and retreat occurred back toward each of the source areas (Hughes *et al.*, 1968). The maximum depth of ice was 1982 m in the study area (Kindle, 1953; Wheeler, 1961) based on the distribution of erratics on slopes. Jackson *et al.* (1991) state that the relief was typically equal to the thickness of ice. Peaks higher than this existed as nunataks (Kindle, 1953; Wheeler, 1961). Mann and Hamilton (1995) state that seaward Coast Ranges glaciers had retreated to their modern positions by 13,500 BP.

During deglaciation, numerous glacial lakes were formed. Kindle (1953), identified Glacial Lake Champagne, described as occupying the Dezadeash River valley, and tributaries up to an elevation of 854 m asl. Rampton (1972) describes the occurrence of glaciolacustrine and other sediments along a transect between Whitehorse and the Alaska Yukon Territory boundary, some of which are associated with the evolution of Glacial Lake Champagne. Day (1962) produced a map and report of soil deposits in the Takhini valley. Much of the soil development was reported to be in sediments of Glacial Lake Champagne. Other glacial lakes have been described throughout southern and southwestern Yukon Territory, including Glacial Lake Kloo (Kindle, 1953), Glacial Lake

Figure 3: Ice flow direction, southern Yukon Territory (Source: Jackson *et al.*, 1991). Key: SEPLC - St. Elias piedmont lobe complex; EL - Eastern Coast Ranges lobe; EC - Eastern Coast Ranges; CL - Cassiar Lobe; C - Cassiar Mountains; P - Pelly Mountains; SL - Selwyn lobe; S - Selwyn Mountains; LL - Liard lobe. Note the divide between the Eastern Coast Mountain lobe, the Cassiar lobe, and the St. Elias piedmont lobe has not been defined.

Jo-Jo (Kindle, 1953), Glacial Lake Carcross (Wheeler, 1961; Hart, 1990), Glacial Lake Nisling (Geurts, 1993). The location of other glacial lakes is important because it indicates the pattern of ice retreat. Morison and Klassen (1991), prepared a map of the surficial geology for the 105 D map sheet, which comprises approximately one half of the study area. This map identifies shorelines and sediments associated with Glacial Lake Champagne, but did not identify any of the deltas described later in the text. This literature review is brief because the published research is limited for the study area, with only the one direct reference to Glacial Lake Champagne.

1.3 Thesis organization

This thesis is organized as follows. Chapter 2 will discuss the regional setting of the study area. Chapter 3 will discuss general aspects of glacial lakes in order to put the subsequent discussion of the findings of the thesis research in perspective. Chapter 4 is a brief discussion of the methodologies used in conducting research for the thesis. Chapter 5 is a presentation, discussion and interpretation of the sedimentology observed in six sections of largely glaciolacustrine sediment in the study area. Chapter 6 discusses aspects of the paleogeography of Glacial Lake Champagne at two recognized stable stands. The nature of the sediments presented in Chapter 5 is used as evidence to define the paleogeography in Chapter 6. Chapter 7 presents limited findings of some of the Holocene geomorphic evolution in Glacial Lake Champagne sediments. Strictly speaking, this is not a part of the main thesis, but is included as it documents interesting findings, and adds a measure of completeness. Chapter 8 concludes and summarizes the findings of the

research discussed in the thesis, and suggests future research that should be undertaken to better understand aspects of Glacial Lake Champagne.

2.0 Regional Setting

2.1 Physiography

The study area falls within the boundaries of the major physiographic unit of the Canadian Cordillera, which is a northwest to southeast trending belt of mountains, plateaux, and valleys, covering an area in excess of 1 500 000 km². This westernmost physiographic region in Canada has been undergoing uplift at different rates in the region since the Cretaceous Period until the present and continues today (Clague, 1989).

The physiographic subunits of the region are the Boundary Ranges (Coast Mountains) and the Yukon Plateau (Figure 4, p. 13). The Boundary Ranges, which are northwest southeast trending unit of the Coast Mountains of the Canadian Cordillera, reach maximum local elevations of 2400 m. These mountains have an upland surface at around 1800 m, with a few rugged peaks reaching above this. The valleys are cut into the upland surface about 900 m (Bostock, 1948).

The Yukon Plateau is a broad tableland that is cut by a network of valleys within the Pleistocene glacial limits (Hughes, 1987). The Yukon Plateau rises towards the southern parts of the study area, where it has an indistinct boundary with the Boundary Ranges. The valleys become deeper and more narrow in the south, creating greater local relief. Towards the north, the occurrence of uplands is reduced, as the landscape becomes more dominated by the wide valleys of Marsh Lake, M'Clintock River, and Lewes River (Bostock, 1948). The isostatic rebound in the study area is not documented, but in the Aishihik region which lies immediately north of the western part of the study area, the

Figure 4: Generalized physiography of southern Yukon Territory. (Source: GSC, 1991).

potential isostatic rebound was calculated to be 237 m, based on a maximum ice cover of 1065 m, and an average ice cover of 710 m (Klassen, 1990). No other estimates for the amount or rate of isostatic rebound exist for the study area, but could potentially be determined from shoreline warp in the Dezadeash Valley which has the longest continuous shorelines.

2.2 Quaternary evolution

The entire study area was glaciated during the Last Glacial Maximum. The ice was part of the Cordilleran Ice Complex, which, at its maximum, covered nearly all of British Columbia, and southern Yukon Territory (Clague, 1989). The early part of each glaciation was typified by mountain (cirque) glaciers growing to become mountain ice caps. Piedmont lobes developed next and these coalesced along mountain fronts (Fulton, 1991). The regional drainage was altered by the advancing piedmont lobes by damming large lakes, such as Glacial Lake Stikine (Ryder and Maynard, 1991), and causing the aggradation of streams by overloading them with outwash materials.

At the climax of each glaciation the coalescent piedmont lobes formed a single Cordilleran Ice Complex. The erosive power of the advancing ice removed sediments previously deposited in almost all cases. The highest peaks in the Canadian Cordillera existed as nunataks. Till was deposited throughout ice covered areas. (Clague, 1989).

Deglaciation occurred mainly as downwasting of the ice sheet, with ice in valley bottoms persisting the longest. This pattern of deglaciation resulted from the ice sheet being topographically confined. The firn line would have had to rise above the elevation of the ice sheet for significant downwasting to occur in the interior. Also, the thickest part of

the ice sheet was in the valley bottoms, and not at the sites of initial accumulation in the mountain cirques (Fulton, 1991). Therefore, the advance and retreat of the Cordilleran Ice Complex did not mirror each other.

Current stratigraphic evidence suggests that the study area was covered by ice sheets at least six times during the Pleistocene (Jackson *et al.*, 1991). During the Last Glacial Maximum, known as the McConnell Glaciation in the study area, ice advanced from the high elevations of the Cassiar Mountains, the St. Elias Mountains, and the Boundary Ranges (Hughes *et al.*, 1968). The morphology of the Southern Lakes system, and Kusawa Lake suggest that these are a result of extensive valley deepening due to north flowing ice from the Boundary Ranges. At the climax of the McConnell Glaciation, ice from the eastern Coast Mountains and the Cassiar Mountains was part of the Cordilleran Ice Complex. This ice was contiguous with ice from the St. Elias Mountains which formed as a piedmont lobe (Jackson *et al.*, 1991).

Ice flow direction was strongly controlled by topography, with merging ice streams following major valleys such as the Teslin/Yukon valley (Jackson *et al.*, 1991). The basis for ice flow direction determination described by Jackson *et al.* (1991) are flow indicators such as whalebacks and crag-and-tail topography.

The Cordilleran Ice Complex disappeared rapidly at the close of the McConnell Glaciation mainly through downwasting and stagnation (Jackson, *et al.*, 1991). The firn line rose above the present day 1830 m level early in the deglaciation. Readvances have been noted in the area, including a readvance of the eastern Coast Mountains lobe in the Aishihik, West Aishihik and Jarvis Valleys (Hughes, 1990), with the Jarvis Valley being

close to the juncture of the eastern Coast Mountains lobe, and the St. Elias Mountains piedmont lobe complex (Jackson *et al.*, 1991).

Deglaciation began on the east side of the St Elias Mountains by $13\,660 \pm 460$ BP (Hughes *et al.*, 1989), and the large Kaskawulsh Glacier was even less extensive than present by 9780 ± 80 BP (Hughes *et al.*, 1989). Quaternary events and deposits are summarised in Table 1. All of the information in the table was gathered by the researchers noted, and is considered to be the most accurate depiction of the regional evolution and correlation of Quaternary evolution in Yukon Territory. The correlations within the table were made by Hughes *et al.* (1989).

2.3 Permafrost

The study area spans two zones of permafrost: alpine and scattered (Brown, 1978), which are both discontinuous. Microclimatic influences are the primary controlling mechanisms in the distribution of permafrost in the discontinuous zone (Smith and Riseborough, 1983). Permafrost is found in the study area in valley floors as discontinuous permafrost (Burn, 1987), and at higher elevations as alpine permafrost. In the Takhini valley, permafrost degradation has led to the development of thermokarst lakes developed in glaciolacustrine sediments (Klassen, 1979).

Slopes that support permafrost are more prone to movement that may disrupt the stratigraphic record than slopes that do not (Williams and Smith, 1989). This may have impacted the preservation of lake features (Schmok and Clarke, 1989).

Table 1: Subdivision of Quaternary events and deposits in Yukon Territory. (Source: Hughes *et al.*, 1989).

	Yukon Plateaus (Boatich, 1968; Hughes <i>et al.</i> , 1989)	Snag-Klurlian area (Rampton, 1971a)	Silver Creek (Denton and Stuver, 1967)	Liard Lowland (Klassen, 1978, 1987)	Southern Ogilvie Mins (Vernon and Hughes, 1966)	Old Crow Flats (Morlan, 1980)
HOLOCENE	POSTGLACIAL	NEOGLACIAL	NEOGLACIAL	POSTGLACIAL		
			SUMS NONGLACIAL INTERVAL 12.5 ka			12.5 ka
PLEISTOCENE	MCCONNELL GLACIATION	MACAULEY GLACIATION	KLUANE GLACIATION	TILL D	LAST GLACIATION	GLACIAL-LACUSTRINE CLAY
	REID SOIL		< 29.6 ka	< 23.9 ka		< 25.2 ka
	SHEEP CREEK TEPHRA (0.84-0.84 Ma)	OLD CROW TEPHRA (0.120 ka, > 48 ka)	BOUTELLIER NONGLACIAL INTERVAL	INTERMILL UNIT C-D		"DISCONFORMITY A" ALLUVIUM QUP P90W TEPHRA, (0.130 ka)
	REID GLACIATION	MIRROR CREEK GLACIATION	ICEFIELD GLACIATION	TILL C	> 53.9 ka INTERMEDIATE GLACIATION	
	PRE-REID SOIL		SILVER NONGLACIAL INTERVAL	INTERMILL UNIT B-C7		ALLUVIUM
	KLAZA GLACIATION	?	SHAWWAK GLACIATION?	TILL B?	OLD GLACIATION?	
	< 1.08 Ma					
	FORT SELKIRK TEPHRA (0.84-0.84 Ma)			INTERMILL UNIT A-B7	MOSQUITO GULCH TEPHRA (1.22 Ma)	LITTLE TIMBER TEPHRA (1.12 Ma)
	NANSEN GLACIATION		SHAWWAK GLACIATION?	> 0.76 Ma TILL A?	OLD GLACIATION	LACUSTRINE CLAY

2.4 Eolian activity

Holocene development of the landscape has also been influenced by eolian activity. Near Whitehorse, there are units of sand that have been mapped by Morison and Klassen (1991) as eolian deposits. Large parts of the study area are topped by eolian deposits reworked from glaciolacustrine sediments, which are documented in Chapters 5 and 7.

3.0 Properties of Glacial Lakes

3.1 Introduction

Glacial lakes exist, or did exist, at a certain level because of an ice barrier on the lake's periphery that effectively blocks lower potential outlets (Farrand and Drexler, 1985). Therefore, a change in the position of a glacial ice blockage will directly impact the extent of a glacial lake. The shorelines of lakes represent an isochronous surface formed at the interface of land (or ice), water, and atmosphere (Goudie *et al.*, 1978), which means that the geometry of the glacial lake in question is defined by the position of the ice front that provides the hydrologic base level. This allows interpretation of the evolution of the deglacial landscape. Chronology is usually relative, but can be absolute where organic or other datable material exists (examples: Miller *et al.*, 1985; Sawicki, 1990).

3.2 Dynamics of glacial lakes

A change in glacial lake geometry (areal extent and depth) is dependent on the interaction of three factors: (1) position of the ice margin; (2) glacioisostatic delevelling; and (3) outlet incision (Lemmen *et al.*, 1994). The position of the ice margin is a product of the climatic conditions that regulate ice advance and retreat. Glacioisostatic delevelling is a function of the thickness and duration of ice cover over a large region. Outlet incision occurs when the height of water is such that it can overtop the lowest barrier, or can escape underneath or through the ice cover forming one of the shorelines (example: Sharpe and Shaw, 1989).

During the retreat of continental ice sheets, glacial lakes form in depressions. Because of the relatively long period of ice coverage, the crust will be isostatically depressed when the ice cover retreats. The abundance of meltwater will cause glacial lakes to form in the lowest areas. These lakes have been shown to drain along the ice front, as was the case for many phases of the proto-Great Lakes (Calkin and Feenstra, 1985), subglacially, or away from the ice sheets.

Changes in drainage of glacial lakes are caused by changes in the equilibrium conditions of the ice-water system. Changes in the drainage regime in a glacial lake system can be a continuum from gradual to catastrophic lowering. This is dependant on the stability of the glacial lake system, which in turn is dependant on a number of factors:

1. Hydrology of the ice-water system
 - changes in thickness and position of proximal ice
 - changes in the volume of meltwater contained in the system
2. Physical characteristics of the bounding basin
 - post-glacial topography
 - post-glacial deposits
 - geology
3. Differential basin uplift due to isostatic rebound

Glacial lakes can undergo several changes in drainage over their existence, and may catastrophically lower several times (examples: Glacial Lake Agassiz: Fisher and Smith, 1994; Huron Basin: Eschman and Karrow, 1985; Neoglacial Lake Alsek: Clague and Rampton, 1982). Each stable or near stable level of the lake is often manifested in the basin geomorphology and sedimentology.

3.3 Sedimentology of glacial lakes

Glacial lake sediments are usually deposited in environments where large loads of unsorted glacial sediment are rapidly deposited to form distinctive facies that may be subsequently further modified by the melting of ice in contact with sediments, or by the draining of ice dammed water into or out of the basin (Gilbert and Desloges, 1987).

Sediments provide a proxy record of environmental conditions within the basin at the time of deposition. By developing a facies model for a basin, based on a number of cores or exposures, the evolution of the environmental conditions within the basin can be investigated. The basis for this investigation is the comparison of modern day glacial lake processes and subsequent deposits, with those deposits left by glacial lakes that no longer exist or did exist in the past with different geometry. This allows correlation of the past deposits to modern day processes. The relationships between certain glaciolacustrine deposits and associated depositional environment(s) are developed below.

Deposition of suspended loads in glaciolacustrine environments

The sedimentary record within a basin is a product of the sediment input, the regional climate, and the limnology within the basin (Sturm, 1978).

The deposits associated with the settling out of suspended loads in a basin yield information on depositional environments, and occasionally can be used as chronostratigraphic markers which can be very important in paleoenvironment reconstruction. Suspended load can be introduced into a basin in five ways:

1. Water - load that is derived from all subaerial runoff including, but not exclusively, streams, subaerial landslides, debris flows, and particles loaded on or in ice.

This also includes particles introduced into the water column in direct contact with glacial ice by meltout, and englacial or subglacial streams

2. Air - load that is deposited within a basin and is transported there as dust by wind, rain, snow, or volcanic eruptions.

3. Autochthonous - load that is formed within the water column by chemical or biological production, and includes precipitates, feces, algae, etc.

4. Re-suspension - load that is derived from activities that occur on the bottom of the basin, and include currents, bioturbation, degassing and dewatering.

5. Upwelling - load that is a product of input via groundwater flow, subaqueous springs, or volcanic activity. (Sturm, 1978).

The importance of these inputs will be different depending on the physical properties of the basin, and the environment the basin is in. In most glaciolacustrine systems clastic sediment derived from subaerial sources, and inputs from direct ice contact would predominate (Sturm, 1978).

Suspension of particles in case 1 considered above occurs when the turbulence of a flow is sufficient to entrain a particle. The energy required to initially entrain a particle is greater than the energy needed to keep it suspended. This means that there must be a significant reduction in flow energy in order for the particle to drop out of suspension. In practical terms for glacial lakes, particles are entrained by running water under, in, on or near glacial ice, and are deposited in the lake basin as the flow energy of the stream input is decreased.

The two most important factors in defining the sedimentology of glaciolacustrine

deposits are the change in clastic influx over time, and the stratification of the water column over time. Sturm (1978), developed a model for depositional features based on the stratification of the water column, and the influx of sediment (Figure 5, p. 24). A continuous influx of suspended sediment will lead to deposits that are massive (cases 1, 2, and 5 in Figure 5, p. 24). Discontinuous influx of suspended sediments lead to a bimodal deposit (cases 3, 4, 6, 7, and 8 in Figure 5, p. 24). In non-stratified and permanently stratified water columns, sediment will settle out according to Stokes' law which defines rates of sedimentation based on grain size. These deposits do not have any inherent temporal cyclicity, as each couplet only represents a pulse of sediment of an indeterminate length of time, which may be independent of seasonality. True varves are created in case 7. The coarsest sediments settle out quickly after they are introduced into the water column, but the clay particles will be trapped in the epilimnion until the water column turns over due to seasonal change. This will create a yearly cycle that has been used as a chronostratigraphic marker for a basin. Case 8 is a modified case 7, where the coarse particles of the present pulse settle out at the same time as the fine particles from the previous pulse. This creates a sand layer within the clayey laminae of the varve couplet. Smith and Ashley (1985) provide criteria for recognizing annual rhythmmites related to seasonality; and surge rhythmmites, which are the record of episodic unsteady underflows usually lasting a few minutes or less. These criteria are:

Figure 5: Idealized sedimentary lacustrine features that result from changes in clastic sediment input, and water column stratification in an oligotrophic lake. Heavy dots represent sand, light dots represent silt, and light dashes represent clay. (Source: Sturm, 1978).

- Annual rhythmites**
1. Occur as couplet; silt layer has normal and inverse grading or no trend in grain size, clay layer fines upward.
 2. Breaks may occur within a rhythmite.
 3. Lebensspuren occur within the rhythmite.
 4. Silt layer thickness varies, clay layer thickness consistent throughout the basin.
- Surge rhythmites**
1. Fining upward, gradual decrease in silt/clay ratio
 2. No breaks within each rhythmite
 3. No lebensspuren within the rhythmite
 4. Silt and clay layer thicknesses in proportion.

Rhythmicity can occur over a number of other time scales, other than the yearly and irregular periodicity presented above. Gilbert and Shaw (1981) and Liverman (1987), present evidence that fine silt laminae may represent a diurnal period; and Smith *et al.* (1982) showed subseasonal laminations in Peyto Lake, Alberta.

In order for suspended sediment deposits to remain part of the stratigraphy of the basin, these deposits must remain undisturbed by bottom currents or bioturbation (Kempe and Degens, 1978). Figure 6 (p. 26) presents a flow chart of the factors that contribute to the formation, preservation and destruction of rhythmite deposits within basins.

Turbidity events and density underflows in glaciolacustrine environments

Turbidity events are episodic surge-type currents formed by subaqueous slumping, and density underflows are generated by underflowing sediment-laden water which produces quasi-continuous currents (Smith and Ashley, 1981), although some authors use

Figure 6: Factors that lead to the formation, preservation and destruction of rhythmites in lakes. pH has a highly complex effect, and has not been included on the flow chart. (Source: Ludlam, 1978).

the term interchangeably. Turbidity events occur when sediment undergoes mass movement underwater causing the shifting of large amounts of sediment. This generates a current which can carry sediments over distances of kilometres (Pickrill and Irwin, 1982). Turbidity events are significant in distributing sediment throughout proglacial lakes, and the resulting sedimentology has been established from the stratigraphic record (Gilbert and Shaw, 1981).

Density underflows are caused by the introduction of sediment laden water into a basin where it flows along the bottom, following the lowest topography. Underflows are capable of carrying much more sediment than can be introduced as suspended sediment in the same lake (Smith and Ashley, 1985). Underflow events tend to overwhelm the effects of thermal density in typical glacier-fed lakes during the maximum melt season because of the high concentration of sediments in the sinking plume. As underflow events tend to manifest during the period of greatest sedimentation in glacial lakes, the deposits of underflows tend to dominate the stratigraphic record (Smith and Ashley, 1985).

As underflows mix with ambient water, are slowed by friction, and deposit sediments their integrity decreases along with their velocity (Smith and Ashley, 1985). Initial velocities range from a few tens of cm/sec (Gustavson, 1975; Lambert *et al.*, 1976; Lambert and Hsu, 1979; Gilbert and Shaw, 1981; Smith *et al.*, 1982; Bogen, 1983) to in excess of 100 cm/sec (Lambert, 1982; Weirich, 1984). Strong underflows are capable of scouring channels in lake bottom sediments (Smith and Ashley, 1985).

Underflow events may continue for days (Gilbert and Shaw, 1981; Lambert *et al.*, 1976; Lambert, 1982) especially in environments of high river discharges (Smith and

Ashley, 1985). Gravitational forces dominate the lateral distributions of underflows, and are not affected strongly by Coriolis effect (Smith and Ashley, 1985).

Turbidity events and density underflows may manifest in many ways in the stratigraphic record. In most cases, turbidity events are reflected in units of relatively coarse grain size (medium sand to coarse silt) (Gilbert and Shaw, 1981). The structure of the deposits may be massive, laminated (Gilbert and Shaw, 1981), laminated units with normal grading; or cross trough beds (Shaw and Archer, 1978), or ripple-drift cross-laminated units (Kaszyski, 1987). One event is discernable when there is grading, with the event spanning from the bottom lag to the finest sediments at the top of the layer. Rip up clasts and load structures (flame, or ball and pillow) are often associated with the deposits of turbidity events (Shaw and Archer, 1978; Gilbert and Shaw, 1981; Kaszyski, 1987). One event can deposit large amounts of material, especially when compared with the rates of sedimentation from suspended sediments (Gilbert and Shaw, 1981).

The differentiation between deposits of turbidity events and density underflows is problematic (Smith and Ashley, 1985). Of the two, density underflows carry by far the bulk of the sediment distributed in glacial lakes dominated by underflows. Because turbidity events are generated by the subaqueous slumping of material, and subsequent movement, the deposits of turbidity flows can be coarser grained than density underflows, and may disrupt typical sedimentary sequences, especially near the source of slumping. On the other hand, density underflows are quasi-continuous, and are found only in topographic lows of the basin.

Rafted sediments in glaciolacustrine environments

Rafted sediment is material which is transported in a glaciolacustrine basin by ice, and is either melted out, or dropped from the ice. Material can be incorporated into ice actively by freezing, or passively by loading the surface of the ice. Passively loaded sediments are more easily lost from the ice surface, and most often form deposits of more than one particle (Gilbert, 1990). Actively loaded materials become concentrated on the ice surface by melting. Sediments can be deposited as a single particle (drop), agglomerations (dump), as frozen aggregates, or as melt out in lake bottom sediment laden ice (Gilbert, 1990). Passively loaded material will most often dump from the surface, whereas actively incorporated particles within floating ice will often rain out as single particles as the ice melts. In addition to sedimentation that can be attributed directly to the release of particles from floating ice, other effects occur. Initial calving, stirring by movement of icebergs, and melting of the iceberg affect lake circulation and mixing patterns (Smith and Ashley, 1985). Grounded icebergs may initiate turbidity flows (Powell, 1981; Rust, 1977), and scour the bottom (Dredge, 1982; Thomas, 1984) but this is less important than the transport of glacial sediments into other parts of the basin (Smith and Ashley, 1985). Table 2 presents characteristics of ice rafted sediment.

Ice-contact versus distal lakes

Ice-contact lakes, or supraglacial lakes have water that is in direct contact with glacial ice at some point, whereas distal lakes are separated from glacial ice. In general, ice-contact lakes are distinguished from distal lakes by rapid lateral and vertical facies changes (Smith and Ashley, 1985). Subaqueous outwash is the term given to material

Table 2: Characteristics of ice rafted sediments

Occurs as	drops, dumps, frozen aggregates, grounded ice contact deposits
Grain size	clay to boulder
Sorting	poor
Surface features	may be striated
Fabric	random (except frozen aggregates, or ice contact deposits)
Fauna	absent
Distance from ice contact	proximal to distal
Water depth	shallow to deep

(Modified from: Gilbert, 1990)

deposited from subglacial or englacial tunnels which meltwater exits, and consists of coarse water sorted sediment (Smith and Ashley, 1985). Other common sedimentary packages include flowtill, and finer grained lake sediments with abundant dropstones. Sedimentation in ice-contact lakes is dominated by underflows at the ice margin, and at a distance from the ice margin (Gustavson, 1975; Reimer, 1984).

Distal lakes tend to have finer grained and more laterally extensive deposits because of the separation of ice and water (Smith and Ashley, 1985). The lake is more prone to develop thermal stratification, which contributes to overflow and interflow as sediment dispersal mechanisms, along with underflows. Sand, silt and clay are moved along the bottom of these lakes by density underflows and turbidity events; and silt and clay are settled out from overflows and interflows (Smith and Ashley, 1985).

3.4 Glacial lake reconstruction and nomenclature

Glacial lake reconstruction has been based upon: 1) stratigraphy, 2) shoreline geomorphology, and 3) outlet controls (Muller and Prest, 1985). Reconstruction aims to define the glacial lake geometry and environmental conditions within the basin through time. Many glacial lake systems have had numerous periods of relative stability that are reflected within the sediments and geomorphology. The basis for naming such periods needs to be clarified, as systems such as Glacial Lake Agassiz were complex due to ice front fluctuations and numerous outlet shifts (Teller, 1985).

Stratigraphy, and to a lesser extent geomorphology is not an adequate basis for the naming of glacial lake events because of the influence of other factors in the system, such as isostatic adjustment, discontinuous deposits, differential sedimentation, and unclear stratigraphic relationships over basins (Farrand and Drexler, 1985; Larsen, 1985; Muller and Prest, 1985).

Muller and Prest (1985), argue that a glacial lake should be identified with only one outlet, as this is the only variable that is consistent through time in the basin with which a glacial lake can be identified. An example of this would be Glacial Lake Iroquois, which they identify as the glacial lake that drained through the Rome outlet to the Mohawk River. When this outlet was abandoned, Glacial Lake Iroquois ceased to exist. This convention is not widely used by other investigators.

Farrand and Drexler (1985) recognized that the terminology used in describing glacial lake systems was inadequate to correctly describe the relationships between glacial lake geometry and stratigraphy. They suggest that a lake level is an inferred water plane

within a basin that is defined by the morpho-sedimentary features of the basin in question. Lake levels are to be related to outlets if possible, but it is suggested that these be named independently, as one or more levels could be related to a single outlet, or vice versa. Lake levels must be inferred, as they may no longer be horizontal due to isostatic adjustment. Lake stage is not synonymous with lake level, as stage is a term that is restricted to chronostratigraphic usage by the convention of the North American Stratigraphic Code (North American Commission, 1983, article 74).

Lake phases are defined by Farrand and Drexler (1985), as the time where certain lake levels were functional. They emphasize the importance of lake phases as chronologic units. Lake phases are based upon the geometry and deposits of the lake, but should be named independently of lakes, outlets, and lake levels to avoid ambiguity (Clayton, 1983).

The wide range of glacial lake environments means that the system could range from a simple glacial lake with one easily identifiable level, to a complex system of interconnected basins, shifting ice fronts, and differential isostatic delevelling. The goal of unambiguously identifying lakes, lake levels, lake stages, and outlets requires different approaches based on the complexity of the system, and the quality, quantity and types of evidence that one can use to reconstruct the glacial lake system. In all cases, the investigator should define the development of the nomenclature, and outline the relationships of the various named lake attributes as completely as possible. In many cases, a shift in outlet may not warrant a change in the name of the lake in question if the lake continues to have similar discrete properties, and position as described in Glacial Lake Agassiz (Fisher and Smith, 1994), and Glacial Lake McConnell (Smith, 1994).

4.0 Methodology

4.1 Introduction

This was predominately a field based study, supplemented by map and airphoto interpretation, as well as lab work. Prior to entering the field, maps and airphotos were consulted to focus the field study but the main information presented in this thesis was obtained in the field. Lab work and analysis took place after the field season to refine the field results, and clarify the interpretation of the sedimentology and paleogeography of Glacial Lake Champagne.

The research process evolved as the study progressed.

- a) Strategy for glaciolacustrine sediment mapping: Initially, the nature of the geomorphology in part of the study area was investigated on airphotos and maps. The presence of lake sediments in the west part of the study area had already been attributed to Glacial Lake Champagne by Kindle (1953). The thick occurrences of glaciolacustrine sediments in the Whitehorse area were hypothesized to have been deposited also by Glacial Lake Champagne: this was investigated by mapping the extent of glaciolacustrine sediments in the area between Whitehorse, and the Dezadeash map area. The glaciolacustrine sediments and features present south of Whitehorse in the Southern Lakes System and were also hypothesized to be caused by the same lake: mapping was then undertaken in order to determine the extent of glaciolacustrine sediments between Whitehorse and the Southern Lakes System.
- b) Strategy for mapping of shoreline features: Reconnaissance fieldwork had indicated that

there appeared to be two stable stands of the lake, an indistinct one at 765 m, and a better developed one at 725 m. It was expected that stable still stands of the lake would be marked by features such as erosional or depositional shoreline remnants and by deltas in the tributary valleys. However, these features are often hard to identify in the field because of poor preservation. The mapping of such landscape features required a study of both the geometry, from air photos and field observation, and, where accessible, a study of the sediments. For example, the ascribing of the term shoreline was based on a number of factors, such as the nature of the sediments in the deposit, the morphology of the deposit, and the situation of the deposit. The geometry of Glacial Lake Champagne was then interpreted from the distribution of the data and the interpretation of the genetic origin of the units in three dimensions. Unfortunately shoreline features were remarkably scarce so that the lake geometry could not be mapped with as much confidence as expected and only the better developed lower stillstand was mapped.

c) Strategy for logging of sedimentary stratigraphy: It was assumed that the extensive sedimentary deposits would contain a record of sedimentary processes and that, given appropriately spaced sample sites, the sedimentary history could be elucidated by stratigraphic logging and facies analysis. Ideally sections would have been sampled representing a variety of situations both proximal and distal to presumed ice margins and shorelines. Sampling in the field situation is often limited in that sections can be sampled only where they exist and are accessible. Nevertheless a variety of sites giving reasonable geographic cover was found and logged. The stratigraphic logging was straightforward. The stratigraphic data had then to be interpreted by facies analysis to define the

depositional environments within the lake. The scarcity of certain features such as shorelines and deltas in itself provided information about the processes in the lake. The sum of the interpretations were synthesized into conclusions about the sedimentology and paleogeography of Glacial Lake Chamapagne.

4.2 Map and airphoto interpretation

Prior to, during, and after field work was conducted, map and airphotos were studied. The main purposes of map and airphoto interpretation were to define and map geomorphic surfaces, and to help in recognizing potential exposures suitable for logging. The basis for the recognition of geomorphic surfaces on maps and airphotos is primarily surface morphology of the surfaces. For example, raised glaciolacustrine sediments usually appear as flat, valley bottom terraces on airphotos and maps if the deposits are extensive enough (Goudie, 1978). One of the main limitations of map and airphoto based interpretation of geomorphology is that many surfaces representing vastly different depositional environments have similar morphologies, such as river terraces and lake terraces. Based on morphological attributes, but recognizing the limitations of map and airphoto interpretation, the geographic extent of the highest stand of Glacial Lake Champagne was mapped, initially on 1: 50 000 scale NTS sheets. This information was then transferred to 1: 250 000 scale NTS sheets, and then to a 1 : 1 000 000 scale NTS map sheet. A list of airphoto flight lines and photo numbers is given in Appendix A.

4.3 Field work

Field work was carried out in three stages, sometimes concurrently. The first stage was to evaluate the ease of access, and an initial reconnaissance of the sites identified in the airphoto and map interpretation, as well as identifying and evaluating other potential field sites. Site evaluation was based on the types of sediment present, the continuity of the deposits within the site, the height of the exposure, and the steepness, and degree of slumping in the site. Reconnaissance of sites included a preliminary stratigraphic log, and bulk sampling of sediments, so that a return visit would not be necessary in order to determine the depositional environment represented in the site.

The second stage of field analysis was detailed stratigraphic logging of the best sites in terms of the above variables. The majority of sedimentary units logged were glaciolacustrine sediments, but occurrences of till, and eolian sediments were also logged. Each site was photographed prior to any cleaning of the surface. Provisional sedimentary units were noted at this point. Units were identified by visual examination of properties of the unit such as sorting, colour, induration, structures and grain size. Changes in these properties were noted as a change in the sedimentary unit. After this the exposure was cleaned in vertical sections. The section was then re-evaluated for the distribution of sedimentary units, and sketched. Units were noted based on the variables listed above, but not in terms of the assumed mode of deposition. The next step was to measure the height of the exposure, and the thickness of the units within it. Grain size was estimated in the field. Sand and larger grain sizes were determined by a field microscope, or by measuring clasts. Bulk samples of at least 100 grams were taken from each distinct sedimentary unit

after it had been photographed. The average clast size within the unit was determined by measuring the b-axis (second longest axis of a three dimensional object) of a random sample of fifty clasts, and taking the mean. The roundness of these clasts was also noted by qualitatively comparing them to the Zingg roundness classification. In addition to this, the ten largest b-axes within the unit were determined. Other samples, such as tephra, lithologies, charcoal, and concretions were taken as appropriate. None of the samples contained materials that were suitable for dating.

The third stage of field work was to map the distribution of glaciolacustrine sediments within the basin. This was accomplished by travelling around the study area, and noting the occurrence of glaciolacustrine sediments, in terms of three dimensional coordinates. In order to be classified as glaciolacustrine sediment, the deposit had to be already exposed by natural or anthropogenic means, have a texture of silt or fine sand, and show no evidence of eolian origin, or reworking. The exposure of the deposit had to be at least 1 m high. In the case of continuous deposits of glaciolacustrine sediments observed in the Yukon and Takhini River valleys, the unit was first identified as glaciolacustrine, and then the complete exposure was followed by boat.

For all sites investigated in the field, the position was fixed in three dimensions by a hand held GPS unit, with an error of ± 30 m, and a pocket altimeter, with a 10 m scale.

4.4 Lab analysis of sediments

For each of the bulk samples of glaciolacustrine sediment, the percentages of clay, silt and sand were determined using a standard wet sieving and settling tube technique

(Torrance, pers. comm., 1997). This method is the easiest for dealing with small samples, and prevents the loss of any particles which may occur during dry sieving because of agitation leading to dust. The results (presented in Appendix B) were used to supplement the field analysis of sedimentary units.

4.5 Data analysis

The detailed stratigraphic logs were redrawn at a larger scale with more detail. At this point facies analyses of the units was done, based on the methods of Fritz and Moore (1988), Lewis and McConchie, (1994), and Boggs (1995). Grain sizes of the units were adjusted on the redrawn logs, as necessary, based on the lab analysis of grain size. The logs were redrawn a third and final time, after the interpretation of the facies and facies associations were complete.

Based on the field work results, the initial interpretation of the paleogeography was modified. The geometry of Glacial Lake Champagne was replotted on the maps by hand to produce the map of the geometry of Glacial Lake Champagne at its highest stand.

5.0 Sedimentology of deposits of Glacial Lake Champagne

5.1 Introduction

Of the many exposures of glaciolacustrine sediment of Glacial Lake Champagne, detailed stratigraphic logging was completed for six, at the sites shown in Figure 7 (p. 40). These six sites were chosen because of the height, and lateral extent of the exposure, distance from other sites in which detailed logging took place, and stratigraphic position. Other exposures were less acceptable for detailed stratigraphic logging due to homogeneity of sediments, slumping, steepness, or inaccessibility. At sites that were not logged in detail, the presence of glaciolacustrine sediments was noted, along with the three dimensional coordinates of the site. These sites will be presented in Chapter 6.

Much of the interpretation of the sedimentology of the sites in this section is justified by comparing the observed deposits with other glacial lake deposits reported in the literature. These comparisons are focussed on previously published studies that are likely to have had similar deglacial conditions compared to the study area. Although these similarities are pointed out, the initial interpretation of the sedimentology was based on the principles described in Fritz and Moore (1988), Lewis and McConchie, (1994), and Boggs (1995), and personal experience.

In the following pages, the stratigraphic logs for each of the six exposures are presented, along with a description of each log. The facies associations are then developed in order to comment specifically on the depositional environment at the site of each section, and generally on the depositional environment within the basin. Facies modelling

is a process of distillation where complex stratigraphic sequences or environments can be rationalized and comprehended (Walker, 1984). A distinct facies represents a mode of sediment deposition, and is recognized by grain size, sorting, structure, and position in stratigraphy.

5.2 Location and situation of stratigraphic logs

Table 3: Location and situation of sites of stratigraphic logs.

Name:	Location (UTM):	Elevation (top and bottom):	Exposed by:	Exposure faces:	Figure number
Whitehorse-Yukon River	08 496 810 E 6 733 830 N	671 m 649 m	Holocene river erosion	East	8
Watson River	08 507 500 E 6 699 020 N	760 m 749 m	Holocene river erosion	West	9
Tagish Lake	08 526 597 E 6 664 751 N	667 m 661 m	Holocene shore erosion	South	10
Whitehorse-Riverdale	08 499 096 E 6 730 369 N	720 m 687 m	Holocene river erosion	South-west	11
Marsh Lake Dam	08 517 201 E 6 716 025 N	709 m 702 m	Holocene river erosion	South	12
Millhaven Bay (Bennett Lake)	08 506 375 E 6 662 673 N	671 m 660 m	Holocene shore erosion	East	13

5.3 Stratigraphic logs

The provisional logs and sedimentary analysis completed in the field provided the basis for the stratigraphic logs for each of the above sites. These were modified after grain size analysis was completed for samples obtained from each of the units of the logs, and a facies model was developed. Table 4 provides a key for the facies associations for each of the logs.

Table 4: Facies associations key for Figures 8-13. (Modified from: Miall, 1977; Kaszyski, 1987; Kerr, 1987; Lewis and McConchie, 1994).

Symbol	Sediment type	Structure	Facies Code
clay	massive	CLm
silt	massive	SLm, SLm(e)
silt	weakly horizontally bedded	SLh
sand	massive	Sm, Sm(e)
sand	horizontally bedded	Sh, Sh(e)
sand	wavy bedded	Sw
sand	dipping beds	Sd(e)
sand	cross beds-trough	St
sand	cross beds-ripple	Sr
gravel	massive, matrix supported	Gms
gravel	horizontal beds, matrix supported	Ghs
diamicton	massive, matrix supported	Dmm
diamicton	deposited in lacustrine environment	Dmm(w)
colluvium	slumped glaciolacustrine sediments	X
tephra
charcoal
concretions
cobbles
boulders		

Note: on logs (Figures 8-13), each tick at the top of the stratigraphic column, from left to right corresponds to: clay, silt, fine sand, medium sand, coarse sand, gravel, and cobble, and the texture indicated corresponds to the mean grain size of the unit. The denotation (e) corresponds to an eolian unit.

csfmcgc
lissarb

I

Figure 8: Stratigraphic log of the Whitehorse-Yukon River site.

esf mege
liss sr b

Figure 9: Stratigraphic log of the Watson River site.

esf mege
liss sr b

White River Ash

SLm(e)

Sm

SLh

SLm

SL

SL

1 m

Figure 10: Stratigraphic log of the Tagish Lake site.

cs fmc gc
li ss srb

Figure 11: Stratigraphic log of the Whitehorse-Riverdale site.

es fm e gc
liss srb
SLm(c) with organics

White River Ash

Charcoal

Sm(c)

Gms

Chs

Sm

SLm

Chs

St

Figure 12: Stratigraphic log of the Marsh Lake Dam site.

Figure 13: Stratigraphic log of the Millhaven Bay site.

5.4 Periods of deposition

In the six sections, there are three periods of deposition. The first period of deposition is glacial, and occurred during ice cover. The only sediment associated with this period is till, recognized as being unsorted, unstratified diamicton. Till occurs only in the Millhaven Bay site.

The second period of deposition occurred during deglaciation, when the study area was inundated by meltwater forming Glacial Lake Champagne. There are a number of deposits associated with this period, described below. Glaciolacustrine sediment in the study area is the most common deposit, and is recognized primarily by its position in the regional stratigraphy; overlying till, and underlying eolian deposits, and also by the texture and structure it exhibits.

In the third period eolian deposits were developed. This took place in the Holocene. The genesis of the deposits is recognized as eolian because of the grain size and sorting; the presence of active eolian processes in places within the study area such as the Whitehorse-Riverdale site, and the Carcross Desert; the uniformity of these deposits across the study area suggesting a region-wide process; the presence of tephra interpreted as the younger White River ash that has been dated at 1200 BP (Clague, 1989) within the deposit; and the presence of charcoal layers and lenses within the deposit. These deposits were likely reworked from exposed glaciolacustrine sediments, which would have blanketed the region after the drainage of the lake.

5.5 Quaternary sedimentary facies

Clay facies

Facies CLm is a massive clay unit that occurs only as thin layers (5 cm) in the Watson River site. This unit is a dark grey olive clay, and has sharp contacts with the above and below units. Although this is a minor localized facies, it is included as a distinct facies because of the high clay content (76.28 %) it exhibits, and the depositional association of this unit with other facies, which will be developed later in the text.

Silt facies

In the stratigraphic logs, two distinct silt facies occur. Facies SLm is a massive silt unit, usually beige to buff in colour. This unit occurs in all of the stratigraphic logs except the Millhaven Bay site. The thickness of the unit is variable, ranging from 30 cm to 19 m. This unit does not exhibit dropstones in any exposures. SLm(e) is a massive eolian silt unit, approximately 1 m thick, that occurs at the top of the Tagish Lake site.

Facies SLh is similar to SLm in texture, colour, and occurrence, except that it exhibits weak horizontal bedding. Facies SLh and SLm are usually found together, and grade into each other, making exact contacts difficult to establish.

Sand facies

Five distinct sand facies occur in the six sites. Sand is the most common grain size that occurs in the stratigraphy. The sand is usually well sorted, and is predominately fine grained, but ranges to coarse grained.

Facies Sm is a massive sand unit, and is well sorted. This unit occurs in the Watson River site, the Whitehorse-Riverdale site, and the Marsh Lake Dam site. This unit occurs

as a continuous thin deposit, ranging from 10 cm to 1 m thick. There are no clast occurrences within the unit. Facies Sm(e) occurs in the Whitehorse-Yukon River site as an eolian cap of fine sand, that is devoid of any clasts or structures, and is well sorted.

Facies Sh is a horizontally laminated unit that occurs in the Marsh Lake Dam site, the Watson River site, the Millhaven Bay site, and the Whitehorse-Riverdale site. The texture of this unit ranges from fine sand to coarse sand. Clasts uncommonly occur in this unit associated with the coarser textures. This unit most often occurs as a thin (10 cm) lateral unit, but can range up to 1 m thick. This facies also occurs in conjunction with St and Sr, as described below. Facies Sh(e) is present only in the Whitehorse-Riverdale site, as a well sorted, medium texture sand unit 2 m thick. Facies Sd(e) is a well sorted, fine sand unit that occurs only in the Whitehorse-Riverdale site as gently (5°) dipping beds that were deposited by wind action.

Facies St is a fine to medium textured sand unit that is bedded in cross trough beds. This unit often occurs with facies Sr, ripple laminated cross beds, that has similar textural properties. Both of these units often occur together, and do not have a distinct contact, with both structures often occurring in close proximity in the stratigraphy (Figure 14, p. 53). The thickness of St/Sr units range from 50 cm to 10 m. These facies are found in every site except the Whitehorse-Yukon River site. In some of the units there are small clasts, occurring as stone lines, or as individual stones. These clasts average 2 cm b-axis, but range up to 7 cm. A rip-up of silt also occurs in these facies in the

Figure 14: Example of facies St and Sr in the Whitehorse-Riverdale site. Facies Sr is well developed immediately right of the pencil, and facies St is weakly developed immediately right of the pencil.

otherwise clean unit in the Whitehorse-Riverdale site. In general, structure in units that contain facies St and Sr has a progression from horizontal beds, to wavy beds, to cross trough beds and/or ripple laminated cross beds. These two facies may be deposited by the same processes, but a distinction is made between them because of the observed differences in bed morphology.

Gravel facies

There are two different gravel facies occurring in the Marsh Lake Dam site, and one in the Millhaven Bay site. Facies Gms is a massive, structureless, medium sand matrix supported gravel that occurs in both sites. Clast size averages about 2 cm, but can range up to 20 cm. This unit ranges from 50 cm to 2 m, and has a distinct contact with the above and below units. The gravel within the unit is moderately sorted.

Facies Ghs is a weakly horizontally bedded unit that occurs only in the Marsh Lake Dam site, and has a medium sand matrix. This unit exhibits strong normal grading from pea gravel to cobbles. This unit is 50 cm thick, and has a sharp contact with the above and below units.

Diamicton facies

There are two examples of diamicton, and both occur in the Millhaven Bay site. Facies Dmm is an unsorted, unstratified diamicton 4.5 m thick that occurs as a continuous unit, and has an indistinct unit contact with facies Dmm(r) that overlies it. Facies Dmm(r) is a diamicton that exhibits strong local sorting in the form of pea gravel pods, and sand lenses. This unit is 3.8 m thick, and has a sharp contact with the overlying unit.

Colluvium facies

There is one example of colluvium at the Watson River site. This unit is the top member, and is 4 m thick. The colluviated material is medium sand with occasional clasts ranging up to 10 cm b-axis. The structure of the unit varies from none to weak beds developed parallel to the slope. This is a poorly indurated unit that slumps readily if disturbed by digging. This unit is included because of its stratigraphic position and the texture it exhibits.

Other deposits

Tephra can be found in the eolian silts of the Marsh Lake Dam site, and the Whitehorse-Yukon River site. Based on the published distribution of tephtras in the Yukon (Clague, 1989), this is interpreted as younger White River tephra. It was deposited 1200 years BP, and covered southern Yukon, with its source being Mt. Bona, Alaska (Bostock, 1952; Lerbekmo and Campbell, 1969; Hughes *et al.*, 1972; Lerbekmo *et al.*, 1975). Charcoal is found above and below the White River tephra in the Marsh Lake Dam site. Concretions of calcite (as determined by reaction with dilute hydrochloric acid) were found in the Watson River site in facies SLh.

5.6 Local paleoenvironmental interpretations

For each site, facies associations are developed so that the local depositional environment can be defined. Regional depositional environments are developed subsequently in section 5.7.

Whitehorse-Yukon River site (Figure 8, p. 44)

At this site, 21.2 m of section was logged. There were only two recorded facies. The top unit of the section was facies Sm(e), and spanned 1.2 m. This unit is interpreted as being an eolian cap, reworked from the local glaciolacustrine sediments. The stability of the unit is uncertain, as there were no markers within it such as the younger White River tephra to provide any sort of control. This unit rests unconformably on top of the lower unit.

The remaining 20.1 m of exposure was logged as SLh and SLm, and is similar to many of the other glaciolacustrine silt deposits of the Canadian Cordillera, such as near Kamloops (Fulton, 1965), the Okanagan Valley (Shaw and Archer, 1979), and the Columbia Valley (Sawicki, 1990). This is one of many extensive outcrops of this sort of material exposed by the Yukon river in Whitehorse. Massive or weakly bedded silts (Figure 15, p. 56) are thought to be deposited in glaciolacustrine environments with no or minor seasonal variation in sediment supply (Sturm, 1978). This type of sediment yield regime may be explained by the release and entrainment of silt particles from glacial ice in contact with the body of water, and the subsequent redeposition of the silt. In ice contact lake basins receiving large amounts of coarse and fine grained sediments throughout the year, a seasonal cycle is either suppressed or cannot be discerned in the sediments (Hsu and Mackenzie, 1985; Eyles, 1987; Eyles and Clague, 1991). This sort of deposit is also found in the northeast Penticton section of the Okanagan Valley (Shaw and Archer, 1979), where it was interpreted as being formed near a stagnant ice body.

Because of the lack of variation within the section, the relationship of the

Figure 15: Massive to weakly bedded glaciolacustrine silts in the Whitehorse-Yukon River site. The vertical exposure is approximately 3.5 m high. This facies can form very steep slopes or cliffs, as shown here.

sediments to the ice front is not clear. This is further complicated by the lack of dropstones within the unit.

Watson River site (Figure 9, p. 45)

At this site 11 m of exposure were logged. The top 4 m consists of colluviated sand with clasts averaging 1 cm in b-axis. This 4 m unit appears to be a small slide scar. Because of this, the unit does not provide any structural information and the original deposit could not be excavated due to the instability of the sand. The sand itself has a medium texture, and numerous subangular to subround clasts. The unit may have been facies St or Sr unit before it slid. The clasts must have been within the unit before it slid.

Two depositional environments were represented in the remaining 7 m of undisturbed exposure. The top 1.3 m of the undisturbed portion of the exposure is interpreted as facies Sr and St. Ripple drift cross laminations originate from density underflows of sediment laden meltwater flowing into glacial lakes (Jopling and Walker, 1968), and are common in many glacial lakes (Gustavson, 1975). Underflow integrity is diminished by mixing with ambient water, sediment deposition, gentle lake floor slope, and bottom friction (Smith and Ashley, 1985). The remaining 5.7 m of exposure is a cyclical progression of SLh and Sh, with interspersed units of Sm and CLm, with some of concretions.

The structure of the units suggests that the entire 7 m of exposure was deposited in a proglacial environment, in an area that was relatively close to active and stagnant ice. Ice proximal deposition is reflected in the stratigraphy as high energy underflows and slump deposits, interspersed with low energy suspended sediment deposition (Smith and

Ashley, 1985) which is consistent with the stratigraphy here. Isolation from source and subsequent stagnation is an important element in the development of the deglacial environment of the Canadian Cordillera (Fulton, 1965; Shaw 1979; Shaw and Archer, 1979; Kaszyski, 1987). The sharp contact between facies Sr and the lower units suggests that Sr is a single density underflow event. The position of the density underflow in relation to the former ice front is unclear, as these events may initiate on an unstable surface of a delta in the basin, but not come to rest (and deposit materials) until the event is well out of the zone of instability (Gilbert, 1975). Underflows in modern lakes have been detected as far as 40 km from the stream mouth (Pickrill and Irwin, 1982), and are known to transport fine sand 15 km in Lake Geneva (Houbolt and Jonker, 1968) and 6 km in Lillooet Lake (Gilbert, 1975). There are no deltas that are associated with Glacial Lake Champagne in the area close to the site. The lower 5.7 m of the in-situ exposure is similar to parts of the northeast Penticton section (Shaw and Archer, 1979), in the cyclical manner in which weakly bedded silt and sand are deposited, with occasional units of massive sand and massive clay, described as unctuous clay by Shaw and Archer. The Watson River section does display a greater amount of weakly bedded sand. Shaw and Archer (1979), interpret the clay units as being deposited out of suspension in winter, and the sand unit as being deposited by density underflows that were initiated in winter. The weakly bedded silt represents summer sedimentation from overflows, interflows, and underflows (Shaw and Archer, 1979). This provides an acceptable framework for the interpretation of the depositional environment in the Watson River section.

It is important to note that the top of the Watson river exposure is only 5 m below

the highest observed stand of Glacial Lake Champagne in the area. The top colluviated and in situ sand could be interpreted as a deltaic deposit, but this hypothesis is rejected for two reasons. First, the deposit is too well sorted and thin to be a deltaic deposit. Second, the surrounding topography is flat bottomed valley, which is unlikely to have drained in the present direction. Wheeler (1961), contends that the Watson River previously drained northward into the Yukon River Valley, instead of the south, Southern Lakes drainage of present.

Tagish Lake site (Figure 10, p. 46)

Of the 6.5 m of exposure in this site, the top 70 cm was deposited in the Holocene by eolian processes, and the remainder was deposited as glaciolacustrine sediment in Glacial Lake Champagne. The eolian cap is facies SLm(e) and has one of the two occurrences of younger White River tephra. This unit lies unconformably on top of the glaciolacustrine sediments, and is likely reworked glaciolacustrine sediments.

There are three facies present in the lower 5.8 m. Facies Sm does exhibit some local bedding, and also rare clasts, which average 1 cm across the b-axis, and are usually subangular. This unit has an indistinct boundary with facies SLm with an occurrence of SLh that is exposed for the next 2.3 m. The bottom unit is 2.5 m of facies St, with a texture of fine sand with some silt.

Facies Sm and facies SLm are common glaciolacustrine sediments, that were deposited from overflows or interflows of sediment laden water (Leeder, 1982). The clasts present in Sm were likely dropped from floating ice, as single units. The other liberated fine sediment was settled out slowly as particles. These particles were therefore deposited

in the same manner as the sediment present in the overflows or interflows (Sturm, 1978).

The bottom unit consists of facies St. This unit is 2.4 m thick, and is fine sand, with some silt. Trough cross beds are not common in glaciolacustrine environments, but have been noted in the Bon Accord section, near Edmonton (Shaw, 1975). In the Tagish Lake site, this facies is associated with ripple drift cross laminated sands, and planar bed sands. This suggests that this unit could have been deposited from a density underflow. A transition from larger wavelengths of 15 cm to smaller wavelengths of 5 cm from bottom to top suggests that the unit may have been deposited by a single event, as decreasing wavelength is a form of grading (Walker, 1967).

Whitehorse-Riverdale site (Figure 11, p. 47)

This is the largest exposure logged (Figure 16, p. 62), consisting of 4.8 m of eolian sand overlying 28 m of glaciolacustrine sediments. The eolian unit consists of facies Sm(e), Sd(e), and Sh(e). The entire unit is well sorted fine to medium sand. The unit was reworked from local glaciolacustrine sediments.

The remaining 28 m of exposure consists of facies SLh, SLm, Sh, Sm, St, Sw and Sr. There are two depositional environments defined by these facies. The finer grain sizes are consistent with deposition from sediment laden overflow and interflow plumes. The sand units were likely deposited by density underflows of various magnitudes. The smallest unit deposited by a density underflow in this site is 13 cm, and the largest is 7.3 m. The structure within these units tends to be an unclear progression from planar bedded sands to ripple drift cross laminated sands to trough cross bedded sands. The lack of clear progressions from one structure to another, especially in the larger units may be due to

Figure 16: The Whitehorse-Riverdale section. The eolian portion of this section lies above the vegetation near the top of the unit, and is still actively eroded and deposited. The remainder of the sediments are glaciolacustrine, and were deposited by suspended sediment settling, or as density underflow deposits. Note the person at the bottom of the section for scale.

multiple events, reworking previously deposited density underflow deposits (Walker, 1967). This interpretation is supported by the magnitude of the larger deposits in the exposure. The single unit of massive sand is also interpreted as being a reworked density underflow deposit that resulted from a subaqueous slump. This interpretation is supported by the presence of a sharp, but wavy contact with the lower unit. The largest density underflow deposit contains a rip up clast of silty clay near the base of the deposit. Rip up clasts are recognized as being associated with density underflow deposits (Gustavson *et al.*, 1975).

The frequency and magnitude of density underflow deposits in this site suggests that it was relatively close to a delta that built into Glacial Lake Champagne, or that the underflows were introduced into the basin directly from an ice front. One source of density underflows is recognized as being oversteepened, unstable, rapidly prograding delta fronts (Pharo and Carmack, 1979). There are no deltas visible in the area near this site, meaning that this explanation is less likely than input directly from a nearby ice front in direct contact with lake water.

Marsh Lake Dam site (Figure 12, p. 48)

At this site, an eolian cap of massive fine sand unconformably overlies glaciolacustrine sediments. The eolian unit consists of facies Sm(e), and is reworked glaciolacustrine sediments. The deposition of the eolian sediments seems to have occurred in the Holocene as intermittent eolian activation, as explained in chapter 7. This is based on the layers of charcoal, and the position of the younger White River tephra. The younger White River tephra occurs at 40 cm from the top of the deposit, which is 1.77 m

thick.

The glaciolacustrine portion of the section is divided into three units of facies Gms, and Ghs; between and below these units the rhythmically deposited facies SLm, and Sm; and the bottom unit facies Sr and St.

The presence of gravel in this section suggests that the site was very close to the ice front. The top unit of gravel has within it silt pods and normally graded pea gravel to coarse sand lenses. The second unit of gravel is weakly horizontally laminated, but strongly normally graded, which is consistent with subaqueous slumping and transport via turbidity events. All three units of gravel are interpreted as ice contact, or near ice deposits, consistent with those described by Kaszyski (1987), in the shield terrain of southern Ontario, near Haliburton. The gravel layers are continuous over tens of metres in this site, which along with the small scale sorting within the massive gravels and the weak laminations and strong grading rule out ice rafting as the sediment source. The gravel was likely deposited as high density, high energy, proximal turbulent underflows (turbidity event) (Rust and Romanelli, 1975). This suggests that the ice front was close during deposition of this unit.

Interspersed between, and beneath the massive gravels are rhythmically deposited silts and fine sands. These units were deposited from suspended sediment laden plumes. The silt-fine sand rhythmicity could be a seasonal variation (Smith, 1981), or another time scale of rhythmic deposition. Smith and Ashley (1985) describe surge rhythmite that are the product of slump generated surge currents. They are noted as having a fining up distribution of grain size, as opposed to a seasonal rhythmite (varve), which has no fining

up trend. The layers in this site do not fine up, which suggests that they were created in a overflow-interflow regime (Smith and Ashley, 1985).

The bottom unit consists of facies St and Sr. None of the structures were very well defined, and the unit itself was poorly sorted, with isolated clasts, stone lines, and units of pea gravel present. This unit is interpreted as a density underflow deposit, but more distal than the massive gravel units. The bottom of the unit was buried, so it was impossible to determine if this deposit represented a single, or multiple events. The position of ice may have changed in the period between the deposition of this unit, and the massive gravel units, but more likely there was a shift in englacial drainage, leading to a different relative position of water input, not of the ice front (Kaszyski, 1987).

Millhaven Bay site (Figure 13, p. 49)

This exposure consists of two units; a glaciolacustrine unit overlying till. This is the only site where till is exposed. There are three units in the glaciolacustrine sediments that represent different depositional environments. The uppermost unit is largely facies Sh with poorly developed facies St and Sr. This is similar to the bottom unit of the Marsh Lake Dam site that was previously described in terms of sorting and texture, but the cross bedding is less well developed. This deposit represents a density underflow event, based on the thickness of the unit, layer morphology, and the presence of clasts. The top of the deposit is truncated, so it is impossible to tell if this represents a single, or multiple events.

Near the bottom of the unit, there is a pod of coarse gravel to cobbles of facies Gms. The origin of this unit is unclear, with two modes of deposition being possible. The pod could have been a large dump of frozen particles that was ice dropped. Another

equally acceptable hypothesis is that it represents a subaqueous mass movement. In either case, it is a high energy deposit.

The bottom glaciolacustrine unit is a diamicton that exhibits local bedding of pea gravel and sand, and has occasional cross beds. This unit is interpreted as a flowtill, that was deposited subaqueously at the ice front. This type of deposit was recognized by Shaw and Archer (1979), at the northeast Penticton section. This unit is not interpreted as reworked till because of the sharp contact between the flowtill and till.

5.7 Regional depositional environments in Glacial Lake Champagne

There are two major modes of depositional processes that led to the sedimentary record in sites from Glacial Lake Champagne. The first major mode is the deposition of silts and fine sands from sediment laden overflow and interflow water entering the basin from subaerial, supraglacial or englacial sources. This sediment is either massive or weakly laminated, with laminations becoming more common as median grain size increases. Laminations seem to be a reflection of seasonal variations. The second major mode is the deposition of sediment entrained from existing bottom deposits, or discharged from subareal, supraglacial, englacial, or subglacial sources by density underflows or turbidity currents that are more dense than the surrounding water. This sediment ranges from fine sand to cobbles. The variation is likely due to the proximity of the deposit to the point of initiation, and the energy of the event. These deposits are manifested as massive, planar bedded, or cross bedded sands; or massive graded gravels; or as a combination of these. The sand packages were deposited in a more distal environment, suggesting deposition

from density underflows (Eyles *et al.*, 1987) whereas gravel deposits were deposited in a proximal environment via subaqueous slumping or high energy turbidity flows (Kaszyski, 1987). Minor modes of deposition include melting of subglacial ice yielding flowtill, and sediment drops and dumps by floating or lodged glacial ice.

Variation in the sedimentary packages of the six investigated sites is large. This suggests that the deglacial environment was largely defined by the presence of stagnant or unconnected lobes of ice, similar to environments described by Fulton (1965), Shaw and Archer (1979), and Kaszyski (1987). Rapid retreat of glacial ice with the onset of the termination of the late Wisconsinan has been noted in other parts of the Canadian Cordillera, such as northern Vancouver Island (Clague, 1980), the Lower Mainland of British Columbia (Saunders *et al.*, 1987), southern Yukon Territory (Jackson *et al.*, 1991), northern British Columbia (Ryder and Maynard, 1991) and southeast Alaska (Mann and Hamilton, 1995). This has important implications for the reconstruction of the paleogeography of Glacial Lake Champagne.

6.0 Paleogeography of Glacial Lake Champagne

6.1 Introduction

The reconstruction of glacial lakes is important in the interpretation of the regional deglacial evolution of an area, as a lake stand may define ice fronts within the region for that particular stand, and as the presence of ice is directly or indirectly responsible for the existence of the lake. Reconstructions are based on the distribution of morphological features interpreted to be associated with the lake stand in question, and on the analysis of the associated sediment.

The original aim of the research was to develop lines of evidence to establish the paleogeography of the two most conspicuous stands of Glacial Lake Champagne that are evident throughout the study area. However, in view of the paucity of field evidence, only the better developed stand can be discussed at all and the conclusions reached can at best be considered to be only tentative. Various provisional hypotheses are discussed below but it must be noted that inferences are at present only poorly supported.

The types of evidence that were investigated and will be discussed are:

1. Presence and nature of glaciolacustrine sediment
2. Presence of raised deltas
3. Presence of paleoshoreline features

The discussion of the paleogeography will include a hypothesis on the blockage mechanism and location, and the provisional chronology of the evolution of the lake basin. A reconstruction of the geometry of Glacial Lake Champagne at the best developed stand

of 765 m will be presented in the form of a map.

6.2 Stratigraphic and geomorphic evidence: Glacial Lake Champagne

Distribution of glaciolacustrine sediments in the study area

Glaciolacustrine sediments in the study area are mostly sands, silts, and silty clays, and blanket the floors of valleys. Exposures of sediment are usually only seen where there is active valley bottom erosion, such as fluvial action, or in sites that have been exposed in road cuts, or other development. The presence of glaciolacustrine sediments in areas that do not have exposures can be interpreted through airphoto interpretation (St-Onge, 1980). Areas that are blanketed by glaciolacustrine sediment often exhibit low relief, and in the study area may have been reworked by strong southerly winds into eolian landforms (Hughes, 1990).

Glaciolacustrine sediment is commonly observed in road cuts and natural erosion surfaces at elevations below approximately 765 m in the study area, and less frequently above 765 m. Glaciolacustrine sediments below 765 m are considered to be deposited by Glacial Lake Champagne and constrain its extent. The geographical extent of glaciolacustrine deposits was determined by noting the three dimensional position. Figure 17 (p. 70) presents a map of the distribution of glaciolacustrine deposits associated with Glacial Lake Champagne. Glaciolacustrine sediments are more extensive than what is mapped in Figure 17, but are not exposed. One of the aims of mapping the occurrence of glaciolacustrine sediments was to establish a linkage between the previously identified sediments of Glacial Lake Champagne, as interpreted by Kindle (1953), the lacustrine

Figure 18: Large section of glaciolacustrine sediments exposed by the Takhini River. The section is actively eroding, as shown by the debris of the exposure. This type of an exposure is typical of those observed in the Takhini River and Yukon River valleys in terms of its steepness, height, and slumping.

sediments of the Whitehorse area (Wheeler, 1961), and Glacial Lake Carcross (Wheeler, 1961; Hart and Radloff, 1990). None of the previous documentation of glaciolacustrine sediment was part of a systematic mapping exercise in order to establish the paleogeography of Glacial Lake Champagne, but was included as part of a comprehensive study of the geology in the region (Kindle, 1953; Wheeler, 1961), or as background to a regional geologic mapping exercise (Hart and Radloff, 1990). A nearly continuous occurrence of these sediments was observed and mapped in the valley bottoms of the Takhini River, and the Yukon River; and many occurrences were observed and mapped in the Southern Lakes district, and along the roads of the study area.

Deltas associated with Glacial Lake Champagne

Raised deltas often appear on maps or airphotos as level perched surfaces next to, or incised by tributaries of larger streams or lakes (Smith, 1992). In the case of Glacial Lake Champagne, potential sites would occur along creeks entering the Southern Lakes system or tributaries of the Yukon River, the Takhini River, or the Dezadeash River. Sites were identified on airphotos or maps, and were subsequently investigated in the field if practical. Of all the sites investigated, only two were identified as deltas of Glacial Lake Champagne. These are noted on Figure 17 (p. 70).

As relict sedimentary feature of glacial lakes, deltas are the most reliable indicator of lake stage, as their geometry is dependent on the lake level at the time of deposition, but they do not necessarily mark the highest stand of the lake (Smith, 1992). The delta at Lime Creek, (Figure 19, p. 73) which is located at UTM coordinates 08 527 073 E, 6 691 233 N, has two levels. The top level is interpreted as forming first, at an elevation of 765

Figure 19: The delta at Lime Creek. View is looking up to 765 m level from the 725 m level. Note the lack of mature trees on the site. This is likely related to the rapid infiltration of water into the coarse deltaic sediments.

m. This level was subsequently incised into as the second delta level built at an elevation of 725 m. Surface pits dug into the deposit at both levels verified that the dominant grain size was gravel, with an average b-axis of 2.5 cm. There were several large boulders present on the surface of the delta, with the largest showing a b-axis measurement of 165 cm. There was no exposure of the deposit, as presently, there is no surface stream flowing through or around it. Lime Creek enters a small pond about 1 km west of the deposit, which has no surface drainage. The pond itself showed no evidence of shoreline fluctuation, as vegetation was well established to the shore. There is a dry, relict channel that is overgrown that cuts to the north of the Lime Creek Delta, but offers no exposure of the sediments. The presence of Lime Creek, and the terminal pond, along with the fact that there is no surficial drainage, and hasn't been for some time suggests that water is draining through the Lime Creek Delta, which further suggests that the internal structure is gravels, and/or sands.

The second exposed delta is along the Watson River at UTM coordinates 08 512 675 E, 6 679 646 N (Figure 20, p. 75). The top of the delta is at 731 m, and the exposed thickness of the unit is approximately 20 m. This delta appears to have prograded into Glacial Lake Champagne, which would have inundated the valley to the south. This delta also has large clasts, on the order of 50 cm b-axis. Based on the presence of crude dipping beds the unit is interpreted to be a crude foreset sequence, with the beds dipping south at an angle of approximately 30°. This section was too steep and unstable to be sampled.

The lack of deltas is important given the presence of large amounts of coarse grained material in the stratigraphy. Either deltas did not form in Glacial Lake Champagne,

Figure 20: The delta at Watson River. View is to the east. The lower section of the delta was exposed by undercutting of the Watson River, and the upper portion was exposed by the construction of the White Pass Railway, which is visible.

or they did form, and have been removed in the Holocene by erosion, or have been obscured in the record by subsequent eolian deposition. Holocene erosion was noted by Smith (1992) as a main explanation for the lack of deltas associated with Glacial Lake Mackenzie. Wheeler (1961) hypothesised that the Watson River shifted from Yukon River (north) drainage to Bennett Lake (south) drainage at the end of the last glaciation. Any deltas in this area may have been obscured by Holocene eolian activity. The delta at Lime Creek still exists because of the relatively small discharge generated within the catchment above being transmitted through the sediments, limiting erosion. The delta at Watson River has been exposed through fluvial action. The Watson River at this site is now eroded into bedrock, limiting further erosion of the deltaic sediments. Both of these situations represent uncommon fluvial conditions.

Holocene erosion or deposition may have removed or obscured many of the deltas associated with Glacial Lake Champagne, as described above. The lack of deltas is better explained by deltas not being formed in many tributary valleys abutting Glacial Lake Champagne. This suggests that stagnant ice filled many of the tributary valley bottoms. This explanation is supported by the sedimentology exhibited and described earlier. Further support for this hypothesis is the lack of valley train deposits that would have been deposited by subaerial streams flowing from the ice front to Glacial Lake Champagne.

Paleoshoreline features of Glacial Lake Champagne

Several investigators (Kindle, 1953; Wheeler, 1961; Morison and Klassen, 1991; Hart and Radloff, 1990), have described the presence of wave cut terraces or beach deposits in the study area, at various elevations (Figure 21, p. 77) between 701 m and 854

Figure 21: Thermokarst and shorelines in the Takhini Valley. View is to the west. A thermokarst depression is just visible in the bottom left of the photograph. Shorelines associated with Glacial Lake Champagne are visible near the bottom of the slope on the right of the photograph as two well developed benches.

m. These features were re-evaluated on airphotos and in the field, if accessible. Not all of the paleoshoreline features are interpreted to be associated with the two stands of Glacial Lake Champagne that are being investigated because of the discrepancies in elevations. These may be associated with earlier phases of Glacial Lake Champagne, or with another event. Figure 17 (p. 70) shows the distribution of paleoshoreline features associated with Glacial Lake Champagne, based on field re-evaluation.

Shorelines that were directly investigated were limited to the Whitehorse map sheet (105 D) because of time constraints. Those of Kindle (1953) were observed from the Alaska Highway, and noted. None of the shorelines had exposures, so the material in them was evaluated by digging pits. Nearly every shoreline showed well sorted beach gravels. It was impossible to note the structure of the deposits, because of the lack of exposures.

6.3 Re-evaluation of the previously published paleogeography of Glacial Lake Champagne

Western portion: map sheet 115 A

Glacial Lake Champagne was initially identified in the Dezadeash map area by Kindle (1953), as part of a Geological Survey of Canada reconnaissance survey. He recognized the large extent of lake features in the area, and named the lake which caused the features Glacial Lake Champagne, after the Village of Champagne, the oldest post within the perimeter of the basin. A map of the shorelines of Glacial Lake Champagne was produced based on airphoto interpretation. Most of the well defined beaches occur on steep slopes between 701 m (2300 ft) and 854 m (2800 ft), but are not individually

identified. Kindle postulated that the beaches are composed of sand and gravel, and are concentrated in areas of pronounced wave action. Large areas of silt were noted in Dezadeash, Takhini, Kathleen, and Aishihik River valleys, with exposures of up to 61 m (200 ft) in height. Ice was thought to have existed in the Mush Lake, Klukshu River, the west half of Kathleen Lake, and the Takhanne River valleys, which would have bounded the southward transgression of Glacial Lake Champagne.

Fieldwork in this portion of the study area focussed on checking the findings of Kindle. For the most part, much of the previously published history was re-interpreted in the same way, but there were some important exceptions. None of the large exposures described by Kindle (1953), or by Johnson and Raup (1974), were found, despite a search. Many road cuts and exposures only up to 3 m in height along the Takhini and Dezadeash River valleys show glaciolacustrine sediments. All of these sites showed massive or weakly laminated fine sand or silt. It may be that the exposures noted by Kindle may be in an extremely remote location, as there is no location for the section given.

Eastern portion: map area 105 D

Wheeler (1961), noted a number of occurrences of glaciolacustrine sediments. Glacial Lake Carcross was described as occupying the lower reaches of the Watson River, Wheaton River, and Tagish Lake, bounded by ice dams that were hypothesized in the Lewes Lake area, Bennett Lake, Windy Arm, and Taku Arm. Silt deposits in the Takhini River valley were interpreted to be part of Glacial Lake Champagne. The silts of the Yukon River, Takhini River, and M'Clintock River valleys near Whitehorse, were assumed to be associated with a stagnant ice complex. Another proglacial lake was

indicated in the lower M'Clintock Valley bounded by ice in the Michie Creek valley, the M'Clintock Lakes and an ice tongue near the mouth of the M'Clintock River.

Based on field investigation and mapping, the sediments described by Wheeler in the above section have been re-interpreted as sediments of Glacial Lake Champagne. There is no geomorphic evidence that there was any blockage within the main valleys that could have split the basin in which Glacial Lake Champagne existed. The lake is interpreted as occupying the valleys of the Takhini, Yukon, Six Mile, M'Clintock, Watson, and Wheaton rivers, and the basins of Marsh, Tagish, Bennett, and Nares lakes, up to a maximum elevation of 765 m.

6.4 Geometry of the ice front-Glacial Lake Champagne interface

Ice front positions described by Kindle (1953) were investigated, by a vehicle based reconnaissance on the Haines highway. Glaciolacustrine sediments appear to end in the areas described as being ice covered at the same time as Glacial Lake Champagne existed. One important ice position that is not investigated by Kindle is the position of the ice front in the Alsek River valley. The toes of glaciers of the Kluane region today are retracted from valley bottoms. The degree of glaciation in the region has fluctuated significantly during the Holocene, with modern glaciers reaching from their tributaries into the axes of main valleys. Neoglacial Lake Alsek was a large Holocene glacial lake that formed in the Alsek River valley due to the damming of the Alsek River by advances of the Lowell Glacier (Clague and Rampton, 1982). This lake is interpreted to have filled and emptied many times. The most recent transgression into the Dezadeash River valley

occurred between AD 1848 and 1891 (Clague and Rampton, 1982). Schmok and Clarke (1989) were unable to determine the exact chronology and evolution of the various phases of Neoglacial Lake Alsek, but stated that there were a number of Neoglacial Lake Alsek phases, and that some phases had multiple fillings and drainings on a variety of time scales.

The drainage of Kluane Lake is also interpreted to have shifted during the late Holocene due to the advance of the Kaskawulsh Glacier, leading to a drainage reversal of the Slims River, forcing Kluane Lake drainage to the north into the Yukon River system (Bostock, 1969). The fact that Holocene glacial limits have reached low enough elevations to cause ice dammed lakes, and drainage reversals means that during the Last Glacial Maximum the Kluane ice complex formed an effective ice margin at some point in the Dezadeash River valley. Inferred ice margins are shown in Figure 22 (p. 82).

The Holocene glacial history in the west half of the Dezadeash map area complicates the interpretation of shoreline associations. Any glaciolacustrine feature found above 700 m must be associated with a Pleistocene glacial lake event because of the regional geomorphology. There is a drainage divide about two kilometres east of Champagne. Dezadeash River flows west of this divide into Pacific draining Alsek River, and the Takhini River flows east into the Yukon River system. Any Neoglacial lake in the region would be bounded by the glacial ice in the Alsek River valley, the natural rise of the landscape to the northwest to 781 m at Kluane Lake, and this low divide. If a Neoglacial lake overtopped the 700 m level, it would have left some erosional feature at this divide as it cut through the unconsolidated sediments. If these sediments were deep enough, there may have been a reversal in drainage.

Ice front positions in the Southern Lakes system are less clear. Ice in this region trended northerly, with the source areas being the Boundary Ranges of British Columbia and Alaska (Wheeler, 1961; Hughes *et al.*, 1968). Ice positions are inferred from the local lack of glaciolacustrine sediments or landforms. Based on this, it is assumed that ice was present in the valleys and basins to just past the Yukon-British Columbia border, as shown in Figure 22 (p. 82).

The extent of Glacial Lake Champagne towards the east, into the Atlin Lake region, is also unclear. This position would also be dependent on the position of the ice front, as there is no bounding topography. Airphoto and field investigation were not able to delineate any glaciolacustrine features east of Marsh Lake, but also do not show any definitive evidence of a stagnant ice body in the lower valleys. The area around Little Atlin Lake is hummocky and pitted, and is a local drainage divide, which may indicate the presence of a stagnant ice body (Kaszyski, 1987).

In the northwest of the study area, the relief is high enough that the 765 m stand of Glacial Lake Champagne would have not overtopped any drainage divides. This is true for the northeast of the study area, except at Lake Laberge. Ice must have existed in this basin during the various stands of Glacial Lake Champagne. A lobe of ice originating in the Cassiar district trended northwest up the strike of the Teslin Valley (Hughes *et al.*, 1968; Jackson *et al.*, 1991) and was the likely source of the blockage as it overran the Laberge basin. Due to the difficulty of accessing this area, no field checking of this hypothesis took place.

The apparent lack of outwash deposits throughout the study area suggests that

Glacial Lake Champagne was in direct contact with most of the remaining valley bottom ice, which is consistent with the description of other glacial lake systems in the Canadian Cordillera (Fulton, 1965; Shaw and Archer, 1979; Sawicki, 1990). This hypothesis is supported by the interpretation of the depositional environments in the Whitehorse-Riverdale site, the Marsh Lake Dam site, and the Millhaven Bay site. The landscape surrounding Whitehorse is kettled, with many lakes up to 2 km² occupying basins up to 30 m deep, suggesting that a large amount of fragmented stagnant ice was present during the existence of Glacial Lake Champagne. Ice-free land was in contact with the lake where shorelines or deltas exist. Therefore, a conceptual model of Glacial Lake Champagne at 765 m would be ice bounding the lake in valley bottoms mostly as stagnant ice, and uplands being ice free, where shorelines formed.

6.5 Geometry of Glacial Lake Champagne

Basic measurements

The highest stand of Glacial Lake Champagne investigated was 765 m. This height is based on the heights of the upper surface of the Lime Creek delta, shorelines in the Takhini Valley, and shorelines near Carcross (Figure 17). At this stand, Glacial Lake Champagne had an area of approximately 2425 km², and a long axis of 260 km. The maximum depth of the lake was at least 159 m. The geometry is shown in Figure 23 (p. 85).

Nomenclature

It is suggested that the name Glacial Lake Champagne be used to identify all lake stands previously defined by Kindle (1953), or in this work. The name Glacial Lake Carcross should be discontinued as this refers to a contiguous portion of Glacial Lake Champagne, as described in this work.

Outlet

There were two possible outlets for Glacial Lake Champagne. One was south through the Alsek River system to the Pacific. The other was north, through the Yukon River system. These are the only two major drainages that the lake straddled. There has been no direct evidence for the occupation of either outlet, but there is strong indirect evidence that rules out the Alsek River outlet. It was earlier shown that the glaciers in the Klumane area occupy low enough terminal positions to have caused major shifts in the regional hydrology. This suggests that coalescent ice from these glaciers would have easily extended far enough out into the Dezadeash Valley to effectively block Glacial Lake Champagne for the early part of its existence. Drainage through the Alsek River system would also suggest that more of the present drainage in the area would have been captured by the Alsek, especially if the drainage of Glacial Lake Champagne was catastrophic, leading to valley bottom erosion. For these reasons, it is hypothesized that drainage of Glacial Lake Champagne was north through the Yukon River system, but drainage through the Alsek River system cannot be ruled out, especially if the lake existed late in the deglacial period. In order for this to be resolved, geomorphology consistent with lake drainage must be found in either the Alsek River system, or the Yukon River

system. This may be problematic in the Alsek River system, as major floods associated with the drainage of Neoglacial Lake Alsek occurred a number of times in the Holocene (Schmok and Clarke, 1989) which may have eradicated or subdued evidence for an earlier flood.

If the lake drained through the Laberge area, drainage would have been slow and non-erosive, possibly as a kame system between stagnant ice and the valley wall in the Laberge basin, as no landforms were found in this area that would be consistent with a catastrophic failure of an ice dam. The problem of the outlet of Glacial Lake Champagne warrants further work.

Chronologic control of Glacial Lake Champagne

The chronology of deglaciation in southern Yukon Territory is poorly understood. In order for Glacial Lake Champagne to have existed, ice blockages must have remained viable in the Kluane region, to prevent drainage into the Alsek River system, and in the Lake Laberge region, to prevent drainage into the Yukon River system. The sediments of Glacial Lake Champagne did not yield any biological material that would have been useful for radiocarbon dating, so a direct date is not available. There is some constraint on the chronology of deglaciation in the Kluane region. Deglaciation began on the east side of the St. Elias Mountains (Kluane region) before $13\,660 \pm 460$ BP based on radiocarbon dating (Hughes *et al.*, 1989), and the Kaskawulsh Glacier was less extensive than present by 9780 ± 180 BP based on radiocarbon dating (Hughes *et al.*, 1989). Based on these dates, it is likely that Glacial Lake Champagne existed sometime 12 500 BP, and 10 500 BP, but this is largely speculation.

The length of time required to deposit large quantities of glaciolacustrine sediments is relatively short. The glacial lake that occupied the Thompson Valley of central British Columbia deposited similar quantities of sediments in 100 to 200 years of existence (Fulton, 1965). This can only be used as an indication of the length of time that is needed for similar amounts of sediment to accumulate in a similar basin under similar environmental conditions, and not as a direct analog for sedimentation rates.

7.0 Holocene geomorphic evolution in Glacial Lake Champagne deposits

7.1 Introduction

Although this thesis is focussed on the evolution of Glacial Lake Champagne, the distribution and type of sediment it has left has largely determined the Holocene evolution of the landscape it covered. This chapter will briefly describe some of the landforms that were noted during the field season.

7.2 Eolian features and impact on the landscape

Much of the surficial deposits in the study area are eolian features. The abundance of silt and fine sand that was deposited by Glacial Lake Champagne, and the strong southerly prevailing wind are the two factors that allowed the movement of sediment by wind. Klassen (1987) noted deactivated dunes in the northwest of the study area. At present, there are some areas that are undergoing active eolian processes, such as the Carcross Desert, which has probably become reactivated because of human recreational use. Grain size analysis of sediment from the Carcross Desert shows that there is no major difference in the texture of this eolian feature, and sediments of Glacial Lake Champagne.

An interesting sequence of sediments was noted at the Marsh Lake Dam site. There is a rhythmic sequence of what appears to be soil, and charcoal and ash (Figure 24, p. 90). The younger White River tephra occurs near the top of the unit. This sequence probably represents a cycle of vegetation growth, forest fire, and reactivation of eolian erosion, followed by stabilization, and a repeat of the cycle. This site could potentially

Figure 24: Eolian deposit at the Marsh Lake Dam site. View is to the east, and the ice axe is 0.8 m long. The bottom of the unit has a very sharp contact with the top of the glaciolacustrine sediments. Dark bands in the unit are charcoal layers, and the younger White River tephra is visible as a discontinuous white band, near the head of the ice axe.

provide a Holocene fire history for the Whitehorse region. It is thought that global change may manifest in northern Canada as much warmer temperatures. Southern Yukon is a dry climate, and an increase in the temperature may lead to desertification. This site may shed light on past climate and vegetation during the Holocene when temperatures were warmer.

7.3 Thermokarst development

Klassen (1979), noted thermokarst development in fine grained sediments near Whitehorse in the sediments of Glacial Lake Champagne. Thermokarst was commonly observed in the Takhini Valley, but was not commonly observed in the sediments of Glacial Lake Champagne in other parts of the study area. If global change does manifest in northern Canada as much warmer temperatures, thermokarst may expand in valley bottoms that have permafrost, and are blanketed by sediments of Glacial Lake Champagne.

8.0 Conclusions and recommendations for future research

8.1 Sedimentology of the Glacial Lake Champagne basin

Detailed stratigraphic logging of the six exposures indicated three periods of deposition:

- i) the first, glacial deposition during ice cover, left till at the base of the sections (exposed only in the Millhaven Bay site);
- ii) the second, lacustrine deposition during deglaciation when the study area was inundated by meltwater forming Glacial Lake Champagne, left glaciolacustrine deposits in all sites;
- iii) the third, eolian deposition during the Holocene, left an eolian cap of reworked glaciolacustrine deposits on top of most sites which in places contains tephra which has been interpreted as the younger White River ash (1200 BP, Clague, 1989) and charcoal layers and lenses.

Facies were categorized into the following units.

Clay facies:

- i) massive, dark grey-olive, clay, CLm, with 76% clay, in thin, sharply defined layers in the Watson River site; deposited from clay settling out of the water column at a point where the energy had become extremely low within the water column.

Silt facies:

- ii) massive, beige-buff, silt, SLm, with no dropstones, a major component of all logs except the Millhaven Bay site; deposited from settling of silt laden water within the column rapidly;

- iii) weakly horizontally bedded, beige-buff, silt, SLh, grading into SLm, also in most sites; deposited in the same manner as the above unit, but with an indeterminate cyclicity, and probably less quickly;
- iv) massive, eolian silt, SLm(e), on top of the Tagish Lake site; reworked from previously deposited lake sediments.

Sand facies:

- v) massive, well sorted sand with no clasts, up to 1 m thick, Sm, in the Watson River site, the Whitehorse-Riverdale site, and the Marsh Lake Dam site. This unit could be the result of either turbidity events where it is associated with sand units with ripple or wavy structures, or as the result of sand settling out of the water column from suspension;
- vi) massive, well sorted, eolian sand with no clasts, Sm(e), on top of the Whitehorse-Yukon River site, interpreted as simple reworked glaciolacustrine sediment;
- vii) horizontally laminated, fine to coarse sand, Sh, with occasional clasts, up to 1 m thick, in the Marsh Lake Dam site, the Watson River site, the Millhaven Bay site, and the Whitehorse-Riverdale site, which is representative of cyclical deposition of suspended sediments;
- viii) well sorted, medium texture, eolian sand, 2 m thick, Sh(e), only in the Whitehorse-Riverdale site that is interpreted as an eolian deposit reworked from previously deposited lake sediments;
- ix) well sorted, fine, eolian sand, Sd(e), gently dipping, only in the Whitehorse-

Riverdale site, interpreted as eolian dune foresets;

x) fine to medium sand, with cross trough beds, St, or with ripple laminated cross beds, Sr, up to 10 m thick, with occasional small clasts, in every site except the Whitehorse-Yukon River site. This material was deposited as the result of density underflow events.

Gravel facies:

xi) massive, structureless gravel, medium sand matrix supported, Gms, in the Marsh Lake Dam site and Millhaven Bay site, up to 2 m thick, with sharp contacts, which was interpreted as being deposited by a high energy, high density turbidity event;

xii) weakly horizontally bedded gravel, medium sand matrix, Ghs, in the Marsh Lake Dam site, with strong normal grading from pea gravel to cobbles, and sharp contacts, also deposited by a high energy, high density turbidity event.

Diamicton facies:

xiii) unsorted, unstratified diamicton, Dmm, in the Millhaven Bay site, interpreted as a till;

xiv) locally reworked diamicton, Dmm(r) that overlies Dmm, and is interpreted as till reworked in a subaqueous lacustrine environment.

Colluvium facies:

xv) medium colluvial sand with clasts, on top of the Watson River site, that is of unknown origin, but is likely slumped lake sediments.

The mechanisms of sediment deposition within the glaciolacustrine environment

were as follows.

i) dispersal and settling of clay, silt, and fine sand particles from the water column.

This process yielded units of SLh, SLm, Sh, and Sm, and comprises a major portion of the stratigraphic record.

ii) slumping and or ice frontal discharge of sediment laden water causing density underflow events, which are manifested in the stratigraphic record as St and Sr units, as well as Sh units that grade from these, and Ghs units. This depositional environment, along with the settling out of fine grained particles were the most important sedimentation processes in the lake.

iii) reworking of till occurred locally in the Millhaven Bay site, where Dmm(r) overlies glacially deposited Dmm.

The six sections show a wide range of sedimentology between each other. This is due to the existence of very different sedimentation regimes operating simultaneously within the basin during the glacial lake event, recorded at each location. When evaluated together, they are representative of the range of deposits found within other Cordilleran glacial lakes, except that these are larger sections than are usually found, and there appears to be more turbidity events recorded as a proportion to the rest of the sedimentary packages in Glacial Lake Champagne than are normally seen in other Cordilleran glacial lakes.

8.2 A tentative reconstruction of the geomorphic evolution of the Glacial Lake Champagne basin

The Cordilleran Ice Complex covered all of southern Yukon during the Last Glacial Maximum. There were three primary regional sources for this ice. Ice from the St. Elias Mountains in the western part of Yukon Territory generally flowed east. Ice from the Cassiar Mountains in southern Yukon Territory generally flowed northeast along the strike of the Teslin Valley. Ice from the Coast Mountains spanning the British Columbia-Alaska border generally flowed north along the axes of the Southern Lakes system. Ice from these three sources coalesced over the area within NTS map sheets 115 A, and 105 D. As climate ameliorated, ice limits downwasted and shrank back towards the source areas. A large part of southern Yukon Territory became ice free as a result. This area was in effect a basin bounded by topography and ice fronts to an elevation of at least 765 m. This basin filled with meltwater from the still retreating ice fronts, giving rise to Glacial Lake Champagne. The work completed for this thesis indicates that two major ice blockages occurred in the Alsek River region, by ice of the St. Elias Range, and in the Lake Laberge region, by ice of the Cassiar Mountain complex. This major glacial lake occurred sometime between 12 500 BP, and 10 500 BP. The outlet of the lake is hypothesized to be in the Lake Laberge region, existing as a kame overflow channel, but may have been in through the Alsek River system. There is no evidence thus far of a catastrophic flood associated with the drainage of the lake water. Stratigraphic logs indicate that the glaciolacustrine depositional environment was typified by two major modes of sedimentation. Most of the sediment associated with the lake was settled out in a

low energy environment from sediment laden overflows or interflows entering the lake as plumes, or as high energy density underflow events initiating on oversteepened delta or fan fronts, or directly from supraglacial, englacial, or subglacial sources. The information provided by the exposures logged, and from the presence of kettles near Whitehorse and Little Atlin Lake suggests that the lake was bounded by rapidly wasting discontinuous stagnant ice in valley bottoms. The formation of deltas and shorelines south of Whitehorse indicate that the lake had at least two stable stands at 765 m, and a later, more poorly developed stand at 725 m. Glacial Lake Champagne existed at these two stands as a contiguous water body in the valleys and basins of the Dezadeash River, the Takhini River, the Yukon River south of Lake Laberge, the M'Clintock River, Six Mile River, the Watson River, the Wheaton River, Marsh Lake, Tagish Lake north of the Yukon-British Columbia border, Nares Lake, and Bennett Lake north of, and including, West Arm. At its maximum investigated extent, Glacial Lake Champagne had a surface area of approximately 2425 km², a long axis of 260 km, and a maximum depth of at least 159 m.

8.3 Recommendations for future research

Future research should be focussed in further defining the depositional environments of Glacial Lake Champagne, and further defining the paleogeography of the Glacial Lake Champagne basin. Specific questions and problems that could be investigated are:

1. Is there a basin-wide signal in the glaciolacustrine sediments of the lowering of Glacial Lake Champagne?

2. How does the frequency of density underflow deposits change with the regional relief? Are these deposits able to manifest at some distance from the presumed source?
3. What were the paleocurrent directions and strengths in the basin? What was the magnitude of the Coriolis effect in the basin?
4. Where was the outlet exactly, and how did it operate as a stable hydrologic control point?
5. How important was stagnant ice in defining the geometry and sedimentation of the basin?
6. When, and how long, did each stand of Glacial Lake Champagne exist?

References

- Bogen, J. 1983. Morphology and sedimentology of deltas in fjord and fjord valley lakes. *Sedimentary Geology*. Vol. 36, p. 245-267.
- Boggs, S. 1995. *Principles of Sedimentology and Stratigraphy*. 774 p.
- Bostock, H.S. 1948. *Physiography of the Canadian Cordillera, With Special Reference to the Area North of the Fifty-Fifth Parallel*. Geological Survey of Canada, Memoir 247.
- Bostock, H.S. 1952. *Geology of northwest Shaskwak Valley, Yukon Territory*. Geological Survey of Canada, Memoir 267.
- Bostock, H.S. 1966. Notes on glaciation in central Yukon Territory. Geological Survey of Canada, Paper 65-36.
- Bostock, H.S. 1969. Kluane Lake, Yukon Territory, its drainage and allied problems. Geological Survey of Canada, Paper 69-28.
- Brown, R.J.E. 1978. Permafrost: Plate 32. *In* Hydrologic Atlas of Canada. Fisheries and Environment Canada, Ottawa, 34 plates.
- Burn, C.R. 1987. Permafrost. *In* Guidebook to Quaternary Research in Yukon. Edited by S.R. Morison and C.A.S. Smith. XII INQUA Congress, Ottawa, Canada. National Research council of Canada, Ottawa, p. 21-25.
- Calkin, P.E., and Feenstra, B.H. 1985. Evolution of the Erie-Basin Great Lakes. *In* Quaternary Evolution of the Great Lakes. Edited by P.F. Karrow and P.E. Calkin. Geological Association of Canada Special Paper 30, p. 149-170.
- Clague, J.J. 1980. Late Quaternary geology and geochronology of British Columbia. Part 1; radiocarbon dates. Geological Survey of Canada, Paper 80-13.
- Clague, J.J. (compiler). 1989. Quaternary Geology of the Canadian Cordillera; Chapter 1 in Quaternary Geology of Canada and Greenland, R.J. Fulton (ed.); Geological Survey of Canada, no. 1 (also Geological Society of America, the Geology of North America, v. K-1).
- Clague, J.J. and Rampton, V.N. 1982. Neoglacial Lake Alsek. *Canadian Journal of Earth Sciences*. Vol. 19 p. 94-117.

- Clayton, L. 1983. Chronology of Lake Agassiz drainage to Lake Superior. *In* Glacial Lake Agassiz. Edited by J.T. Teller and L. Clayton. Geological Association of Canada Special Paper 26, p. 291-307.
- Day, J.H. 1962. Reconnaissance Soil Survey of the Takhini and Dezadeash Valleys in the Yukon Territory. Research Branch, Canada Department of Agriculture.
- Denton, G.H., and Stuiver, M. 1967. Late Pleistocene Glacial Stratigraphy and Chronology, Northeastern St. Elias Mountains, Yukon Territory, Canada. Geological Society of America Bulletin. Vol 78, p. 485-510.
- Desloges, J.R., and Gilbert, R. 1991. Sedimentary record of Harrison Lake: implications for deglaciation in southwestern British Columbia. Canada Journal of Earth Science. Vol. 28 p. 800-815.
- Dredge, L.A. 1982. Relict ice-scour marks and the late phases of Lake Agassiz in northernmost Manitoba. Canadian Journal of Earth Sciences. Vol 19, p. 1079-1087.
- Duk-Rodkin, A.S., and Hughes, O.L. 1991. Age relationships of Laurentide and montane glaciations, Mackenzie Mountains, Northwest Territories. Geographie physique and Quaternaire. Vol. 45, p. 79-90.
- Eschman, D.F. and Karrow, P.F. 1985. Huron Basin Glacial Lakes: a review. *In* Quaternary Evolution of the Great Lakes. Edited by P.F. Karrow and P.E. Calkin. Geological Association of Canada Special Paper 30, p. 81-93.
- Eyles, N. 1987. Late Pleistocene debris-flow deposits in large glacial lakes in British Columbia and Alaska. Sedimentary Geology Vol. 53, p. 33-71.
- Eyles, N. and Clague, J.J. 1991. Glaciolacustrine sedimentation during advance and retreat of the Cordilleran ice sheet in central British Columbia. Geographie Physique and Quaternaire. Vol. 45, p. 317-331.
- Eyles, N., Clark, B.M., and Clague, J.J. 1987. Coarse-grained sediment gravity flow facies in a large supraglacial lake. Sedimentology. Vol. 34, p. 193-216.
- Farrand, W.R., and Drexler, C.W. 1985. Late Wisconsinan and Holocene history of the Lake Superior basin. *In* Quaternary Evolution of the Great Lakes. Edited by P.F. Karrow and P.E. Calkin. Geological Association of Canada Special Paper 30, p. 17-32.
- Fisher, T.G., and Smith, D.G. 1994. Glacial Lake Agassiz: its northwest maximum extent and outlet in Saskatchewan (Emerson phase). Quaternary Science Reviews. Vol. 13 p. 845-858.

- Foscolos, A.E., Rutter, N.W., and Hughes, O.L. 1977. The use of pedological studies in integrating the Quaternary history of central Yukon. Geological Survey of Canada, Bulletin 271.
- Fritz, W.J., and Moore, J.N. 1988. Basics of Physical Stratigraphy and Sedimentology. 371 p.
- Fulton, R.J. 1965. Silt deposition in late-glacial lakes of southern British Columbia. American Journal of Science, Vol. 263, p. 553-570.
- Fulton, R.J. 1991. A conceptual model for the growth and decay of the Cordilleran Ice Sheet. Geographie Physique et Quaternaire. Vol. 45, p. 281-286.
- Geological Survey of Canada. 1991. Physiography of the Canadian Cordillera. Map 1701A 1:5 000 000
- Gilbert, R. 1975. Sedimentation in Lillooet Lake, British Columbia. Canadian Journal of Earth Sciences, Vol. 12, p. 1697-1711.
- Gilbert, R. 1990. Rafting in glacimarine environments. *In* Glacimarine Environments: Processes and Sediments. Edited by J.A. Dowdeswell and J.D. Scourse. Geological Society Special Publication No. 53 p. 105-120
- Gilbert, R., and Desloges, J.R. 1987. Sediments of ice-dammed, self draining Ape Lake, British Columbia. Canadian Journal of Earth Sciences. Vol. 24 p. 1735-1747.
- Gilbert, R., and Shaw, J. 1981. Sedimentation in proglacial Sunwapta Lake, Alberta. Canadian Journal of Earth Sciences. Vol. 18 p. 81-93.
- Goudie, A., Lewin, J., Richards, K., Anderson, M., Burt, T., Whalley, B., and Worsley, P. 1978. Geomorphological Techniques. 187 p.
- Guerts, M.A., and Dewez, V. 1993. Le Lac Glaciaire Nisling et le Pleistocene Dans le Bassin Superieur de la Nisling River, au Yukon. Geographie Physique et Quaternaire, Vol. 47, no. 1, p. 81-92.
- Gustavson, T.C. 1975. Sedimentation and physical limnology in proglacial Malaspina Lake, southeastern Alaska. *In* Glaciofluvial and Glaciolacustrine Sedimentation. Edited by A.V. Jopling, and B.C. McDonald. Society of Economic Paleontologists and Mineralogists, special publication no. 23. p. 249-263.
- Gustavson, T.C., Ashley, G.M., and Boothroyd, J.C. 1975. Depositional sequences in glaciolacustrine deltas. *In* Glaciofluvial and Glaciolacustrine Sedimentation. Edited by

- A.V. Jopling, and B.C. McDonald. Society of Economic Paleontologists and Mineralogists, special publication no. 23. p. 264-280.
- Houbolt, J.J.H.C., and Jonker, J.B.M. 1968. Recent sediments in the eastern part of the Lake of Geneva (Lac Lemman). *Geologie en Mijnbouw*. Vol. 47, p. 131-148.
- Hsu, K.J. and McKenzie, J.A. 1985. Swiss lakes as a geological laboratory; Part 2, Varves. *Naturwissenschaften*, Vol. 72, p. 1132-1144.
- Hughes, O.L. 1987. Quaternary Geology. *In* Guidebook to Quaternary Research in Yukon. Edited by S.R. Morison and C.A.S. Smith. XII INQUA Congress, Ottawa, Canada. National Research council of Canada, Ottawa, p. 12-16.
- Hughes, O.L. 1990. Surficial geology and geomorphology, Aishihik Lake, Yukon Territory. Geological Survey of Canada, Paper 87-29.
- Hughes, O.L., Campbell, R.B., Muller, J.E., and Wheeler, J.O. 1969. Glacial Limits and Flow Patterns, Yukon Territory, South of 65 Degrees North Latitude. Geological Survey of Canada Paper 68-34.
- Hughes, O.L., Rampton, V.N., and Rutter, N.W. 1972. Quaternary geology and geomorphology, southern and central Yukon (northern Canada). 24th International Geological Congress (Montreal, Quebec), Guidebook, Field Excursion A11.
- Hughes, O.L., Rutter, N.W., and Clague, J.J. 1989. Yukon Territory; Chapter 1 *In* Quaternary Geology of Canada and Greenland, Edited by R.J. Fulton. Geological Survey of Canada, no. 1 (also Geological Society of America, the Geology of North America, v. K-1).
- Jackson, L.E., Ward, B., Duk-Rodkin, A., and Hughes, O.L. 1991. The last Cordilleran Ice Sheet in southern Yukon Territory. *Geographie Physique et Quaternaire*. Vol. 45, p. 341-354.
- Johnson, G., and Raup, P. 1974. Investigations in Southwest Yukon: Geobotanical and Archaeological Reconnaissances. Papers of the Robert S. Peabody Foundation for Archaeology, Vol. 6, p. 1-198.
- Jopling, A.V., and Walker, R.G. 1968. Morphology and origin of ripple-drift cross-lamination, with examples from the Pleistocene of Massachusetts. *Journal of Sedimentary Petrology*, Vol. 38, p. 971-984.
- Kaszyski, C.A. 1987. A model for glacial and proglacial sedimentation in the shield terrane of southern Ontario. *Canadian Journal of Earth Science*. Vol. 24 p. 2373-2391.

- Kempe, S., and Degens, E.T. 1978. Varves in the Black Sea and in Lake Van (Turkey). *In* Moraines and Varves. Edited by C. Schluchter. Proceedings of an INQUA symposium on genesis and lithology of Quaternary deposits. p. 309-318.
- Kerr, D.E. 1987. Depositional environments during a glaciolacustrine to marine transition in the Richardson and Rae River basin, N.W.T. Canadian Journal of Earth Sciences. Vol. 24, p. 2130-2140.
- Kindle, E.D. 1953. Dezadeash Map-Area, Yukon Territory. Geological Survey of Canada Memoir 268.
- Klassen, R.W. 1978. A unique stratigraphic record of late Tertiary-Quaternary events in southeastern Yukon Territory. Canadian Journal of Earth Sciences. Vol. 15, p. 1884-1886.
- Klassen, R.W. 1979. Thermokarst terrain near Whitehorse, Yukon Territory. *In* Current Research, Part A. Geological Survey of Canada, Paper 79-1A. p. 385-388.
- Klassen, R.W. 1987. The Tertiary-Pleistocene stratigraphy of the Liard Plain, southeastern Yukon Territory. Geological Survey of Canada, Paper 86-17.
- Lambert, A. 1982. Trubestrome des Rheins am Grund des Bodensees (Turbidity currents from the Rhein River on the bottom of Lake Constance). Wasserwirtschaft. Vol. 72, p. 1-4.
- Lambert, A.M., and Hsu, K.J. 1979. Non-annual cycles of varve-like sedimentation in Walensee, Switzerland. Sedimentology. Vol. 26, p. 453-461.
- Lambert, A.M., Kelts, K.R., and Marshall, N.F. 1976. Measurement of density underflows from Walensee, Switzerland. Sedimentology. Vol. 23, p. 87-105.
- Larsen, C.E. 1985. Lake level, uplift, and outlet incision, the Nipissing and Algoma Great Lakes. *In* Quaternary Evolution of the Great Lakes. Edited by P.F. Karrow and P.E. Calkin. Geological Association of Canada Special Paper 30, p. 63-77.
- Leeder, M.R. 1982. Sedimentology. 344 p.
- Lemmen, D.S., Duk-Rodkin, A., and Bednarski, J.M. 1994. Late glacial drainage systems along the northwestern margin of the Laurentide ice sheet. Quaternary Science Reviews. Vol. 13, p. 805-828.
- Lewis, D.W., and McConchie, D. 1994. Practical Sedimentology. 213 p.

Lerbekmo, J.F., and Campbell, F.A. 1969. Distribution, composition, and source of the White River ash, Yukon Territory. *Canadian Journal of Earth Sciences*, Vol. 6, p. 109-116.

Lerbekmo, J.F., Westgate, J.A., Smith, D.G.W., and Denton, G.H. 1975. New data on the character and history of the White River volcanic eruption, Alaska. *In Quaternary Studies Edited by R.P. Suggate, and M.M. Cresswell*. Royal Society of New Zealand, Bulletin 13, p.203-209.

Liverman, D.G.E. 1987. Sedimentation in ice-dammed Hazard Lake, Yukon. *Canadian Journal of Earth Sciences*. Vol. 24, p. 1797-1806.

Ludlam, S.D. 1978. Rhythmite deposition in lakes of the northeastern United States. *In Moraines and Varves. Edited by C. Schluchter*. Proceedings of an INQUA symposium on genesis and lithology of Quaternary deposits. p. 295-302.

Mann, D.H. and Hamilton, T.D. 1995. Late Pleistocene and Holocene paleoenvironments of the North Pacific Coast. *Quaternary Science Reviews*. Vol. 14 p. 449-471.

Mathews, W.H. 1986. Physiographic map of the Canadian Cordillera. Geological Survey of Canada. Map 1701A, scale 1:5 000 000.

Miall, A.D. 1977. A review of the braided-river depositional environment. *Earth-Science Reviews* Vol. 13, p 1-62.

Miller, B.B., Karrow, P.F., and Mackie, G.L. 1985. Late Quaternary molluscan faunal changes in the Huron Basin. *In Quaternary Evolution of the Great Lakes. Edited by P.F. Karrow and P.E. Calkin*. Geological Association of Canada Special Paper 30, p. 95-107

Monger, J.W.H., Price, R.H., and Tempelman-Kluit, D.J. 1982. Tectonic accretion and the origin of two major metamorphic and plutonic belts in the Canadian Cordillera. *Geology* Vol. 10 p. 70-75

Morison, S.R. and Klassen, R.W. 1991. Surficial geology, Whitehorse, Yukon Territory. Geological Survey of Canada, Map 12-1990, scale 1:250 000.

Morlan, R.E. 1980. Taphonomy and archaeology in the Upper Pleistocene of the northern Yukon Territory: a glimpse of the peopling of the New World. National Museums of Canada, National Museum of Man, Mercury Series, Archaeological Survey of Canada, Paper 94.

Muller, E.H. and Prest, V.K. 1985. Glacial Lakes in the Ontario Basin. *In Quaternary Evolution of the Great Lakes. Edited by P.F. Karrow and P.E. Calkin*. Geological

Association of Canada Special Paper 30, p. 213-229.

North American Commission on Stratigraphic Nomenclature. 1983. North American Stratigraphic Code. American Association of Petroleum Geologists Bulletin. Vol. 67 p. 841-875.

Pharo, C.H., and Carmack, E.C. 1979. Sedimentation processes in a short residence-time intermontane lake, Kamloops Lake, British Columbia. Sedimentology. Vol. 26 p. 523-541.

Pickrill, R.A., and Irwin, J. 1982. Predominant headwater inflow and its control of lake-river interactions in Lake Wakatipu. New Zealand Journal of Freshwater Resources. Vol. 16, p. 201-213.

Powell, R.D. 1981. A model for sedimentation by tidewater glaciers. Annals of Glaciology. Vol. 2, p. 129-134.

Rampton, V.N. 1971. Late Pleistocene glaciations of the Snag Klutlan area, southwestern Yukon Territory, Canada. Arctic. Vol. 24 p. 277-300.

Rampton, V.N. 1972. Surficial Deposits: Alaska Highway, Whitehorse to Alaska-Yukon Boundary. In Current Research, Part B. Geological Survey of Canada, Paper 72-1B. p. 119-122.

Reimer, G.E. 1984. The sedimentology and stratigraphy of the southern basin of Glacial Lake Passaic, New Jersey. Unpublished M.S. thesis, Rutgers University, New Brunswick, New Jersey.

Rust, B.R. 1977. Mass flow deposits in a Quaternary succession near Ottawa, Canada: diagnostic criteria for subaqueous outwash. Canadian Journal of Earth Sciences. Vol. 14, p. 175-184.

Rust, B.R., and Romanelli, R. 1975. Late Quaternary subaqueous outwash deposits near Ottawa, Canada. *In* Glaciofluvial and Glaciolacustrine Sedimentation. Edited by A. V. Jopling, and B.C. McDonald. Society of Economic Paleontologists and Mineralogists, special publication no. 23. p. 177-192.

Ryder, J.M., and Maynard, D. 1991. The Cordilleran Ice Sheet in northern British Columbia. Geographie Physique and Quaternaire. Vol. 45, p. 353-363.

Saunders, I.R., Clague, J.J., and Roberts, M.C. 1987. Deglaciation of Chilliwack River valley, British Columbia. Canadian Journal of Earth Sciences, Vol. 24, p. 915-923.

Sawicki, O. 1990. Geomorphology of Glacial Lake Invermere, upper Columbia River

valley, British Columbia. Unpublished M.Sc. thesis, University of Calgary, Calgary.

Schmok, J.P., and Clarke, G.K.C. 1989. Lacustrine sedimentary record of ice-dammed Neoglacial Lake Alsek. *Canadian Journal of Earth Sciences*. Vol. 26, p. 2092-2105.

Sharpe, D.R., and Shaw, J. 1989. Erosion of bedrock by subglacial meltwater, Cantley, Quebec. *Geological Society of America Bulletin*. Vol. 101, p. 1011-1020.

Shaw, J. 1975. Sedimentary successions in Pleistocene ice-marginal lakes. *In* *Glaciofluvial and Glaciolacustrine Sediments*. Edited by A.V. Jopling, and B.C. McDonald. Society of Economic Paleontologists and Mineralogists, special publication no. 23. p. 281-303.

Shaw, J. 1979. Genesis of the Sveg tills and Rogen moraines of central Sweden: a model of basal meltout. *Boreas*, Vol. 8, p. 409-426.

Shaw, J., and Archer, J. 1979. Deglaciation and glaciolacustrine sedimentation conditions, Okanagan Valley, British Columbia, Canada. *In* *Moraines and Varves*. Edited by C. Schluchter. Proceedings of an INQUA symposium on genesis and lithology of Quaternary deposits. p.347-351.

Smith C.A.S., Tarnocai, C., and Hughes, O.L. 1986. Pedological investigations of Pleistocene glacial drift surfaces in the central Yukon. *Geographie Physique and Quaternaire*. Vol. 40 p. 29-37.

Smith, D.G. 1992. Glacial Lake Mackenzie, Mackenzie Valley, Northwest Territories, Canada. *Canadian Journal of Earth Sciences*. Vol. 29 p. 1756-1766.

Smith, D.G. 1994. Glacial Lake McConnell: paleogeography, age, duration, and associated river deltas, Mackenzie River basin, western Canada. *Quaternary Science Reviews*. Vol. 13 p. 829-843.

Smith M.W., and Riseborough, D.W. 1983. Permafrost sensitivity to climatic change. Proceedings, Fourth International Conference on Permafrost, N.A.S., Washington, D.C. p. 1173-1183.

Smith, N.D. 1981. The effect of changing sediment supply on sedimentation in a glacier-fed lake. *Arctic and Alpine Research*, Vol. 13, p. 75-82.

Smith, N.D., and Ashley, G. 1985. Proglacial lacustrine environment. *In* *Glacial sedimentary environments*. Edited by G.M. Ashley, J. Shaw, and N.D. Smith. SEPM short course no. 16.

Smith, N.D., Vendl, M.A., and Kennedy, S-K. 1982. Comparison of sedimentation

regimes in four glacier-fed lakes of western Alberta. *In* Research in Glacial, Glacio-Fluvial, and Glaciolacustrine Systems. Edited by R. Davidson-Arnott, W. Nickling, and B.D. Fahey. Proceedings of the 6th Guelph Symposium on Geomorphology. p. 671-688.

St-Onge, D.A. 1980. Glacial Lake Coppermine, north-central District of Mackenzie, Northwest Territories. *Canadian Journal of Earth Sciences*. Vol. 17 p. 1310-1315.

Sturm, M. 1978. Origin and composition of clastic varves. *In* Moraines and Varves. Edited by C. Schluchter. Proceedings of an INQUA symposium on genesis and lithology of Quaternary deposits. p. 281-285

Tarnocai, C., Smith, S. and Hughes, O.L. 1985. Soil development on Quaternary deposits of various ages in the central Yukon Territory. *In* Current Research, Part A. Geological Survey of Canada, Paper 85-1A. p. 229-238.

Teller, J.T. 1985. Glacial Lake Agassiz and its influence on the Great Lakes. *In* Quaternary Evolution of the Great Lakes. Edited by P.F. Karrow and P.E. Calkin. Geological Association of Canada Special Paper 30, p. 1-16.

Tempelman-Kluit, D.J. 1974. Reconnaissance geology of Aishihik Lake, Snag, and part of Stewart River map-areas, west-central Yukon. Geological Survey of Canada, Paper 73-41.

Thomas, G.S.P. 1984. Sedimentation of a sub-aqueous esker-delta at Strabathie, Aberdeenshire. *Scottish Journal of Geology*. Vol. 20, p. 9-20.

Vernon, P., and Hughes, O.L. 1966. Surficial geology, Dawson, Larsen Creek, and Nash Creek map-areas, Yukon Territory (116 B and 116 C E1/2, 116 A, and 106 D). Geological Survey of Canada, Bulletin 136.

Walker, R.G. 1964. Turbidite sedimentary structures and their relationship to proximal and distal depositional environments. *Journal of Sedimentary Petrology*, Vol. 37, p. 25-43.

Walker, R.G. 1984. Facies models, 2nd edition. Geoscience Canada reprint series 1. 317 p.

Weirich, F.H. 1984. Turbidity currents: monitoring their occurrence and movement with a three dimensional sensor network. *Science*. Vol. 224, p. 384-387.

Williams, P.J., and Smith, M.W. 1989. *The Frozen Earth*. 306 p.

Wheeler, J.O. 1961. Whitehorse map-area, Yukon Territory. Geological Survey of Canada, Memoir 312.

Appendix A: List of airphoto flight lines and photo numbers

Flight Line	Photo numbers
A22437	79
A11524	124-133; 244-251
A11631	361-371
A11521	309-316; 319; 459-466
A10552	11-21; 26-40; 112-130
A11447	256
A10560	78-93
A10563	120-150
A10567	58-86
A11478	55-92

Appendix B: Results of grain size analysis of samples

Sample number/location		%clay	%silt	%sand
3.8	Marsh Lake Dam	11.12	13.11	75.77
3.9	Marsh Lake Dam	23.89	43.42	32.69
3.10	Marsh Lake Dam	5.05	4.85	90.10
3.13a	Marsh Lake Dam	18.49	44.63	36.88
3.13b	Marsh Lake Dam	4.56	7.38	88.07
3.14	Marsh Lake Dam	0.58	2.04	97.38
4.1	Whitehorse-Yukon River	7.71	83.03	9.22
5.1	Whitehorse-Riverdale	3.26	5.93	90.81
5.2	Whitehorse-Riverdale	1.53	8.17	90.30
5.3	Whitehorse-Riverdale	4.55	13.09	82.35
5.4	Whitehorse-Riverdale	4.19	1.32	94.50
5.5	Whitehorse-Riverdale	1.47	0.16	98.37
5.6	Whitehorse-Riverdale	7.84	65.44	26.73
5.7	Whitehorse-Riverdale	7.04	45.24	47.72
5.8	Whitehorse-Riverdale	9.24	70.35	20.41
5.9	Whitehorse-Riverdale	4.41	60.74	34.75
5.10	Whitehorse-Riverdale	5.53	24.79	69.68
5.11	Whitehorse-Riverdale	1.81	6.46	91.73
5.12	Whitehorse-Riverdale	3.79	9.20	87.01
5.13	Whitehorse-Riverdale	3.54	15.21	81.25
5.14	Whitehorse-Riverdale	10.41	87.79	1.80
5.15	Whitehorse-Riverdale	9.86	44.51	45.63
5.16	Whitehorse-Riverdale	11.24	65.39	23.37
5.18	Whitehorse-Riverdale	11.73	47.52	40.75
5.19	Whitehorse-Riverdale	4.76	90.49	4.75
5.20	Whitehorse-Riverdale	4.30	21.45	74.25
5.22	Whitehorse-Riverdale	1.85	64.38	33.77
10.1	Watson River	1.48	0.40	98.12
10.2	Watson River	1.93	11.67	86.39
10.3	Watson River	3.78	67.24	28.98
11.1	on Carcross Highway	76.28	22.88	0.83
15.1	Tagish Lake	2.96	82.78	14.26
15.2	Tagish Lake	1.97	34.39	63.65
16.1	Millhaven Bay	0.65	2.63	96.72
17.3	on Takhini River	1.57	41.69	56.74
18.1	on Yukon River	1.72	97.82	0.47
Carcross Desert		0.08	1.46	98.45

IMAGE EVALUATION TEST TARGET (QA-3)

APPLIED IMAGE, Inc.
1653 East Main Street
Rochester, NY 14609 USA
Phone: 716/482-0300
Fax: 716/288-5989

© 1993, Applied Image, Inc., All Rights Reserved