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 cryostratigraphic and isotopic characteristics of "muck"
deposits, Klondike area, Yukon Territory.**

by

Erica Kotler

A thesis submitted to the Faculty of
Graduate Studies in partial fulfillment of
the requirements for the degree of
Master of Science

Carleton University
Ottawa, Ontario
September 14, 1998

© 1998 Erica Kotler

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.

0-612-36933-1

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

**The cryostratigraphic and isotopic characteristics of “muck”
deposits, Klondike area, Yukon Territory**
submitted by **Erica Kotler**, in partial
fulfilment of the requirements for the
degree of Master of Science

Chair, Department of Earth Sciences

Thesis Supervisor

Abstract

Four cryostratigraphic units have been recognized within frozen, unconsolidated materials (muck) which overlie creek gravel in the Klondike area. They represent deposition during Late Pleistocene and Holocene glacial/interglacial periods. The lowermost Last Chance Creek Formation, is a Reid-McConnell interglacial deposit, consisting of ice-rich silt and organic material >40 ka old. Ice in this formation has $\delta^{18}\text{O}$ values between -29 and -26 ‰. The Quartz Creek Formation is a McConnell-age deposit consisting of redeposited loess with $\delta^{18}\text{O}$ values for pore ice between -32 and -29 ‰. The Dago Hill Formation is an ice-rich, late-glacial deposit consisting of colluviated silt and organic material, which has $\delta^{18}\text{O}$ ratios for ice between -28 and -25 ‰. Accumulation began by $11\,620 \pm 90$ BP at the start of the Bølling-Allerød warm interval. Ice in the uppermost Holocene Organic unit has $\delta^{18}\text{O}$ values between -25 and -20 ‰. Thaw unconformities mark the base of the Dago Hill Formation and Organic unit. The sediments contain, on average, 70 % ice by volume.

Acknowledgements

Field and laboratory work in 1997 and 1998 was possible through support from the Exploration and Geological Services Division of the Department of Indian and Northern Affairs, Canada, a Northern Scientific Training Program grant, and a National Science and Engineering Council of Canada grant to C.R. Burn. I owe much of the success of the fieldwork to my two companions: Anne-Pascale Bartleman, whose unwavering forearms and constant enthusiasm kept the samples coming and the mood cheerful, and Lancelot, who never failed to provide entertainment. None of the fieldwork would have been possible without the outstanding generosity of the Klondike miners and their families, who allowed us access to their sites and whose unsatiable curiosity about muck and the girls that study it, always led to helpful discussions. In particular, I thank Lee Olynyk, Ron Toews, John Alton, Torfin Djukastein, Bob Cattermole, Jerry Bryde, Lou Peronovitch, Dave Miller, Brenda and Roger Kay, and Jerry and Betsy Ahnert. In Mayo, the hospitality of Jim and Shann Carmichael, Michel Menelon, the Coopers, and the McGregors is gratefully acknowledged. Many thanks go to Bill Lebarge for supporting the research and assisting at various times throughout the fieldwork. Helpful discussions were provided by Scott Smith, Wayne Pollard and John Storer. In Ottawa, thanks go to Larry Boyle and colleagues at the Geocryology Research Laboratory, Wendi Abdi at the G.G. Hatch Isotope Laboratory and Chris Burn, for keeping me on track and providing constant encouragement. Finally, I thank the

people who made my time at Carleton interesting throughout: Laurent Godin (for well-timed doses of snow), Steven Turgeon, Jeff Kiernan, Dan Gibson, Hassan Mirnejad and Tim Kilby.

Table of Contents

Thesis Acceptance	ii
Abstract	iii
Acknowledgements	iv
Table of Contents	vi
List of Tables	viii
List of Figures	ix
List of Appendices	x
 CHAPTER 1: INTRODUCTION	 1
 CHAPTER 2: GEOLOGICAL SETTING	 4
2.1 Tectonic Setting	4
2.2 Bedrock Geology	4
2.3 Cenozoic Stratigraphy	6
2.3.1 Gravelly Deposits	6
2.3.2 Muck Deposits	8
2.4 Valley Evolution	10
2.5 Glacial History	11
 CHAPTER 3: BACKGROUND GEOCRYOLOGY	 16
3.1 Cryostratigraphy	16
3.1.1 Ice Types	16
3.1.2 Structure	23
3.1.3 Contacts	24
3.1.4 Summary	25
3.2 H-O Isotopes	25
3.2.1 Fractionation and the Meteoric Water Line	26
3.2.2 ^{18}O and ^2H in Permafrost	27
3.2.3 Summary	29
3.3 Chapter Summary	30
 CHAPTER 4: METHODS	 31
4.1 Study Area	31
4.2 Cryostratigraphy	33
4.2.1 Stratigraphy	33
4.2.2 Ice Characteristics and Content	36
4.2.2.1 <i>Field Determinations</i>	38
4.2.2.2 <i>Laboratory Determinations</i>	38

4.3 Isotope Geochemistry	42
4.3.1 Field Methods	42
4.3.2 Laboratory Methods	43
CHAPTER 5: RESULTS	44
5.1 Distribution of Deposits	44
5.2 Cryostratigraphy	46
5.2.1 Stratigraphy	46
5.2.2 Ice Characteristics	56
5.2.3 Ice Content	64
5.3.4 Summary of Cryostratigraphy	70
5.3 Isotope Geochemistry	70
5.3.1 $\delta^{18}\text{O}$ Profile	72
5.3.2 Co-Isotope Relations	72
5.4 Chapter Summary	75
CHAPTER 6: DISCUSSION OF RESULTS	76
6.1 Distribution of Deposits	76
6.2 Cryostratigraphy	76
6.2.1 Depositional and Freezing History	77
6.2.2 Ice Content Determinations	83
6.3 Isotope Geochemistry	84
6.4 Comparison of Klondike and Alaskan Deposits	87
6.5 Chapter Summary	90
CHAPTER 7: CONCLUSIONS	91
REFERENCES	93
Appendix I	103
Appendix II	115

List of Tables

Table 4.1: Section locations.	35
Table 4.2: Nomenclature used in ice content determinations	40
Table 5.1: Results of radiocarbon dating	49

List of Figures

2.1: Study area and physiographic regions of Yukon	5
2.2: Development of valley-in-valley profiles	12
2.3: Summary of Quaternary history of unglaciated and glaciated Yukon	15
3.1: Development of syngenetic, epigenetic and anti-syngenetic ice wedges	21
4.1: Monitoring on Gold Bottom Creek	32
4.2: Study and drilling locations within Klondike area	34
4.3: Water-jet drilling on Gold Bottom Creek	37
5.1: Results of drilling on Gold Bottom Creek	45
5.2: Last Chance Creek Formation with syngenetic ice wedge	48
5.3: Tephra bed within Quartz Creek Formation	51
5.4: Contact between Quartz Creek Formation and Dago Hill Formation	53
5.5: Folded sediments within Dago Hill Formation	54
5.6: Radiocarbon sample locations at DC-2	55
5.7: Cryostratigraphy at LCC-1 section	58
5.8: Cryostratigraphy at CAMP-1 section	60
5.9: Deformed Holocene ice wedge at CAMP-1 section	62
5.10: Intrusive ice at DG-1 section	63
5.11a: Fractures within sediments at DG-1	65
5.11b: Banding observed within fracture-infill ice	65
5.12: Results of ice content determinations of Dago Hill Formation	67
5.13: Determination of the volumetric ice content of a deposit	69
5.14: Generalized cryostratigraphy of muck in the Klondike	71
5.15: Isotope profile across Quartz Creek-Dago Hill Formation contact	73
5.16: Co-isotope plot of all Klondike samples	74

List of Appendices

I: Section Illustrations	
Legend to accompany illustrations	103
LCC-1	104
CAMP-1	105
DC-1	106
DC-2	107
DG-1	108
DG-2	109
DP-1	110
FV-1	111
GB-1	112
GB-2	113
RB-1	114
II: Complete results of isotope geochemistry	115

CHAPTER 1: INTRODUCTION

A portion of northwestern North America remained unglaciated throughout Late Tertiary to Quaternary time, while most of the continent was intermittently blanketed by ice sheets. This unique environment constitutes Beringia. The deposits which represent deposition during this time, have been the focus of numerous studies in Alaska (e.g., Péwé 1955, 1966, 1975, 1989; Péwé et al. 1997), and fewer in Yukon Territory (e.g., Tyrrell 1917; Harington and Clulow 1973; Naldrett 1981; Fraser and Burn 1997). Primarily consisting of loess (Fraser and Burn 1997), these frozen, unconsolidated, ice and organic-rich sediments, referred to as "muck" (Harington and Clulow 1973), overlie auriferous creek gravels in valley-bottoms in the Klondike area of Yukon Territory.

In the Klondike area, the stratigraphy and sedimentology of these deposits was described in detail by Fraser and Burn (1997), who recognized two units: the Silt unit, consisting of primary and redeposited loess of Late Wisconsinan age, and the Organic unit, a disorganized to crudely-bedded Holocene deposit consisting mostly of organic material. The geocryology of these deposits was examined to some extent by Naldrett (1981) and Fraser and Burn (1997), who recognized epigenetic and syngenetic ice wedges within the two units. French and Pollard (1986) attempted to identify massive ice exposed within muck on Hunker Creek based on its cryostratigraphic location and petrographic characteristics. No regional examination of the cryostratigraphy or isotopic characteristics of these deposits has been undertaken, nor have any comparisons been made between the

Klondike deposits and those of adjacent Alaska.

Cryostratigraphy has been used in the western Canadian Arctic, in Yukon and in Siberia to infer permafrost conditions during and after deposition of sediments, by identifying ground ice types and examining their relations to adjacent sediments (e.g., French et al. 1982; Black 1983; Burn et al. 1986; Vasil'chuk and Vasil'chuk 1995; Burn 1997). The isotope geochemistry of ground ice has been used to infer paleoclimatic conditions, by relating $^{18}\text{O}/^{16}\text{O}$ and $^2\text{H}/\text{H}$ ratios of permafrost waters to surface temperatures at the time of precipitation (e.g., Michel and Fritz 1982; Mackay 1983 α ; Burn et al. 1986; Michel 1990).

The main objectives of this study are: (1) to build upon the existing sedimentological framework for muck in the Klondike area, by examining in detail the cryostratigraphy and isotope geochemistry of these ice-rich deposits; (2) to deduce paleoclimatic and permafrost conditions at the time of deposition; (3) to determine the characteristic ice content of the deposits; and, (4) to compare these deposits with those reported in Alaska.

Radiocarbon dating of organic material has been used to constrain depositional ages. In this thesis, dates are presented in uncalibrated radiocarbon years before present, where 'present' refers to 1950 AD. The calendric ages corresponding to 8000, 10 000 and 12 000 radiocarbon years BP, are approximately 9000, 11 500 and 14 000 years BP, respectively, as suggested by Bard et al. (1990) and Bartlein et al. (1995).

No formal names exist for stratigraphic units recognized within these

deposits. Three names are herein introduced. These are informal designations which are proposed to ease in identification and correlation of the sediments.

CHAPTER 2: GEOLOGICAL SETTING

2.1 TECTONIC SETTING

The study area lies within Klondike Plateau, a Tertiary-age highland bounded by the Kluane Plateau to the south and Tintina Trench to the north (Bostock 1970; Tempelman-Kluit 1980, Figure 2.1). Tintina Trench is the surface expression of a northwest-trending dextral strike-slip fault, associated with the Columbian orogeny, which has experienced up to 700 km of offset (Roddick 1967; Tempelman-Kluit 1972). A series of northwest-trending lineaments within the Klondike area are surface expressions of secondary faults associated with movement along the trench (Gleeson 1970; Milner 1976).

2.2 BEDROCK GEOLOGY

Rocks of the Klondike Plateau are part of the Omineca Belt, a medium to high-grade suite of Proterozoic and Paleozoic metasedimentary, gneissic and ultramafic rocks including the Klondike Series and Nasina Series (McConnell 1905; Green 1972). The Klondike Series comprises quartz, quartz-sericite and chlorite schists of the Klondike Schist, and granites and mylonites of the Pelly Gneiss. The Nasina Series includes recessive graphitic and pyritic phyllites, quartzites and marbles. The graphitic phyllite of the Nasina Series is the dominant rock type of the Klondike area. It is commonly fragmented and contains secondary gold, either precipitated from groundwater or incorporated by erosion from overlying, auriferous gravel deposits. Ultramafic rocks include peridotite,

Figure 2.1: Location of study area and major physiographic regions.

serpentinite and steatite, associated either with alteration along lineaments, or displacement of ophiolites along the Tintina Fault (Green 1972; Naldrett 1981). Late Cretaceous to Lower Tertiary rocks include paraconglomerates, which are locally intruded by basic dykes, intermediate pyroclastic rocks and felsic plugs (Naldrett 1981). These volcanic rocks are Late Tertiary in age and form part of the Carmacks Group (Tempelman-Kluit 1974).

2.3 CENOZOIC STRATIGRAPHY

The Cenozoic stratigraphy comprises unconsolidated deposits which unconformably overlie bedrock. They are divided here into two general units: gravelly deposits and muck deposits (silt and organic material).

2.3.1 Gravelly Deposits

Three units make up the gravelly deposits of the Klondike area: White Channel Gravel, Klondike Gravel and Creek Gravel (McConnell 1907; Milner 1976).

White Channel Gravel

The partially-consolidated White Channel Gravel unit overlies bedrock and forms prominent high-level terraces in the valleys of the Klondike area. The abundance of gold found within this unit makes it one of the richest placer deposits in the world. As many as fourteen lithofacies have been described, consisting of

massive to imbricated to cross-bedded quartz-rich sediments ranging in size from sand to cobbles (Morison 1985). The unit is up to 45 m thick (Duk-Rodkin 1996), and has been interpreted as being dominantly fluvial in origin (Morison 1989; Froese and Hein 1996). Ice-wedge casts have been found near the top of the unit indicating that permafrost developed and persisted for some time after the deposition of the gravel (Milner 1976; Morison 1985; Duk-Rodkin 1996).

Klondike Gravel

The Klondike Gravel unit overlies, and is interbedded with, White Channel Gravel in intermediate to high-level terraces. It consists of mostly sub-rounded to rounded clasts of orange-yellow quartzite and chert derived from the Ogilvie Mountains to the north (Bostock 1966). This unit has been extensively altered in places by hydrothermal fluids, giving it a chalky, weathered appearance, and a yellowish stain. Cross-beds, imbrication and channel structures dominate, with a decrease in development towards the top of the unit (Naldrett 1981). The Klondike Gravel unit is glaciofluvial in origin and corresponds with deposition during advance and retreat of a Late Pliocene Pre-Reid glaciation (Hughes et al. 1972; Naldrett 1981). Ice-wedge casts have been reported within this unit (Froese 1997).

Creek Gravel

These poorly to very well sorted gravels are found in the valley bottoms of all major drainages of the Klondike area. The large variation in lithology and

structure observed within these deposits reflects the diversity of origin of the sediments. The unit is largely composed of reworked colluvial material, including schistose bedrock, yellow and white gravels, and organic material. Minor concentrations of gold are found where there has been reworking and accumulation of White Channel Gravel. *In situ* organic material occurs as lenses within the gravel reflecting fluctuations in stream levels during deposition (Naldrett 1981).

2.3.2 Muck Deposits

Brown to black, organic-rich muck deposits overlie creek gravel and bedrock in valley-bottoms of the Klondike. If undisturbed, these deposits are frozen, but they emit a putrid odour when thawing (Richardson 1861). Gold does not occur in economic quantities in these sediments, although extra-fine gold is believed to occur as a precipitate (Placer Mining Section 1996). Facies within this unit have previously been identified based on their clastic characteristics, particularly foreign clast content and abundance (Campbell 1952; Milner 1976; Naldrett 1981). The unit has been interpreted as a colluvial deposit, derived from loess transported to the Klondike during the Late Pleistocene glaciation (Campbell 1952). In detailed investigations on the origin of these deposits, Fraser and Burn (1997) recognized two distinct stratigraphic units: a Silt unit and an Organic unit.

Silt unit

The Silt unit consists of light grey to grey-brown massive and bedded loess

(87% by weight medium silt to fine sand), consisting of dominantly subangular chlorite, micas, quartz and plagioclase feldspar. The unit unconformably overlies gravels and bedrock in valley-bottoms and on slopes. Colluviation and reworking of this material has redeposited some of the sediment into valley-bottoms, where it is up to 25 m thick. Several tephra beds have been identified, some with evidence of fluvial reworking. Grassbeds and squirrel nests occur in the lower levels of the unit, and rhizomes are dispersed throughout (Fraser and Burn 1997). Beringian mammal remains, preserved at the base of the unit, have been identified (Harrington and Clulow 1973).

Fraser and Burn (1997) interpreted this unit as a Late Pleistocene deposit which began accumulating in a grassland environment 27 000 ^{14}C years BP. The silt is believed to originate from the floodplain of the Yukon River, in a system comparable to that of modern-day Slims River delta in southern Yukon (Nickling 1978; Laxton et al. 1996).

Massive ice occurring in the lower portions of this unit at one section was interpreted as either buried snowbank or segregated ice (French and Pollard 1986). Epigenetic or syngenetic ice wedges have been identified in the upper portions of this unit (e.g., Fraser and Burn 1997). These wedges do not extend into the overlying unit.

Organic Unit

The Organic unit unconformably overlies either the Silt unit or creek gravel.

It is composed mostly of brown to black peat, and contains organic macrofossils such as tree branches, roots and stumps. Material is disorganized to crudely bedded and contains colluvial pebbles in places (Fraser and Burn 1997). Basal dates suggest accumulation beginning by 9945 ± 150 BP (BGS-1772) at the start of the Holocene when the climate became abruptly wetter (Fraser and Burn 1997). Small ice wedges occur at various stratigraphic positions within this unit, including ones which are truncated by the present active layer.

Muck in Alaska

Fine-grained, organic and ice-rich deposits also occur in unglaciated portions of Alaska, to the west of the Klondike. Examination of these deposits has been much more extensive than in Canada (e.g. Tuck 1940; Péwé 1952, 1989; Hamilton and Brigham-Grette 1991; Péwé et al. 1997), and hence there is a more developed understanding of the regional stratigraphy, depositional environments and paleoclimate. Recent thermoluminescence dating of tephra beds found within the sediments has enabled stratigraphic correlation between exposures (e.g., Berger 1987; Berger et al. 1996). Alaskan muck deposits have not been directly correlated with those of the Yukon despite the similarities noted by Fraser and Burn (1997).

2.4 VALLEY EVOLUTION

The development of physiography in Yukon has largely been controlled by regional tectonics and glaciation. The valley-in-valley profiles seen in the

Klondike area are shown schematically in Figure 2.2 (Bostock 1957; Hughes et al. 1972). During the initial Miocene ? uplift of the Yukon Plateau, the Yukon River likely flowed south via the Alsek drainage, towards the Pacific (Tempelman-Kluit 1980). Immature streams began dissecting the surface of the plateau, creating v-shaped valleys. Following downcutting, the valleys began widening, until stabilization of base level allowed accumulation of the White Channel Gravel. Tempelman-Kluit (1980) postulates that Pliocene uplift of the St. Elias Mountains, combined with damming by early (Pre-Reid) ice advances from the south, caused Yukon River to reverse its course to a northern route. In the Klondike area, the resulting rise in base level corresponds to the deposition of the Klondike Gravels on top of the White Channel sequence. Incision of these deposits to a level below the White Channel terraces occurred in the Pleistocene as a result of regional uplifts, such as that of the Swede Dome and King Solomon Dome within the Klondike itself (Hughes et al. 1972; Milner 1976). These regional uplifts have influenced drainage patterns, creating major NW-SE trending streams which have cut off underfit, shorter streams perpendicular to them (Milner 1976).

2.5 GLACIAL HISTORY

Most of central and northwestern Yukon escaped glaciation. This unglaciated terrain forms part of the Beringian refugium which spanned most of Alaska and eastern Siberia. As many as seven northwesterly advances of the Cordilleran ice sheet have been recorded in Yukon Territory (Duk-Rodkin et al.,

Figure 2.2: Development of valley-in-valley profiles (after Hughes et al. 1972; Bostock 1957). 1: Klondike Plateau erosion surface, 2: incision and valley-widening, 3: Deposition of White Channel Gravel, 4: Pleistocene incision.

1995). The first four advances, collectively termed Pre-Reid glaciations, were the most extensive, reaching as far west as the Dawson area (Bostock, 1966; Hughes, 1987). These glaciations are Tertiary to Early Pleistocene in age, with the last of the outwash (Klondike Gravel) being deposited at about 730 ka. The ensuing Middle Pleistocene Reid glaciation in central Yukon corresponds with the Happy Interval of unglaciated Beringia. This glaciation was less extensive, reaching to within 150 km of Dawson (Hughes et al. 1989). Milner (1976) suggests that loess began accumulating in the Klondike during this glacial interval, which ended by 190 ka, as suggested by Berger et al. (1996). Sand wedges observed along the Stewart River to the east of the Klondike indicate that permafrost conditions existed during this time (Hughes 1987). The Sangamonian-Early Wisconsinan Reid-McConnell interglacial in central Yukon (Boutellier Interval in Beringia) coincided with the development of the Diversion Creek Paleosol (Tarnocai et al. 1985). The interglacial was terminated by the McConnell glaciation at about 28 ka (Bostock 1966). In the Klondike, McConnell (Late Wisconsinan) deposits are the loess sequences described in section 2.3.2. Permafrost is known to have been present from this time at least, due to the preservation of ice wedges within these sediments (Fraser and Burn 1997; Naldrett 1981). Glacial maximum was reached around 18 ka in central Yukon (Jackson and Harington 1991), and periglacial conditions existed until 11.5 ka (Bond 1997). The narrowness of ice wedges developed on Reid and Pre-Reid surfaces in central Yukon has been suggested as indicative of a short and intense glacial period (Rutter et al. 1979). This most

recent glacial interval ended by about 10 ka at the start of the Holocene (Hughes et al. 1989). The Quaternary history of central and western Yukon Territory is summarized in Figure 2.3.

Figure 2.3: Summary of Quaternary history of unglaciated and glaciated Yukon Territory. After Bostock (1966), Lerbekmo et al. (1975), Hopkins (1982), Schweger and Matthews (1985), Heginbottom and Vincent (1986), Hughes (1989), Berger et al. (1996).

CHAPTER 3: BACKGROUND GEOCRYOLOGY

This chapter introduces the aspects of geocryology used in this study, primarily cryostratigraphy and analysis of the stable isotopes of water. The origin of various types of ice and the sedimentology of frozen sediments is discussed.

3.1 CRYOSTRATIGRAPHY

Cryostratigraphy involves the identification of cryostratigraphic units within frozen, unconsolidated sediments, based on the nature and distribution of ground ice and the interpreted locations of thaw contacts (e.g., French et al. 1982; Burn et al. 1986; French and Harry 1990; Murton and French 1994; Burn 1997). The description of cryostratigraphy includes identification of ice types, the structure of both ice and sediment, the relationship between ice and sediment and the nature of contacts between units.

The cryostratigraphy of sediments can be used to interpret both the depositional and freezing history of materials found in areas where permafrost has persisted since shortly after deposition. The identification of ice types and thaw contacts can yield important paleoclimatic information, specifically temperature and moisture conditions (e.g., Black 1983; Hyatt 1990; Murton and French 1994; Vasil'chuk and Vasil'chuk 1995; Burn 1997).

3.1.1 Ice Types

Ground ice includes all types found within frozen ground, excluding that

which has a buried origin (Permafrost Subcommittee 1988). Thus, any ice found within frozen sediments can be classified as buried ice or intrasedimental ice, which includes injection ice, pool ice, segregated ice, aggradational ice and ice wedges.

Buried Ice

A buried glacier-ice origin has been proposed for several thick, tabular bodies of ice observed in the western Canadian Arctic (e.g. Fujino et al. 1983; Lorrain and Demeur 1985; French and Harry 1990). In the unglaciated Klondike area, French and Pollard (1986) proposed that a large body of ice within frozen sediments could have been a buried snowbank, which was metamorphosed into massive ice. The massive ice had a dispersed sediment content and the orientation of ice crystals was oblique to the valley side. Ice of a buried origin may have bubbles at the upper ice-sediment contact which are truncated and infilled with sediment, indicating that the ice must have originated at the surface, since the ice must be older than the sediments that infill it (Mackay 1989; French and Harry 1990).

Injection Ice

This type of ice forms following the expulsion of water into permafrost from a proximal source. The water is usually under pressure and the frozen ground may deform as a result of the injection of the water and the growth of ice, creating a

mound. Characteristically, pingos form in this way (Mackay 1962; Mackay 1972). Pingo ice is often underlain by coarse-grained sediments. Frost mounds form by similar processes to open-system pingos, but are contained within the active layer, and as such, may disappear annually (Van Everdingen 1978, Pollard and French 1984). Other near-surface mounds in permafrost can be due to the growth of ice in cavities produced by methane gas. These types of mounds may last for several years (Mackay 1965). Intrusive ice may also occur as a tapering ice-dyke which penetrates upwards through frozen sediments (Mackay and Dallimore 1992).

Pool Ice

Ice wedge troughs often channel meltwater and lake water. If this water freezes, it forms pool ice. The ice is usually clear and has bubbles which radiate towards the center (Mackay 1997).

Segregated Ice

Segregated ice is a general term used for ice which forms by the addition of water to pore ice by a negative pressure created at the freezing front (Williams 1967). The result is a lens of ice, which may be from ten centimetres to several hundred metres long, and up to 30 m high, although ones only a few centimetres high are most common (e.g. French and Harry 1990; Mackay 1971). Typically, ice lenses are found within finer-grained sediments where the unfrozen water content is high and water can be continuously supplied to the growing ice lens (Penner

1967; Anderson and Tice 1972). However, this type of ice has also been observed within coarse-grained sediments (e.g. Mackay 1992). Large masses of segregated ice which form within mineral soil, create perennial frost mounds, called palsas (e.g., Pollard and French 1984; Seppälä 1986). Although it is sometimes difficult to differentiate between large masses of segregated ice and buried ice, the nature of the upper ice-sediment contact may provide some clues. Bubbles extending downwards from the contact, the presence of ice veins perpendicular to the contact, or an increase in abundance of ice lenses closer to the contact, may indicate a segregated origin for the ice body (Mackay and Dallimore 1992).

Aggradational Ice

This type of ice is ubiquitous in permafrost terrain. It is segregated ice which has formed within the active layer, but has since become part of permafrost by the upward aggradation of the permafrost table either by a decrease in the thickness of the active layer or by the addition of sediment to the surface, such as is the case in alluvial fans, deltas and sites of peat growth or eolian deposition (Mackay 1972). This type of ice, therefore, preserves the base of a former active layer which may truncate frozen sediments or other types of ground ice. Thick ground ice within the active layer may also develop due to repeated-segregation: a multi-annual process whereby the ice content at the top of permafrost gradually increases due to unequal migration of frozen water, self-purification of the segregated ice and the addition of material to the ground surface. It only occurs

in mostly fine-grained soils (Cheng 1983).

Ice Wedges

Ice wedges are found in many areas underlain by permafrost, and are diagnostic of its presence. Ice wedges join to form ice-wedge polygons, which are readily observable in lowland settings. On hillsides, polygons may become elongate, forming stripes. Wedges form by the infilling of thermal contraction cracks by spring snowmelt (Mackay 1983*a*). Dostovalov and Popov (1963) recognize two types of ice wedges; syngenetic and epigenetic, and describe the conditions for their growth based on rates of sedimentation. Epigenetic wedges are those which grow where there is no addition of material to the surface during ice wedge growth, creating v-shaped wedges which have upturned sediments at their margins (Mackay 1974). Syngenetic wedges develop where there is accumulation of material at the ground surface during ice wedge growth, creating narrow, elongate wedges, sometimes nested in chevron pattern, and often with rounded tops. Mackay (1990*a*) recognises a third type of ice wedge, the anti-syngenetic wedge, which grows during a lowering of the ground surface by removal of material, creating very wide wedges (Figure 3.1). The frequency of ice wedge cracking increases with decreasing ground temperatures (Romanovskij 1973), and decreasing snow cover (Mackay 1974).

Ice wedges are truncated at their top by the base of the active layer. If there is a decrease in active-layer thickness, or an accumulation of material, the

Figure 3.1: The development of epigenetic, syngenetic and anti-syngenetic ice wedges. The tops of the wedges are truncated by thaw, and indicate maximum depth of the active layer at the time. From Mackay (1990a).

truncated portion will indicate the position of the base of the old active layer. Since the initiation of cracking is at the top of the ice wedge (Mackay 1984), if the ice wedge is rejuvenated after some period of time, it will likely re-crack along the same line as the older wedge (Mackay 1995, 1990*a*). Since ice wedges grow downward, perpendicular to the ground surface (Dostovalov and Popov 1963), slope movement can cause the deformation of these wedges both during and after cessation of growth (e.g., Hyatt 1990; Burn 1997).

Wedges formed by thermal contraction of the ground need not necessarily be filled by ice. Several authors have identified wedges along the coast of the Beaufort Sea which are filled entirely with eolian sand, rather than ice (e.g., Carter 1983; Murton and French 1993), and ones that are filled with sand and ice (e.g., Burn 1997). The identification of these types of wedges can be of great paleoenvironmental significance. If permafrost has persisted since the time of formation of these wedges, then a cold, arid environment with minimal snowcover was likely to exist (Murton and French 1993). Sand-filled wedges can also be observed where permafrost no longer exists (e.g., Brown and Péwé 1973). In this case, the sand infill is secondary, and the wedges are considered "fossil" (Black 1976).

As an ice wedge grows, and volume is added to permafrost, the ground on either side will be displaced upwards, creating a trough immediately above the ice wedge where water will collect in the summer (e.g., Mackay 1974). If there is no deepening of the active layer, then this water does not induce melting of the ice

wedge, and is supplied to growing ice lenses within the active layer at the sides of the trough during fall freeze-back (Mackay 1974). Melting and erosion of an ice wedge begins when standing water drains, exposing the ice wedge, surface temperature increases, or there is a disturbance of the insulative vegetative cover, causing a deepening of the active layer, which now encroaches on the top of the wedge. Channels formed along ice-wedge troughs by melting and erosion may fill with sediments, as well as other types of ice (e.g., Mackay 1997).

3.1.2 Structure

The distribution of ground ice within frozen sediments defines the cryostructure (Murton and French 1994). There are numerous cryostructural classification systems (e.g., Shumskii 1964; Katasonov 1969; Demek 1978; Kudriavtsev 1978; ACGR 1988). The simplest and most comprehensive classification system however, is that of Murton and French (1994). This system is based on other classification systems and includes structures observed in ice-free sediments as described by Collinson and Thomson (1982). According to this system, sediments and ice can be divided into cryofacies based on their volumetric ice content. Five facies are recognized: ice-poor sediment (≤ 25 %), ice-rich sediment (> 25 % to ≤ 50 %), sediment-rich ice (> 50 % to ≤ 75 %), sediment-poor ice (> 75 %), and pure ice (100 %). Ice lenses can be described by their shape as planar, wavy or reticulate.

3.1.3 Contacts

Recognizing discontinuities within the cryostratigraphic record is of utmost importance as they yield valuable paleoclimatic information. Two types of cryostratigraphic contacts are recognized: the unconformity and the conformity. An unconformity denotes the maximum depth of thaw, or an erosional surface. A thaw unconformity is located at the base of a paleoactive layer, where ice wedges (or other types of ice) are truncated, or aggradational ice exists (e.g. French et al. 1982; Burn et al. 1986; Murton and French 1994). Numerous periods of thaw have been identified by the truncation of several generations of ice wedges (e.g. French et al. 1982, Black 1983, Hyatt 1990). Radiocarbon dating of materials immediately above these contacts can give a minimum age of the occurrence of the thaw (e.g. Burn et al. 1986; Mackay 1995; Burn 1997). *In situ* organic material such as rhizomes give more accurate dates than wood pieces which can be cryoturbated into position. The recognition of thaw unconformities is important in that it indicates that there has either been an accumulation of material or a decrease in active-layer thickness since the time of maximum thaw, and that the permafrost table has since moved upwards. An erosional unconformity is produced by the truncation of ice by exposure, such as in a thaw slump (e.g. Mackay and Dallimore 1992).

Conformities are ice-sediment contacts produced where there has been downward freezing. The ice in question is usually segregated in origin, and has bubbles extending downwards from the overlying sediment, indicating freezing

from above (Mackay 1989).

3.1.4 Summary

The identification of ice types, ice-sediment relationships and the identification of thaw unconformities are fundamental to the study of cryostratigraphy. Valuable paleoclimatic information can be extracted if the genesis of the ice is considered. Placing ages on thaw unconformities likewise identifies periods of change. In the Klondike area, the cryostratigraphy of a frozen section exposed on Hunker Creek has been examined by French and Pollard (1986). Six cryostratigraphic units were identified based on ice types and ice-sediment relationships. The ice types consisted of two generations of large ice wedges, smaller lenses of segregated ice, and a larger body of ice near the base of the section (unit 4), which remains unidentified, but is proposed to be either of buried snowbank or segregated origin. Fraser and Burn (1997) have recognized large epigenetic and syngenetic ice wedges within the upper portion of the Silt unit, and smaller ice wedges and lenses within the Organic unit (see section 2.3.2).

3.2 H-O ISOTOPES

The analysis of the heavy isotopes of hydrogen and oxygen is also useful in geocryology. This section examines the theory behind the fractionation of the isotopes of water, and the utility of analyzing these isotopes in determining the origin of ice within sediments and the existing conditions during its formation.

The most common isotopes of water are ^{16}O and ^1H . Four other isotopes (^{18}O , ^{17}O , ^3H and ^2H) exist, but are much less abundant. Standard ocean water (VSMOW) has an $^{18}\text{O}/^{16}\text{O}$ ratio of 2005.2×10^{-6} and $^2\text{H}/^1\text{H}$ ratio of 155.76×10^{-6} . Changes in the abundances of ^{18}O and ^2H with respect to these standards are expressed as $\delta^{18}\text{O}$ and $\delta^2\text{H}$. The most important control on the fractionation of water is temperature (Craig 1961).

3.2.1 Fractionation and the Meteoric Water Line

If a body of water is left to evaporate, the lighter isotopes ^1H and ^{16}O will preferentially enter the vapour phase, thereby enriching the remaining water in the heavier isotopes, ^2H and ^{18}O . Upon condensation, the opposite occurs, the heavier isotopes are preferentially removed, and the residual vapour is depleted. Global precipitation follows a Rayleigh distillation process whereby the system is closed and fractionation progresses until all heavy isotopes have been removed from the vapour (Kirshenbaum 1951; Craig 1961). This means that precipitation that is the first to form from seawater is enriched in the heavier isotopes, and that which is last to form is significantly depleted (Craig 1961; Dansgaard 1964). The relation between $\delta^2\text{H}$ and $\delta^{18}\text{O}$ in most global precipitation is linear, and can be expressed as:

$$\delta^{18}\text{O} = 8 \delta^2\text{H} + 10 \text{‰} \quad (\text{Craig 1961}) \quad [3.1]$$

It is referred to as the meteoric water line. Water from warmer regions is enriched in the heavier isotopes whereas it is depleted in colder regions. Michel (1982)

demonstrated experimentally that equilibrium fractionation can occur during slow freezing. On the other hand, water will retain its isotopic composition when melted from ice (Jouzel and Souchez 1982).

3.2.2 ^{18}O and ^2H in Permafrost

There are numerous isotopic analyses of ice from within frozen sediments in Yukon Territory (e.g. Michel 1982, 1983; Michel and Fritz 1982). In sampling pore ice at regular intervals across stratigraphic sections, isotopic discontinuities have sometimes been revealed (e.g., Michel and Fritz 1982; Mackay 1983; Michel 1983; Burn et al. 1986).

Dansgaard (1964) demonstrated that for global precipitation, $\delta^{18}\text{O}$ decreases with decreasing mean annual surface temperature, by the relation;

$$\delta^{18}\text{O} = 0.69T - 13.6 \text{ ‰} \quad (\text{Dansgaard 1964}) \quad [3.2]$$

Although this relation should not be used directly, it is sufficient to note that a decrease in surface temperature results in a decrease in $\delta^{18}\text{O}$ of precipitation. The $\delta^{18}\text{O}$ -temperature relationship seems to work well for ground ice as well, even though it may be derived from precipitation whose $\delta^{18}\text{O}$ may vary considerably seasonally and annually. The isotopic discontinuities revealed within frozen sediments have been related to long-term temperature changes associated with glacial/non-glacial intervals. Ice within Late Wisconsinan sediments has $\delta^{18}\text{O}$ values which are more negative (av. -27 ‰) than those of modern ice (av. -21 ‰), indicating that temperatures at that time were colder than at present (e.g. Michel

1982; Michel and Fritz 1982). Nikolayev and Mikhalev (1995) have used samples from Siberia to correlate $\delta^{18}\text{O}$ values with surface temperatures at the time of formation of ground ice (pore and vein ice). They detected depletions of between 0.83 and 0.86 ‰ per degree drop in mean temperature of the cold season and depletions of 0.5 to 0.6 ‰ per degree drop in mean January temperature.

Analysis of the isotopic composition of ice wedges yields similar trends of depletion within older wedges (Michel 1990). Some uncertainty exists when sampling ice wedges however, since as discussed in section 3.1.1, cracks may penetrate older wedges and erosion and incorporation of other types of ice may lead to sampling of ice which was not derived from snow meltwater (e.g. Mackay 1997). Isotopic variations of up to 3 ‰ within the wedges themselves have been detected (Michel 1990). Michel also noted that the composition of ice wedges can be significantly different from enclosing sediments, because ice wedges can develop when surface conditions are favourable, and, therefore, can be much younger than the sediments they penetrate. Despite this, sediments within 0.5 m of the ice wedge may show values closer to those of the wedge, as cracking is not always confined to the wedge, and erosion of wedges by surface waters causes seepage into surrounding sediments (Michel 1990).

A hypsithermal thaw unconformity (8765 ± 230 BP, GX-4352; Mackay 1983a) is widely observed in the western Canadian Arctic. At Illisarvik, Richards Island, ice wedges above this unconformity provided $\delta^{18}\text{O}$ ratios between -22 and -26 ‰ (Michel and Fritz 1982), whereas ice from a wedge below this unconformity

near Tuktoyaktuk, had a lower ratio of -32 ‰ (Mackay 1983*a*). The isotopic composition of ice wedges in relation to the meteoric water line has been studied by Lauriol et al. (1995). Late Holocene ice wedges plot below the meteoric water line. They suggest that this indicates that ice wedge infill is meltwater from a later stage of snowmelt, which has been subjected to refreezing, rather than the first meltwater formed, which, derived from winter precipitation, is depleted in both ^{18}O and ^2H .

It is apparent that the location of thaw unconformities caused by changing climate may be confirmed by isotopic analyses, but this is not always the case. Sampling across unconformities at Illisarvik (Michel and Fritz 1982), and on Pelly Island, Northwest Territories, Burn (1997) revealed no discontinuity due to mixing of surface and permafrost waters (Mackay 1983*b*).

3.2.3 Summary

Isotopic analysis of ice can be useful if the origin and age of the ice in question is to be determined. Sampling of pore ice can give insight into the regional temperature at the time of formation, if no subsequent mixing has occurred. Isotopic ratios from ice wedges, likewise, can provide insight into surface conditions at the time of formation of the ice wedge, again provided that the ice sampled is known to be derived from surface infiltration, rather than erosion and subsequent refreezing. If a thaw unconformity is suspected, analysis of isotopes across this contact, may confirm its location, and provide clues to

conditions during deposition, with the same assumptions as stated above.

3.3 CHAPTER SUMMARY

The description and identification of ground ice types, ice-sediment relations and contacts, and analysis of hydrogen and oxygen isotopes has not previously been attempted on a regional basis in the Klondike area. The literature reviewed above forms the background for study of the cryostratigraphic and isotopic characteristics of muck deposits.

CHAPTER 4: METHODS

Examination of the geocryology of the unconsolidated valley-bottom sediments of the Klondike area was accomplished by detailed observations of the cryostratigraphy and analysis of the isotope geochemistry of the ice. This chapter reviews both the field and laboratory methodology used. Section 4.1 reviews the study area, including site selection and description. Methodology used in describing the cryostratigraphy is provided in section 4.2, and section 4.3 outlines the field and laboratory procedures used in the isotope geochemistry.

4.1 STUDY AREA

Five drainage basins in the region were included in this study. They comprise the north-flowing Bonanza and Hunker Creeks, tributaries of Klondike River, and the south-flowing Dominion, Sulphur and Quartz Creeks, tributaries of Indian River. They are separated by an east-west-trending drainage divide topped by King Solomon Dome, at an elevation of 1306 m.

Placer mining has occurred in the area since the late nineteenth century. Present mining operations require the removal of the ice-rich overburden materials to reach the underlying gold-bearing gravels. High-pressure water is directed at frozen sediments to thaw and remove them (Figure 4.1). The resulting exposures are vertical to overhanging cliffs 10 to 300 m wide which reveal materials from the surface to bedrock. Once exposed, these materials thaw quickly and slump naturally until the slope restabilizes, often within six years. Therefore, study sites

Figure 4.1: High-pressure water is directed at frozen sediments to thaw and remove them. Gold Bottom Creek, August 1997.

were chosen where mining was active, or had been abandoned for only a few years. Nineteen such sites were chosen, all accessible by road (Figure 4.2, Table 4.1).

4.2 CRYOSTRATIGRAPHY

As outlined in section 3.1, cryostratigraphy involves the description of, and the relations between, ice and sediment. In keeping with the objectives of this study, this section is divided into two sub-sections; one devoted to discussing the methodology in re-examining the stratigraphy from a sedimentological perspective, the other describing the methods in examining the ice and its relations with these sediments.

4.2.1 Stratigraphy

Each exposure was referenced with a Global Positioning System, precise to approximately 50 m. The aspect of the mining cut was recorded, as well as a general description of the morphology of the area, such as the relief and location of the mining cut in the valley. Units were identified based on their colour, texture, structure and the nature of organic material. Where units were not accessible, measurements of thickness and sediment characteristics were made visually. Contacts were described as either conformable or unconformable, and bedding was defined by changes in organic content or competency of the sediments, largely a function of ice content. Detailed particle-size investigations have already been undertaken (Fraser and Burn 1997), so the textural

Figure 4.2: Study and drilling locations within Klondike area. Exact coordinates are given in Table 4.1.

NAME	UTM N	UTM E	LOCATION
BC-1	7096 690	580 178	left limit Bonanza Creek, upstream of Sourdough Gulch
BC-2	7090 330	582 009	left limit Bonanza Creek at Eldorado Creek
IG-1	7087 703	582 529	right limit Irish Gulch, 500 m upstream of Eldorado Creek
UBC-1	7088 937	584 204	left limit Upper Bonanza Creek, across from Spring Gulch
RB-1	7085 932	591 438	right limit Ready Bullion Ck, 1 km upstream of Upper Bonanza Ck.
CAMP-1	7093 299	581 424	right limit Bonanza Creek, between Queen and Mosquito Gulches
DG-1	7099 903	591 575	right limit, base of Dago Gulch
DG-2	7099 830	591 379	left limit Hunker Creek, immediately below Dago Gulch
LCC-1	7095 649	591 200	right limit of Right Fork of Last Chance Creek
FV-1	7094 645	599 064	left limit Hunker Creek below Temperance Hill
GB-1	7087 331	598 664	right limit Gold Bottom Creek, 1 km above Gold Bottom Gulch
GB-2	7086 820	598 830	right limit Gold Bottom Creek, 1.5 km above Gold Bottom Gulch
DP-1	7082 557	604 576	right limit Discovery Pup at Dominion Creek
DC-1	7079 961	610 572	right limit Dominion Creek, 1 km above Coarse Gold Pup
DC-2	7080 060	611 135	right limit Dominion Creek, 1.5 km above Coarse Gold Pup
GR-1	7065 812	618 140	left limit Gold Run Creek, 1 km above Dominion Creek
SC-1	7075 123	602 251	right limit Sulphur Creek at Toronto Creek
SC-2	7069 424	606 144	right limit Sulphur Creek, across from Brimstone Gulch
QC-1	7081 171	597 158	left limit Quartz Creek, 1.5 km upstream of Mack Fork

Table 4.1: Section locations in Klondike area. UTM zone 7V.

characteristics of the sediments were estimated visually.

Two locations were drilled by water jet to determine the depth of sediments up valley sides (Figure 4.2). Selection of these sites was limited due to the availability of water, and the requirement of a settling pond for discharge. Lines were surveyed perpendicular to the valley bottom, using the position of the current creek elevation as datum. These lines were extended to the anticipated end of the deposit. Locations were marked every 15 m, and elevations were recorded to the nearest 1 cm. A Wajax pump and 2 inch hose were used to pump water through 3/4 inch drill pipe in order to thaw the underlying sediments (Figure 4.3). Depth was recorded when a gravel interface was encountered.

Organic material was collected at the base of units where possible for radiocarbon dating. Samples were extracted frozen from the sediment, described briefly, wrapped in aluminum foil, and sealed in plastic bags. Samples were stored in a cool, dry container and transported to Ottawa before analysis. Accelerator mass spectrometer ^{14}C dates were obtained from the University of Toronto, and conventional radiocarbon dates from Brock University.

4.2.2 Ice Characteristics and Content

Ice within the sediments was classified into two broad groups: bulk ice and pore ice. This section describes the methods used in identifying ice types and determining ice contents of the sediments both in the field and in the laboratory.

Figure 4.3: Water-jet drilling on Gold Bottom Creek. Water is being taken from a settling pond in the valley below.

4.2.2.1 Field Determinations

Bulk Ice

Visible ice bodies within sections were identified based on physical characteristics such as shape, colour, translucency, texture, bubble distribution and their relations with surrounding sediments, particularly the nature of contacts. Each exposure was sketched from a distance on grid paper for visual estimation of bulk ice distribution.

Pore Ice

Over 100 samples of frozen sediment and ice were collected from exposures for ice content determinations. Two samples were extracted from within each stratigraphic unit accessible within each exposure, using a chisel and hammer. The frozen sample was immediately wrapped in plastic wrap and immersed in a full bucket of water. The overflow was captured in a graduated plastic container, accurate to 20 mL. This volume was recorded as the volume of the sample extracted. For ice samples collected, the bubble density was visually estimated. The sample was immediately sealed in double plastic bags to minimize water loss due to evaporation. Samples were transported out of the field once per week for weighing and drying.

4.2.2.2 Laboratory Determinations

The expressions used in calculations for volumetric ice content are presented

in Table 4.2. Ice content determinations were completed at J.V. Clark School in Mayo. The volume of ice within samples was determined using the following principles.

For saturated sediment;

$$M_t = M_{sat} = M_{pw} + M_d \quad [4.1]$$

therefore;

$$M_{pw} = M_{sat} - M_d \quad [4.2]$$

If it is assumed that pore ice is bubble-free and unfrozen water content negligible, then;

$$V_{pi} = M_{pw} / \rho_i \quad [4.3]$$

The volumetric proportion of pore ice within a saturated sample, is then;

$$V_i = (V_{pi} / V_t) \times 100 \% \quad [4.4]$$

If a sediment is supersaturated, then there is excess water, so that;

$$M_t = M_{sat} + M_{ew} \quad [4.5]$$

rearranging;

$$M_{ew} = M_t - M_{sat} \quad [4.6]$$

Again, knowing the density of ice, and for a moment discounting bubble content;

$$V_{ei} = M_{ew} / \rho_i \quad [4.7]$$

If a supersaturated frozen sample is considered, then;

$$V_t = V_{pi} + V_{ei} + V_{sed} \quad [4.8]$$

Since we are only concerned with the amount of ice in a sample;

$$\% \text{ ice} = [(V_{pi} + V_{ei}) / V_t] \times 100 \% \quad [4.9]$$

Nomenclature			
	symbol	description	unit
major	V	volume	mL, cm
	M	mass	g
	ρ	density	g/cm^3
subscript	w	water	
	pw	porewater	saturated amount
	ew	excess water	supersaturated
	i	ice	
	pi	pore ice	
	ei	excess ice	
	sat	saturated sediment	
	t	total, as extracted	
	d	dry sediment	
	α	bubble density	0 to 1.0

Table 4.2: Nomenclature used in ice content determinations.

For most ice bodies, the situation is complicated further by the presence of bubbles within the ice. To account for this, equation [4.9] is modified to include α ;

$$\% \text{ ice} = \{ [V_{pi} + V_{ei}(1 - \alpha)] / V_t \} \times 100 \% \quad [4.10]$$

V_t and α were determined in the field, as discussed above; M_t was determined by weighing the thawed sample, together with its bags and then subtracting the mass of the bags. To estimate M_{sat} , the sample was transferred to paint tins or aluminum trays, using a squirt bottle to ensure that all sediment was removed from the bags. The excess water in the trays was poured off into a paint tin where the suspended sediment was allowed to settle overnight. Once clear, the water was siphoned off, leaving only saturated sediment; the mass of the saturated sediment was then obtained. The sample was then placed in a 105°C oven and allowed to dry. M_d was obtained by weighing the dry sample.

Masses of all containers were obtained using an electronic balance, precise to 0.1 g. All other measurements are precise to 1 g, so that if a maximum of 3 separate containers were used for a sample, the precision of masses for each set of measurements can be considered as ± 3.3 g. Since each sample required weighing for M_t , M_{sat} and M_d , the precision is now ± 9.9 g, i.e., ± 10 g. Sample M_t was, on average, 600 to 700 g, so error is approximately 2 to 3 %. Using these measurements, all ice contents were determined using equations [4.4], [4.9] and [4.10].

4.3 ISOTOPE GEOCHEMISTRY

Fifty-two ice samples were collected from the studied sections for $^{18}\text{O}/^{16}\text{O}$ and $^2\text{H}/^1\text{H}$ isotopic ratios. Three categories of ice were sampled: relatively sediment-free ice from ice wedges, massive (undifferentiated) ice, and pore ice from frozen sediments. The field procedures for sampling are outlined in section 4.3.1 and the laboratory methods are described in section 4.3.2.

4.3.1 Field Methods

In addition to sampling representative ice types exposed at each section, one section was sampled to obtain a profile across a prominent contact. Access was a key problem in all sampling if contamination due to surface meltwater was to be avoided. The surface surrounding the sampling site was first dried with paper towels, then several were used to build a 'dam' above the sample site to absorb and divert meltwater from the face above. Using a hammer and a dry 1" chisel, frozen material was removed to a depth of at least 2 cm, from a 15 cm x 15 cm area. The chisel was cleaned and dried again, and paper towels replaced as necessary to ensure that the area remained free of meltwater. The sample was extracted by chiselling off small pieces into a plastic bag. This bag was sealed, removing as much air as possible, and placed and sealed within a second bag. Samples were transferred to 7 mL plastic, screw-top vials for storage at the end of every day. This involved cleaning, drying and cutting one corner of the bag, so that meltwater could be transferred to the vial. Vials were filled to the top, closed, and sealed

with tape. Ice-poor samples were transferred to larger vials to ensure that enough water would be available for isotopic analysis.

4.3.2 Laboratory Methods

Samples were analysed for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ at the G.G. Hatch Isotope Laboratory at the University of Ottawa. Determinations of the $^{18}\text{O}/^{16}\text{O}$ ratio were by automated procedures. 1 mL of sample was transferred to a glass vial attached to the 20-sample extraction line. Oxygen was extracted by equilibration with CO_2 . Three distilled water standards were analysed for every seven samples. The extraction of hydrogen was done by reduction with zinc, following procedures of the G.G. Hatch Isotope Laboratory. Both $\delta^{18}\text{O}/^{16}\text{O}$ and $\delta^2\text{H}/^1\text{H}$ were determined using a gas-source mass spectrometer, with values determined from an average of 6 pairings of sample to reference gas such that;

$$\delta \text{ av. R} = [(\text{av. R}_{\text{sample}} / \text{av. R}_{\text{reference}}) - 1] \times 1000 \text{ ‰ ref.} \quad [4.11]$$

(Gonfiantini 1981), where R represents the mass ratio of heavy to light isotope. The average ratios for standards were corrected to VSMOW by creating a linear regression where:

$$\delta \text{ R}_{\text{VSMOW}} = m(\delta \text{ av. R}) + b \quad [4.12]$$

This equation was then used to correct all O and H sample ratios. $\delta^{18}\text{O}$ are precise to $\pm 0.2 \text{ ‰}$ and $\delta^2\text{H}$ to $\pm 2 \text{ ‰}$.

CHAPTER 5: RESULTS

This chapter contains the results of field measurements and laboratory studies of samples collected from 19 sections in the Klondike area. First, the setting of the deposits is described, followed by detailed observations of the sedimentological and geocryological characteristics of the deposits. The chapter concludes with some laboratory results of the isotopic analyses.

5.1 DISTRIBUTION OF DEPOSITS

Muck is found on north or north-east facing slopes throughout the Klondike and in valleys of small tributaries where mining operations have not removed the surficial materials. The thickness of the deposits generally increases southwards in the area, so that 1 to 10 m of exposure is common on lower Bonanza and Hunker Creeks, whereas the thickness exceeds 20 m on Quartz and Dominion Creeks.

Two locations were drilled to determine muck thickness up valley-sides. At one, Upper Bonanza Creek, large amounts of colluvium were encountered, rendering it impossible to determine the depth to the base of muck. The results of the drilling on Gold Bottom Creek are shown in Figure 5.1. A 95 m transect, perpendicular to the valley-bottom, on the east side of Gold Bottom Creek, was drilled approximately every 15 m beyond the extent of disturbed sediments. The sediment encountered was olive-brown 2.5Y 5/3 to 4/3 ice-rich silt, with some pockets of pebbly sand. There was no section exposed to examine the stratigraphy of the sediments. Figure 5.1 shows the interpreted valley morphology based on these drilling results. Although

Figure 5.1: Drilling results from Gold Bottom Creek. Dashed line indicates that no bottom was reached. At hole 3, at least 15 m of fine material was encountered. Gravel was encountered at the base of holes 1 and 4. This may be creek gravel or colluvium.

incomplete, it shows that fine-grained sediments extend well up the valley sides and that modern valley morphology may be significantly different from that at the time of gravel deposition and/or reworking, prior to muck emplacement.

5.2 CRYOSTRATIGRAPHY

Discussion of the cryostratigraphy of the Klondike deposits is presented in three sections. The first is devoted to defining and describing recognized units, based on the sedimentology only. The timing of deposition of these units is confirmed by the results of radiocarbon dating. Sections 5.2.2 and 5.2.3 describe the ice characteristics and present results of ice content determinations. A detailed cryostratigraphy of most sections can be found in Appendix I.

5.2.1 Stratigraphy

Four units have been recognized within these materials. There are no formal names for any of the sedimentary units overlying gravels in the Klondike area, and therefore, informal names are used here to provide an easy means of referring to units which have been previously defined by their sedimentological characteristics (e.g., Fraser and Burn 1997). Fraser and Burn (1997) recognized two units within these deposits, but the stratigraphy described here differs in two important ways. First, the Silt unit has been divided into two separate units; second, an additional unit has been recognized underlying the Silt unit. The description of the newly designated Last Chance Creek Formation, Quartz Creek Formation and Dago Hill Formation will

proceed from oldest to youngest. The Organic unit of Fraser and Burn (1997) is not discussed in detail since their observations do not differ from those obtained in this study.

Last Chance Creek Formation

This sedimentary unit is only exposed at one location, LCC-1, and is, therefore the type section. It can be divided into two distinct facies: The lower 0.75 to 1 m consists of dark olive grey (5Y 3/2) massive silt with rhizomes and the upper 0.5 to 0.75 m is a dark brown (7.5YR 3/2) cryoturbated bed, consisting of silt from the facies below and organic matter, particularly twigs and small branches (Figure 5.2). The contact between the two facies is distinct and recognizable primarily by the presence of larger organic matter and the cryoturbated texture. The contact with the underlying gravel is irregular and sharp. Rhizomes collected from the base of the unit, directly above gravel have been dated to $45\,500 \pm 5\,800$ years BP (BGS-2019) and wood collected from the cryoturbated facies has been dated to $40\,060 \pm 2\,800$ years BP (BGS 2018), (see Table 5.1). Since both dates are at or beyond the limit of conventional radiocarbon dating, they should be considered as $> 40\,000$ ^{14}C years BP.

Quartz Creek Formation

This unit forms the lowermost portion of the Silt unit defined by Fraser and Burn (1997). It is found overlying the Last Chance Creek Formation, or above gravel. It has been observed up to 10 m thick, but may be as much as 17 m thick if

Figure 5.2: Last Chance Creek Fm: ice-rich silt is overlain by organic-rich cryoturbated bed. The wedge formed syngenetically with deposition. The age of this unit is >40 000 BP. Scale = 1.5 m.

Field No.	Lab. No.	type	radiocarbon age	location	description
DG-O-01	Beta-111606	radiometric	13 910 +/- 70	DG-1	squirrel nest
DG2-Org-1	TO-6869	AMS	11 620 +/- 90	DG-2	twigs from base of Dago Hill Fm
DC2-Org-1	BGS 2015	radiometric	49 000 +/- 9 000	DC-2	wood above gravel
DC2-Org-2	BGS 2016	radiometric	10 125 +/- 380	DC-2	twigs and peat above first generation ice wedges
DC2-Org-3	BGS 2017	radiometric	10 178 +/- 200	DC-2	wood fragments above second generation ice wedges
LCC-Org-1	BGS 2018	radiometric	40 060 +/- 2 800	LCC-1	wood from cryoturbated bed, Last Chance Creek Fm
LCC2-Grass-1	BGS 2019	radiometric	45 500 +/- 5 800	LCC-1	rhizomes above gravel, Last Chance Creek Fm
RB-Org-bT	TO-6967	AMS	22 300 +/- 190	RB-1	twigs above tephra bed, Quartz Creek Fm
RB-Org-aT	TO-6968	AMS	23 520 +/- 210	RB-1	twigs below tephra bed, Quartz Creek Fm

Table 5.1: Results of radiocarbon dating.

material drilled on Gold Bottom Creek was entirely of this unit. The material is 87% by weight medium silt to fine sand (Fraser and Burn 1997), and is olive brown to grey brown 2.5Y 4/4 to 10YR 5/2. At QC-1, a colour gradation to dark brown (7.5YR 3/2) was detected in the upper 1 m of the unit. There are some silty beds, 10 to 20 cm thick (Figure 5.3), which contain rhizomes and small twigs and have been interpreted as colluvial in origin (Fraser and Burn 1997). Several 1 to 10 cm pebbly sand beds were observed in the lower portions of the unit. Mammal bones were also observed at various levels, and are of types described by Harington and Clulow (1973). A discontinuous to continuous 1 to 30 cm tephra bed was seen close to the base of this unit (Figure 5.3). Organic material sampled above and below this bed at RB-1 has been dated to $23\,520 \pm 210$ (TO-6967) and $22\,300 \pm 190$ (TO-6968) years BP, respectively, the latter representing the oldest possible date for the deposition of this tephra. This tephra has not been reported in other parts of the Yukon, but a conventional date of $24\,025 \pm 550$ BP (BGS-1755) was obtained for organic material underlying a tephra bed reported by Fraser and Burn (1997). The Quartz Creek Formation is absent from the broad valleys of Dominion, Sulphur, Hunker, lower Quartz and Bonanza Creeks.

Dago Hill Formation

This is the redeposited (upper) portion of the Silt unit of Fraser and Burn (1997). The differences in sedimentological and geocryological characteristics from the Quartz Creek Formation, as well as its timing of deposition warrant its

Figure 5.3: Tephra bed within Quartz Creek Fm at RB-1. Organic samples collected above and below this bed give a radiocarbon date of around 23 000 BP for the deposition of this tephra. Note the bedded nature of the sediments.

designation as a separate unit. It consists of interbedded, redeposited silt derived from the Quartz Creek Formation, and organic material such as peat, twigs and small branches. The contact with the underlying unit is planar and sharp (Figure 5.4). The base of this unit has an accumulation of organic material, making it more competent than the unit below. Although interbedded with what appears to be silt from the unit below, the sediments are very dark brown to black (7.5YR2.5/1), distinguishing them from the Quartz Creek Formation. The unit is up to 12 m thick, and, in places, rests unconformably on top of gravels. There is evidence of intense colluviation, shown by folded beds (Figure 5.5) and fluvial reworking, shown by numerous sandy lenses throughout the unit at several locations. A 1 to 2 m angular diamicton unit is present at several locations, near the base of this unit, and is likely a local colluvial deposit. Wood collected from the base of this unit at DG-2 has been dated to $11\,620 \pm 90$ years BP (TO-6869), and material near to the base of this unit, where it overlies creek gravel (DC-2), has been dated to $10\,125 \pm 380$ years BP (BGS 2016, Figure 5.6). Wood taken directly above gravel at this location was dated to $>40\,000$ years BP (BGS 2015), but is interpreted as a remnant, rather than being indicative of the timing of emplacement of this unit.

Organic Unit

This unit is 1 to 3 m thick, and consists of disorganized organic material, as described by Fraser and Burn (1997). The observations made in this study do not differ significantly, so the description here is abbreviated. The lower contact with

Figure 5.4: Contact between Quartz Creek Fm (below) and Dago Hill Fm. Note colour change towards the top of the unit. Organic material from the base of the Dago Hill Fm has been dated to 11 620 +/- 90 BP.

Figure 5.5: Folded sediments within Dago Hill Fm, suggestive of intense colluviation. Red of scale is 25 cm.

Figure 5.6: Location of samples collected for 14-C dating at DC-2, 1 m above gravel (BGS-2016: 10 125 +/- 380; BGS-2017: 10 178 +/- 200 years BP). The syngenetic ice wedge indicates that there has been an accumulation of material at this location. Scale increments are 5 cm.

the Dago Hill Formation is unconformable, irregular and sharp. Basal dates reported for *in situ* material, by Fraser and Burn (1997) suggest accumulation beginning by 9945 ± 150 BP; BGS-1772). The unit is entirely of Holocene age.

5.2.2 Ice Characteristics

This section describes the ice present within the four units discussed above and its relations with the surrounding sediments. Again, the ice characteristics have been described previously by Fraser and Burn (1997) and French and Pollard (1986). Keeping with the aims of this study, particular attention is focussed on the relations of ice and sediment, and the definition of units based on ice and sediment characteristics, rather than strictly sedimentological characteristics, the difference being that the identification of thaw unconformities is important, in addition to erosional or non-depositional unconformities.

Last Chance Creek Formation

It is perhaps the relative abundance of bulk ice within this unit that best distinguishes it from the overlying Quartz Creek Formation. Discontinuous, planar ice lenses, up to 20 cm long are abundant. Narrow syngenetic or epigenetic ice wedges, up to 4 m in height, penetrate the underlying gravel (Figure 5.2). They have a distinct, nearly vertical foliation. The sediments adjacent to their margins are upturned and the contacts are sharp. These ice wedges show at least two episodes of growth, indicating that there has been accumulation of material or a decrease in the

active-layer thickness since the start of their growth. Doming of the centers of these ice wedges has displaced sediments immediately above them, and is likely due to diapiric uplift (see for example Mackay 1995). The wedges are truncated by the base of the cryoturbated bed discussed above (Figure 5.7).

Quartz Creek Formation

No ice wedges were observed within this unit, where it is overlain by the Dago Hill Fm. At locations where the Dago Hill Fm was absent however, ice wedges were observed near the top of this unit. In these cases, the ice wedges are likely epigenetic, post-dating the deposition of the Quartz Creek Fm, since no ice wedges were observed within this unit where there has been subsequent deposition (i.e., Dago Hill Fm). Lens ice was observed within the Quartz Creek Fm, but was much less abundant than within the Last Chance Creek Fm. Most ice within this deposit is pore ice. The lower contact of this unit, cryostratigraphically, is a thaw unconformity, located at the truncation of the underlying ice wedges, at the base of the cryoturbated bed of the Last Chance Creek Fm (Figure 5.7), and sedimentologically, is an erosional unconformity above gravels where the Last Chance Creek Fm is absent.

Dago Hill Formation

What is perhaps most striking about this unit, is the abundance of ice. Ice wedges, up to 8 m in height are ubiquitous. Their shape varies depending on location in the valley. In valley bottoms, such as at DC-1,2, DG-1, SC-1,2, FV-1, GB1,2,

Figure 5.7: LCC-1 section showing interpreted cryostratigraphy. Thaw unconformities are dashed in yellow. Note diapiric uplift of lowermost ice wedges.

they are of epigenetic or syngenetic type, i.e., either they are very narrow, or form chevrons indicating accumulation of material (e.g., Mackay 1974), (Figure 5.6). On valley sides (all other sections), the ice wedges have a much broader form, where the tops of adjacent wedges are joined. These wedges are interpreted as of the anti-syngenetic type (Figure 5.7). They are identified as ice wedges because of their prominent sediment foliation and bubble trains. The foliation is nearly vertical near the centers of the wedges, and near horizontal at the margins of large wedges. Bubbles typically occupy up to 30 % of the ice and are spherical to tubular. The orientation of the bubbles was not related to foliation in many of the wedges examined. At the CAMP-1 section, bubble orientations measured throughout four exposed ice wedges indicate that bubbles were uniformly oriented, oblique to the valley side, rather than to ice-wedge foliation as one might expect if freezing were perpendicular to the axis of the wedge. Many of the centers of the ice wedges are eroded and contain both organic and mineral material of the same unit, and other types of ice, such as pool ice, which have been incorporated, likely due to erosion along the troughs of the wedges (Figure 5.8). The ice wedges extend well into underlying sediments, upturning beds at their margins. The sediments themselves, on inspection are ice-rich, and ice lenses, 10 to 20 cm long commonly occur parallel to bedding. In addition to ice wedges and pool ice, a continuous, 0.75 m bed of ice usually joins adjacent wedges. The ice composing this band has no foliation, and contains on average 20 % 1 to 3 mm spherical bubbles. Sediment is dispersed throughout the bed. This ice remains unidentifiable, and is perhaps analogous to that

Figure 5.8: CEMP-1 section showing erosion of anti-syngenetic ice wedge, and incorporation of sediments and pool ice. Scales are each 1.5 m.

described by French and Pollard (1986), but a repeated-segregated origin for this ice should not be discounted (Cheng, 1983). A thaw unconformity marks the top of this unit cryostratigraphically. It truncates wedge ice and pool ice. Above this, is a paleo-active layer, up to 1 m thick which contains abundant planar to wavy ice lenses, up to 40 cm long, which are interpreted as aggradational ice (Figures 5.7, 5.8).

Organic Unit

Ice wedges occur at various levels within this unit, and are narrow and up to 3 m in height. They have a very prominent sediment foliation. They are not as common as in the underlying unit, but where present, are found rejuvenating older wedges. Sediments adjacent to the margins are upturned. The upper truncation of these wedges occurs within the organic sediments, typically 1m below the surface. At several locations, the tops of the wedges have been deformed downslope, indicative of long-term slope movement, particularly creep (Figure 5.9). Pool ice and intrusive ice were seen locally within this unit.

General

In addition to ice found repeatedly within the stratigraphic units, two types of ice occur independently of the sediments. Injection ice was found near to the base of the deposits at several locations. This ice was identified based on its high transparency, lack of sediment, and bubble trains indicative of chilled margins (Figure 5.10). It was found most commonly penetrating upwards through sediments

Figure 5.9: Deformed ice syngenetic ice wedge within Organic unit at CAMP-1. Arrows demarcate growth steps. Scale bars mark horizontal and vertical orientation.

Figure 5.10: Intrusive ice from DG-1 showing lack of sediment and chilled margins.

from the underlying gravel. Fracture infill ice was observed at DG-1, near the top of the unit, indicative of brittle deformation at this location, as a result of downslope sediment movement (Figure 5.11a, b).

5.2.3 Ice Content

This section presents results of bulk and pore ice content determinations for the deposits. This information is combined to estimate the overall ice content of the deposits in the Klondike area.

Bulk Ice

Of the bulk ice types, visible (wedge, injection, pool, unidentified), ice-wedge ice was by far the most abundant. The area of an ice wedge (a), exposed in true width (w_{true}), is, approximately:

$$a = 0.5w_{\text{true}}h \quad [5.1]$$

where h is the height of the ice wedge. According to Mackay (1977), however, ice wedges are rarely viewed in true cross section. On average, an ice wedge makes an angle of 40° with the face, so that the true width is, on average, 0.64 times that of the exposed width. If the area of ice wedge exposure is to be calculated, then the equation becomes;

$$a = 0.32 w_{\text{exposed}}h \quad [5.2]$$

Using this for ice wedges only, the average visible areal ice content of all sections was determined to be 26 %. Some sections had areal ice contents of near 10 %, others

Figure 5.11: a) Fractures at DG-1 indicative of downslope movement, b) sub-horizontal banding observed within this fracture-infill ice.

were nearer to 35 %. Since the largest abundance of ice wedges is within the Dago Hill Fm, sections that were dominated by this unit had greater ice contents.

Pore Ice

Although over 100 samples were collected for ice-content determinations the results of many were deemed unusable. Bulk ice samples collected for sediment content were problematic because of the necessity to estimate the bubble content, and due to their rapid melting upon immersion into water. Other problems were encountered where the organic content of a sample was high. The water contained within the cellulose fibres of organic matter caused an overestimation of the ice content. An insufficient number of samples was collected from the Quartz Creek Fm to be statistically valid, but the average ice content for 10 samples was 65 % by volume. The results of ice content determinations for 43 samples collected from the Dago Hill Fm are presented in Figure 5.12. The average ice content for sediments of this unit, is 70 %, with a standard deviation of 15.3 %. Since frozen samples were collected to represent all types of sediments within this unit, the range of ice contents is quite high. As mentioned, samples with high organic content had ice contents of up to 95 %, and those from within siltier material were as low as 35 %.

The volumetric ice content of an entire deposit can be estimated if the approximate depth and lateral (i.e., up valley side) extent of sediments overlying creek gravel or bedrock is known. Bulk ice (mostly ice wedges) occupies, on average, 26 % of the area of exposed sections. The actual cross-sectional area of the

Figure 5.12: Frequency distribution of ice content determinations of Dago Hill Fm. Of 43 samples, the average volumetric intrasedimental ice content was 70 %.

ice wedges becomes $0.26(\cos 40^\circ)$ (Figure 5.13a) or 20 % of the area of the exposed section. Frozen sediments, therefore, occupy the other 80 % of the true cross-section (Figure 5.13a).

The volume of ice within a 1 m-thick deposit, of depth (d) and width (W) can be determined by adding the volume of ice wedges to the volume of pore ice. Since, as mentioned, ice wedges make an angle of 40° with the face, the 1 m-thick section will have an ice wedge which is $(\sin 40^\circ)^{-1}$, or 1.56 m long, within it (Figure 5.13b). The ice-wedge volume is then;

$$1.56 \text{ m}(0.20dW) = 0.31 \text{ dW m}^3 \quad [5.3]$$

The sediments occupy the remaining 80 % of the volume, and contain 70 % ice. The total volume of ice in the remaining sediments, is therefore;

$$1 \text{ m}(0.80dW) \times 0.7 = 0.56 \text{ dW m}^3 \quad [5.4]$$

The total volume of ice within a 1 m-thick deposit is the sum of the two (Figure 5.13c);

$$0.31 \text{ dW} + 0.56 \text{ dW} = 0.87 \text{ dW m}^3 \quad [5.5]$$

The total volume of ice within a deposit of thickness (t) and depth f(t) is then:

$$V_{\text{ice}} = 0.87 W \int_0^t f(t) dt \quad [5.6]$$

or, ice occupies 87 % of the volume of a muck deposit composed mostly of the Dago Hill Formation. This is illustrated in Figure 5.13d. This is a first approximation, since deposits can differ significantly.

Figure 5.13: Volumetric ice content of a deposit. a) True cross-sectional area occupied by ice wedges is 20 %; b) a 1 m-thick section will have a 1.56 m-long ice wedge within it; c) total volume of ice in a 1 m-thick section is 0.87 dW; d) the volumetric ice content of a deposit can be approximated if the thickness and depth up valley are known.

5.3.4 Summary of Cryostratigraphy

The generalized cryostratigraphy of muck deposits is shown in Figure 5.14. The lowermost unit, the Last Chance Creek Fm, has Pleistocene basal dates > 40 000 years BP. It contains epi- or syngenetic ice wedges and abundant ice lenses. A cryoturbated bed, also of age > 40 ka overlies this, and whose base truncates the ice wedges, marking a thaw unconformity. The overlying Quartz Creek Fm is a bedded silt, containing mammal bones and rhizomes. There are no ice wedges, but ice lenses are common. A *circa* 23 000 BP tephra bed was seen nearer to the base of the unit. The Dago Hill Fm consists of reworked, interbedded silt of the unit below and organic material. Large ice wedges dominate this unit, and penetrate underlying sediments. They are truncated at their tops by a thaw unconformity, above which is an abundance of aggradational ice. Basal dates suggest a very Late Pleistocene age for the start of accumulation of this unit. The uppermost Organic unit consists of peat and organic macrofossils. Small ice wedges penetrate the underlying unit.

The volumetric intrasedimental ice content of the Dago Hill Fm, is 70 %, and bulk ice occupies, on average, 26 % of exposed faces.

5.3 ISOTOPE GEOCHEMISTRY

The results of the $\delta^{18}\text{O}$ and $\delta^2\text{H}$ determinations for all ice samples are presented here. The section is divided into two subsections: one presenting the $\delta^{18}\text{O}$ profile, the other presenting the $\delta^2\text{H}$ - $\delta^{18}\text{O}$ co-isotope plot. Data for all samples can be found in Appendix II.

Figure 5.14: Generalized cryostratigraphy of frozen, unconsolidated materials found in the Klondike area.

5.3.1 $\delta^{18}\text{O}$ Profile

Thirteen ice samples were collected at approximately 50 cm intervals, across the contact between the Quartz Creek Fm and the Dago Hill Fm, at QC-1. Analysis of the $\delta^{18}\text{O}$ of these samples, revealed up to 5 ‰ enrichment upwards across this contact. Ratios 2 m below the contact are < -31 ‰, and above the contact, are > -27 ‰. There is a 2 m zone of intermediate values below the contact, which is interpreted as a mixing zone, and might be indicative of the paleo active layer during the initial stages of deposition of the Dago Hill Fm (Figure 5.15).

5.3.2 Co-isotope Relations

Fifty-two ice samples lie on a straight line on a co-isotope plot (Figure 5.16). The slope of this line is 7.84 ± 0.31 with $r^2 = 0.95$. A least-squares linear regression was used, rather than a least-squares cubic method or functional analysis, since the difference in the slopes obtained would likely not be significant (Burn and Maxwell 1993). The samples have been divided into pore ice, ice wedges and massive ice, which was not identified, but most resembled ice-wedge ice. Three distinct groupings can be seen on this plot. The most depleted samples are those of pore and wedge ice collected from the Quartz Creek Fm ($\delta^{18}\text{O}$: -31.9 to -29.3 ‰). The central grouping comprises ice collected from the Last Chance Creek and Dago Hill Fms ($\delta^{18}\text{O}$: -28.3 to -24.9 ‰). The most enriched samples come from the Organic unit ($\delta^{18}\text{O}$: -24.6 to -20.3 ‰). The significance of these groupings is discussed in the following chapter.

Figure 5.15: Isotope profile across contact between Quartz Creek Fm and Dago Hill Formation. $\delta^{18}\text{O}$ is up to 5‰ depleted below the unconformity.

Figure 5.16: Co-isotope plot of all ice samples collected in Klondike area. The least-squares linear regression has a slope of 7.84 ± 0.31 .

5.4 CHAPTER SUMMARY

Examination of the cryostratigraphy of the unconsolidated overburden in the Klondike area, has revealed four units within these deposits. The previously unidentified Last Chance Creek Fm started accumulating by at least 40 ka BP. The Silt unit of Fraser and Burn (1997) has been subdivided into two units, the Quartz Creek Fm and the Dago Hill Fm. The contact between the two is marked by an abundance of organic material, bulk ice, a 5 ‰ enrichment of $\delta^{18}\text{O}$, and a basal date of 11 620 BP. The Organic unit of Fraser and Burn (1997) unconformably overlies the Dago Hill Fm. The Last Chance Creek Fm and Dago Hill Fm contain ice of similar oxygen-isotopic composition. Ice from the Quartz Creek Fm is isotopically depleted, and from the Organic unit is isotopically enriched, with respect to this group.

CHAPTER 6: DISCUSSION OF RESULTS

This chapter presents a discussion of field and laboratory data. The first section examines the distribution of deposits in the Klondike area; the second and third discuss cryostratigraphy and isotope geochemistry; the Klondike deposits are compared to those of Alaska in the final section.

6.1 DISTRIBUTION OF DEPOSITS

The restriction of muck to north-facing slopes in larger valleys may be due to several factors: first, north-facing slopes receive less incident solar radiation than south-facing slopes, thus sustaining permafrost to a greater degree. When mining exposes material on south-facing slopes, it melts back more quickly since it is exposed directly to the sun. North-facing exposures on the other hand, are shaded from the sun, and although melting does occur, permafrost is re-established after the slope has stabilized, preserving any remaining muck. This is likely why west and south-facing valley walls expose stoney colluvium or bedrock. Second, larger valleys, especially those closer to Dawson, were the first to be exploited for their gold by early miners, because of the ease of accessibility. This may be the reason for lesser amounts of muck at these locations.

6.2 CRYOSTRATIGRAPHY

The first subsection of this section discusses the depositional and freezing history of the deposits based on the observed sediment and ice characteristics. The

second subsection is devoted to the ice content determinations of the deposits, and the problems encountered.

6.2.1 Depositional and Freezing History

The age of basal gravels exposed at several sections is unknown. These gravels are gold-bearing, and in most cases, it is these gravels that are being exploited by placer miners. Since Klondike Gravels do not contain significant quantities of gold, and White Channel Gravels are found on high-level terraces, exposed creek bottom gravels are most likely derived from reworked White Channel Gravel and eroded underlying bedrock (Naldrett 1981). The age of these creek-bottom gravels is not known, except that they must be older than 40 000 years BP (BGS-2015), the oldest age of material obtained above gravel in this study. This age is similar to the age obtained for creek and alluvial gravels in the Sixtymile area by Hughes et al. (1987).

Last Chance Creek Formation

The Late Pleistocene basal dates obtained in this study correspond with the end of the Reid-McConnell interglacial period in central Yukon, and the Boutellier Interval in unglaciated Beringia (Hopkins 1972). The presence of ice wedges suggests that permafrost existed in the Klondike at the time, and that there was sufficient snow cover to infill contraction cracks with meltwater. Figure 5.2 shows that the ice wedges have experienced two periods of growth. The first stage of

growth, represented by the largest portion of the ice wedges, is truncated just above the gravel. This indicates that the wedge must post-date the deposition of the gravel, since there must have been at least some material present above the gravel at the time of ice-wedge growth to allow for an active layer, likely the silt which makes up the lower facies of this unit. The second, smaller wedge which rejuvenated the lower wedge of this unit, is truncated by the cryoturbated facies. If this cryoturbated bed represents the thickness of the paleoactive layer, then the silt facies may have comprised the "relict" paleoactive layer (Harris et al. 1988). Ice lenses within this facies may be interpreted as aggradational ice, which formed as the active-layer thickness decreased with lowering of ground surface temperatures at the end of the Reid-McConnell interglacial and onset of McConnell glaciation in central Yukon.

Quartz Creek Formation

This unit is distinguishable from the underlying Last Chance Creek Formation, due to the relative absence of ice and tree material. Radiocarbon dates from grassbeds obtained at the base of this unit suggest accumulation began by 27 000 years BP, at the onset of McConnell glaciation (Fraser and Burn 1997). Permafrost must have developed syngenetically and persisted since deposition, as mammal carcasses have been preserved in these deposits (Harrington et al. 1994; Fraser and Burn 1997). No ice wedges were reported within the bottom portion of the Silt unit by Fraser and Burn (1997). The only ice wedges reported in this

study, are those which penetrate the upper portions of this unit from the overlying unit, or those which occur where the overlying unit is absent. There is no evidence, in the form of multiple truncations, or 'steps' that any of these ice wedges grew syngenetically during deposition of the Quartz Creek Fm (see Figure 3.1). Although sufficient water must have been available episodically for the sheetwash processes responsible for the development of bedding within these deposits (Fraser and Burn 1997), it is possible that snow melt was insufficient for the development of ice wedges. Interestingly, no other evidence of thermal contraction has been observed, either in the form of ice wedge casts, or sand wedges. Since Mackay (1974) has shown that cracking frequency of ice wedges increases with less snow cover, the absence of ice wedges within this unit is difficult to explain, unless a very thick snowcover existed, inhibiting ground cracking. The strong winds responsible for loess distribution in the area might have caused excessive snow to accumulate in valleys. There are several reasons why this hypothesis is not supported here. First, ice wedges are absent from valleys up to 300 m wide. Indeed an exceptional amount of snow must have accumulated to fill a valley this wide. Second, previous studies have shown convincingly that a dry, grassland environment existed at the time (e.g. Hopkins 1982), and third, it would be expected that if valley-bottoms had a thick snowcover, and the tops of hills were essentially bare, there would exist some locations (hillsides, for example), where insufficient snowcover *would* allow ice wedges to form.

A more reasonable explanation for the lack of ice wedges exists, if one considers that the thickness of these sediments was at one time much greater, that ice wedges *did* form, and what is currently exposed at all locations, is only the lower portion of ice wedges and material which was originally deposited. If ice wedges *did* form during the McConnell glacial, then they must have been epigenetic, forming after deposition had ceased. To summarize, either sediment thickness now represents the original thickness of material, and ice wedges did not develop, either due to deep snowcover, or insufficient moisture, or, the thickness of material was at one time greater, and only the bases of epigenetic ice wedges have now been preserved. The absence of this unit from broad valleys supports the erosion hypothesis, but, since a squirrel nest collected at DG-1 has been dated to $13\,910 \pm 70$ (Beta-111606; W.P. Lebarge, pers. comm.), close to the end of the McConnell glaciation, the hypothesis that most of the original material is present, and that no ice wedges formed in this period, must be considered in some locations.

Dago Hill Formation

The oldest date obtained from the base of this unit, suggests accumulation beginning by $11\,620 \pm 90$ years BP (TO-6869), towards the end of the Late Wisconsinan McConnell glaciation in central Yukon. The presence of larger organic material within this unit, although not *in situ*, suggests that post-glacial warming began by this time, three or four thousand years before the Early

Holocene warm interval recorded in other parts of the Canadian arctic (e.g., Burn et al. 1986; Mackay 1995; Burn 1997). The planar and convoluted nature of the beds of this unit, together with sandy lenses, suggests that intense fluvially-controlled colluviation occurred at the time, taking silt derived from the Quartz Creek Fm, together with organic material, and redepositing it downslope. The deposition of this unit may have been preceded or accompanied by erosion, since the Quartz Creek Fm is absent at several locations. If, as discussed, snowcover was insufficient during the McConnell glacial period for ice wedge formation, then in addition to warming, the start of deposition of this unit likely represents a change to wetter climate, allowing the formation of ice wedges. The age of the base of this unit corresponds closely with warming associated with the Late Wisconsinan Bølling-Allerød Interstadial recorded in Greenland ice cores which began 14.7 ka calendar years ago, approximately 12 ka radiocarbon years ago (e.g., Johnson et al. 1992; Taylor et al. 1993).

The syngenetic ice wedges observed on Dominion Creek had begun growing by $10\,125 \pm 380$ (BGS-2016; Figure 5.6). The syngenetic origin of these ice wedges suggests that there was continuous accumulation of material at this and other locations within broad creek valley-bottoms. On the valley sides of larger creeks, the ice wedges observed had a much broader form, with upturned sediments at their margins, suggesting either an epigenetic or anti-syngenetic origin (e.g., Mackay 1990). The ice wedges described by Fraser and Burn (1997), were interpreted as syngenetic or epigenetic in origin. It is proposed here, that at least

some of the very wide ice wedges with joined tops, found on valley sides, are anti-syngenetic in origin. The unit is consistently much thicker in valley bottoms than it is on valley sides, so if ice wedge growth had begun on valley sides at the same time as in valley bottoms, then early during the deposition of this unit, ice wedges may have formed syngenetically, but turned anti-syngenetic as some of the sediment was transported into valley-bottoms, causing a lowering of the ground surface. Furthermore, if epigenetic ice wedges existed within the Quartz Creek Fm, then denudation of this material would have caused anti-syngenetic ice wedges to develop, as in areas where the Dago Hill Fm is absent, e.g., GB-2 and RB-1.

The presence of large pockets of sediments, organic material and pool ice at the tops of the wedges, suggests that erosion of the ice wedges occurred close to the end of the depositional period (e.g., Mackay 1989). In places, the centres are completely eroded and filled with sediments and pool ice, so that only the margins of the ice wedges remain.

The continuous band of ice present near the base of this unit, where it is in contact with the Quartz Creek Fm is upturned at the margins of the ice wedges. In places it is seen above and below the Quartz Creek Fm/Dago Hill Fm unconformity. The ice might be of segregated or repeated-segregation origin, developing at the base of an active layer (e.g., Cheng 1983; Mackay 1989), or may be buried snowbank ice, similar to that observed on Hunker Creek by French and Pollard (1986). The ice in question was never accessible, and therefore detailed observations were not made as to the nature of the upper and lower contacts,

sediment inclusions or bubble content, all of which would help in identifying this ice.

Organic Unit

Organic material of this unit began to accumulate 9945 ± 150 years BP, at the start of the Holocene (Fraser and Burn 1997). The base of aggradational ice at the top of the Dago Hill Fm marks a thaw unconformity, which represents the maximum depth of the active layer at the start of the Early Holocene warm interval, dated to 8870 ± 200 in central Yukon (Burn et al. 1986) and from as early as 9480 ± 100 BP (Murton and French 1993) to 8765 ± 230 BP (Mackay 1983a) in the Mackenzie Delta area. Subsequent cooling of climate has caused a decrease in active-layer thickness, preserving the aggradational ice. Small, epigenetic or syngenetic ice wedges have developed within the organic material, and are truncated at the base of the current active layer. These ice wedges penetrate the older ice wedges truncated by the thaw unconformity. This indicates that they did not begin growing before the start of the Holocene. Deformation of these wedges, where present on valley sides, indicates that long-term slope movement has been taking place since the start of the Holocene (Figure 5.9).

6.2.2 Ice Content Determinations

Several problems were encountered when trying to determine the ice contents of sediments. The determination of sample volume was problematic for

two reasons: first, excess water or air was likely entrapped in the plastic wrap in which a sample was placed before immersion, and second, water would have been lost due to melting upon immersion. This would cause an over- estimation, or under-estimation of the volume of the sample. Samples with excess ice, or those containing mostly ice, were problematic as well, due to the errors associated with estimating the bubble content. The determination of ice content of the deposits was intended to be approximate. If further detailed investigations are to be made, then samples should be kept frozen until they can be transported to a location where more advanced laboratory techniques can be used.

6.3 ISOTOPE GEOCHEMISTRY

The results of isotopic analyses lie straight on a co-isotope plot, with a slope of 7.84 (Figure 5.16). The slope suggests that the bulk and intrasedimental ice sampled may originate directly from surface waters which has been subjected to minimal melting-freezing events (Craig 1961), rather than of segregated origin, in which case the slope might be closer to six (Jouzel and Souchez 1982) or ice which was derived from freezing in a closed system where the slope would be closer to 5 (Souchez and Jouzel 1984). Pingo ice has demonstrated slopes of 7.2, and although ice is derived from the freezing of groundwater in a closed system, the co-isotope slope is closer to that of precipitation, demonstrating that the slope is not diagnostic of water origin (Mackay 1990*b*). Although this may be the case, Figure 5.16 shows that ice-wedge ice and pore ice lie along the same line, at least

indicating similar origin. Ice-wedge ice is known to be infilled with meteoric waters (e.g., Dostovalov and Popov 1963; Romanovskij 1973; Mackay 1972), so, assuming that only ice-wedge ice is being sampled, pore ice may be considered to originate directly from surface waters as well.

The three distinct groupings of samples seen in Figure 5.16 are interpreted as being a direct result of temperature changes associated with glacial/non-glacial intervals during deposition of the sediments. Pore ice samples from the Quartz Creek Formation were the most depleted in ^{18}O and ^2H , one sample from the Organic unit was the most enriched, and Dago Hill and Last Chance Creek Fm samples had intermediate ratios. If the $\delta^{18}\text{O}$ - temperature relation presented by Dansgaard (1964) is valid in this context (i.e., equilibrium fractionation), then sediments of the Quartz Creek Fm were deposited when surface temperatures were colder than at any other time represented in the stratigraphy of the Klondike muck deposits. Deposition of these sediments occurred during the McConnell glacial (e.g., Fraser and Burn 1997). Samples taken from ice wedges which penetrate this unit are, on average, 3 ‰ enriched with respect to the sediments. Mackay (1983a) notes that ice wedges may be up to 5 ‰ *depleted* in $\delta^{18}\text{O}$ compared to sediments, due to infilling by the first meltwater of the snow, whereas Lauriol et al. (1995), have noted that ice wedges may be up to 6 ‰ *enriched* with respect to precipitation, suggesting that infill is from water released during later stages of snowmelt. Michel (1990) has shown that although some mixing of ice-wedge and sediment water occurs near the margins of ice wedges, the $\delta^{18}\text{O}$ of ice wedges closely

resembles that of sediments of the same age. If this is valid, then the more enriched ice wedges likely indicate that they are from a different period.

All ice taken from the Dago Hill Fm was enriched in heavy isotopes with respect to the Quartz Creek Fm. Basal dates for this unit indicate deposition near the end of the McConnell glacial, at the start of the Bølling-Allerød. This enrichment of isotopes further supports climate warming as indicated by the cryostratigraphy. Pore and wedge ice from the Last Chance Creek Fm had similar isotopic ratios to this unit. If the isotopic ratios reflect temperatures near the time of deposition of the sediments, then temperatures during deposition of the Last Chance Creek Fm were similar to those at the end of the McConnell glacial. Syngenetic wedges from the Dago Hill Formation were enriched compared to epi and anti-syngenetic ice wedges of the same unit.

The most enriched samples came from ice of the Organic unit. This unit is known to have been deposited at the start of the hypsithermal, when climate became abruptly warmer (e.g., Fraser and Burn 1997). Isotopic ratios obtained reflect this change to warmer conditions, and are similar to those obtained in other parts of the Canadian arctic (e.g., Michel and Fritz 1982; Mackay 1983a).

The $\delta^{18}\text{O}$ profile obtained across the contact between the Quartz Creek Fm (below), and the Dago Hill Fm, provides further evidence for warming at the end of the Late Wisconsinan McConnell glacial period (Figure 5.15). Values 2 m below the unconformity range between -30.9 ‰ and -31.9 ‰ and those above the unconformity are -26.8 ‰ to -25.3 ‰. There is a 1 m zone below the unconformity

where $\delta^{18}\text{O}$ becomes are similar to those above the unconformity, and 1 sample 1.5 m below the unconformity which shows an intermediate value. This is interpreted as a zone of mixing, where surface waters of late Late Wisconsinan age infiltrated sediments of Late Wisconsinan age, causing intrasedimental water to have intermediate isotopic ratios. The exact depth of this paleoactive layer is not known, since only one sample of intermediate value exists, and since surface waters have also been demonstrated to infiltrate into the top of permafrost from the active layer (Mackay 1983*b*). The presence of this mixing zone provides evidence of a thaw unconformity, even though one is not recognizable by examining the cryostratigraphy alone. The more negative ratios correspond with those of ice collected from correlative sediments, obtained throughout the Canadian arctic (e.g., Michel and Fritz 1982; Mackay 1983*a*). The less enriched ratios found in the profile, on the other hand, are not widely correlatable, but -26 ‰ to -27 ‰ ratios have been reported from the Eagle Plain area, below the modern sediments (Michel 1983), and below the Early Holocene thaw unconformity near Mayo (Burn et al. 1986).

The analysis of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ of intrasedimental and ice-wedge ice proved valuable, as the ratios support the long-term temperature shifts associated with glacial/non-glacial periods, represented in the cryostratigraphy.

6.4 COMPARISON OF KLONDIKE AND ALASKAN DEPOSITS

This section briefly presents a comparison of the Klondike deposits to those

of adjacent Alaska, based on what is described in the literature. An attempt is made to correlate deposits of the two areas.

The stratigraphy of Alaskan deposits spans a much larger time period than that of the Klondike. The oldest unit, the Gold Hill Loess, represents loess deposition and erosion between 3 Ma and 130 ka (e.g., Péwé 1975; Péwé et al. 1997). Unconformably overlying this, is the Eva Forest bed, a 125 ka old unit consisting dominantly of spruce wood, interpreted as an interglacial deposit (Péwé et al. 1997). Radiocarbon ages of wood collected from this unit vary from >29 000 (Péwé et al. 1997) to 56 900 (Matthews 1970) years BP, demonstrating that wood of this unit is beyond the limits of radiocarbon dating. The age of this forest bed has been determined by thermoluminescence dating of tephra and loess from above and below the unconformity. These dates suggest that the interglacial might have begun as early as 134 ka and ended as late as 90 ka (Péwé et al. 1997). Overlying the Eva Forest Bed in valley-bottoms, is the Goldstream Fm, a Late Wisconsinan dark, ice-wedge-rich retransported loess, containing mammal remains (Péwé 1975; Guthrie 1990). The Ready Bullion Formation unconformably overlies the Goldstream Formation, and consists of Holocene (10 ka) organic-rich loess, and the Giddings Forest Bed (Péwé 1995).

In the Klondike deposits, the >40 ka ages obtained for wood overlying gravel at DC-1, and from within the cryoturbated bed of the Last Chance Creek Formation may be correlatable to the Eva Forest Bed in Alaska, since both were near or beyond the limits of radiocarbon dating. The sediments of the Last Chance

Creek Formation, however, are not correlatable, since there is no record of deposition during the interglacial period in Alaska, only the development of a forest. There is also no evidence of permafrost existing during this period within the Alaskan deposits, whereas the Klondike deposits clearly show the development of ice wedges.

The Quartz Creek Formation in the Klondike area, is correlatable with the Goldstream Formation in Alaska, with respect to timing of deposition (Late Wisconsinan), but the ice and sediment characteristics of the two differ in several respects. The Alaskan deposits are grey to black and contain abundant ice wedges (Péwé et al. 1997). The Klondike deposits are light brown to grey-brown and contain no ice wedges. Both consist of retransported loess and have mammal remains. The 23 ka tephra bed identified in the Klondike deposits is not reported in Alaska.

There is no Alaskan unit correlatable to the Dago Hill Formation. The thaw unconformity at 11 620 years BP, and the intensely colluviated sediments do not appear in the Alaskan sedimentary record. The Organic unit can be correlated with the Ready Bullion Formation, the base of which marks the start of the 10 ka warming interval (Péwé 1975).

Further research would be appropriate to investigate the differences and similarities of deposits of the two areas. In particular, the age of the Last Chance Creek Formation, and the distribution of the Dago Hill Formation should be determined. The latter, if not detectable sedimentologically, could be identified

through the use of $\delta^{18}\text{O}$ ratios of the intrasedimental ice.

6.5 CHAPTER SUMMARY

In the Klondike area, muck is found primarily on north to northeast-facing slopes where it is likely that less incident solar radiation favours the preservation and re-establishment of permafrost after disturbance.

Overlying creek gravel, the Last Chance Creek Formation is interpreted as a Reid-McConnell age late interglacial deposit, where permafrost existed and sufficient moisture was available for the development of ice wedges. Climate was warmer than during the ensuing glacial period. The Quartz Creek Formation was deposited during cold, full-glacial conditions (McConnell-age), when permafrost existed, but most probably, insufficient snowcover was present to allow the formation of ice wedges. The end of the McConnell glacial period is marked by a shift to warmer and wetter climatic conditions, allowing intense colluviation, the growth of trees and the development of ice wedges to take place, over a period of about two thousand years, synchronous with the pre-hypsithermal Bølling-Allerød interval recorded in Greenland ice cores.

The Quartz Creek Formation and Organic unit are likely correlatable to Alaskan units, but the correlatability of the Last Chance Creek and Dago Hill Formations remains questionable.

CHAPTER 7: CONCLUSIONS

Several conclusions can be drawn from the results of cryostratigraphic and isotopic investigations of muck deposits in the Klondike area, and names are proposed for three of the four units recognized within these deposits. These units represent the latest Cenozoic sediment accumulation, being Wisconsinan loess deposits, overlain by Holocene organic material.

The lowermost unit, the Last Chance Creek Formation, unconformably overlies gravel. It is ice-rich silt, containing cryoturbated organic material and syngenetic or epigenetic ice wedges. $\delta^{18}\text{O}$ ratios between -29 and -26 ‰, and organic material of age >40 ka suggest that this unit was deposited during, or in later stages of the Reid-McConnell interglacial period, when conditions were wet enough to allow the development of ice wedges.

The Quartz Creek Formation is found in narrow valleys, unconformably overlying either the Last Chance Creek Formation or gravels. $\delta^{18}\text{O}$ ratios between -32 and -29 ‰ and the absence of ice wedges, suggest that climatic conditions were cold and dry at the time of deposition, during the Late Wisconsinan McConnell glacial period. A 23 ka tephra bed occurs near the base of this unit.

The Dago Hill Formation unconformably overlies the Quartz Creek Formation in narrow valleys, and overlies gravel in broad valleys. It consists of colluviated silt and organic material, particularly twigs and branches. Ice wedges dominate this unit, and are syngenetic in origin in valley-bottoms where there has been an accumulation of material, and anti-syngenetic on valley sides where

material has been removed. Ice wedges are truncated at their tops by a thaw unconformity, above which is aggradational ice, suggesting a decrease in active-layer depth, or accumulation of material since ice-wedge development. $\delta^{18}\text{O}$ ratios between -28 and -25 ‰ and the abundance of ice within this unit suggest a change to warmer and wetter conditions associated with the start of deposition of this unit, $11\,620 \pm 90$ years BP.

The Organic unit unconformably overlies the Dago Hill Formation. Syngenetic or epigenetic ice wedges from the Organic unit penetrate the underlying Dago Hill Formation. $\delta^{18}\text{O}$ ratios between -25 and -20 ‰ support previous work suggesting further warming of climate associated with the deposition of this unit at the start of the Holocene.

Sediments of the Dago Hill Formation have a volumetric ice content of 70%, and bulk ice occupies, on average, 26 % of exposed faces. Co-isotope relations indicate that most ice within the deposits was frozen directly from surface water.

Further research opportunities exist in obtaining AMS dates for organic material from the base of the Quartz Creek Formation, identifying organic material from the Last Chance Creek Formation, and identifying and more accurately dating the tephra bed. In a more regional study, these deposits should be correlated with those of Alaska. Following this study, a more extensive drilling program could be undertaken, and ground ice estimates refined.

REFERENCES

- Anderson, D.M. and Tice, A.R. 1972. Predicting unfrozen water contents in frozen soils from surface area measurements. Highway Research Record, **393**: 12-18.
- Bard, E., Hamelin, B., Fairbanks, R.G. and Zindler, A. 1990. Calibration of the ^{14}C timescale over the past 30,000 years using mass spectrometric U-Th ages from Barbados corals. *Nature*, **345**: 405-410.
- Bartlein, P.J., Edwards, M.E., Shafer, S.L. and Barker, E.D., Jr. 1995. Calibration of radiocarbon ages and the interpretation of paleoenvironmental records. *Quaternary Research*, **44**: 417-424.
- Berger, G.W. 1987. Thermoluminescence dating of Pleistocene Old Crow tephra and adjacent loess, near Fairbanks, Alaska. *Canadian Journal of Earth Sciences*, **24**: 1975-1984.
- Berger, G.W., Péwé, T.L., Westgate, J.A. and Preece, S.J. 1996. Age of Sheep Creek tephra (Pleistocene) in central Alaska from thermoluminescence dating of bracketing loess. *Quaternary Research*, **45**: 263-270.
- Black, R.F. 1983. Three superposed systems of ice wedges at McLeod Point, northern Alaska, may span most of the Wisconsinan stage and Holocene. *In* Proceedings of the Fourth International Conference on Permafrost, Alaska, National Academy Press, pp. 68-73.
- Bond, J.D. 1996. Quaternary history of McQuesten map area, central Yukon. *Yukon Quaternary Geology*, **1**: 27-46.
- Bostock, H.S. 1966. Notes on Glaciation in Central Yukon Territory. Geological Survey of Canada Paper 65-36.
- Bostock, H.S. 1970. Physiographic regions of Canada. Geological Survey of Canada, Map 1254A.
- Burn, C.R. 1997. Cryostratigraphy, paleogeography, and climate change during the early Holocene warm interval, western Arctic coast, Canada. *Canadian Journal of Earth Sciences*, **34**: 912-925.
- Burn, C.R. and Maxwell, M.G. 1993. Proper determination of the $\delta^{18}\text{O}$ - δD relationship for ice and water by least-squares cubic regression. *Canadian Journal of Earth Sciences*, **30**: 109-112.

- Burn, C.R., Michel, F.A. and Smith, M.W. 1986. Stratigraphic, isotopic and mineralogical evidence for an early Holocene thaw unconformity at Mayo, Yukon Territory. *Canadian Journal of Earth Sciences*, **23**: 794-803.
- Campbell, J.D. 1952. The paleobotany and stratigraphic sequence of the Pleistocene Klondike "muck" deposits. Ph.D. thesis, McGill University, Montréal.
- Cheng, G. 1983. The mechanism of repeated-segregation for the formation of thick layered ground ice. *Cold Regions Science and Technology*, **8**: 57-66.
- Collinson, J.D. and Thompson, D.B. 1982. *Sedimentary Structures*. Chapman and Hall, 207 pp.
- Craig, H. 1961. Isotopic variations in meteoric waters. *Science*, **133**: 1702-1703.
- Dansgaard, W. 1964. Stable isotopes in precipitation. *Tellus*, **16**: 435-468.
- Dawson, G.M. 1894. Notes on the occurrence of mammoth remains in the Yukon district of Canada and in Alaska. *Geological Society of London*, Feb. pp. 1-9.
- Demek, J. 1978. Periglacial Geomorphology. *In* *Geomorphology: present problems and future prospects*. Embleton, Brunsden and Jones (eds.), Oxford University Press, Oxford, pp. 139-155.
- Dostovalov, B.N. and Popov, A.I. 1963. Polygonal systems of ice-wedges and conditions of their development. *In* *Proceedings of the First International Conference on Permafrost*, Lafayette, Indiana. National Academy of Sciences-National Research Council, Washington, no. 1287, pp. 102-105.
- Duk-Rodkin, A. 1996. *Surficial Geology of Dawson Yukon Territory*. Geological Survey of Canada Open File 3288. 1:250 000 scale.
- Duk-Rodkin, A., Bond, J., Jackson, L.E., Jr., and Barendregt, R. 1995. Late Cenozoic record of glaciations in the northern Canadian Cordillera. *In* *Abstracts, 14th International Congress, International Union for Quaternary Research*, August 1995. Free University Press, Berlin, p. 70.
- Fraser, T. 1995. On the nature and origin of muck deposits, Klondike District, Yukon Territory. M.A. thesis, Carleton University, Ottawa, 203 pp.

- Fraser, T.A. and Burn, C.R. 1997. On the nature and origin of "muck" deposits in the Klondike area, Yukon Territory. *Canadian Journal of Earth Sciences*, **34**: 1333-1344.
- French, H.M. and Harry, D.G. 1990. Observations on buried glacier ice and massive segregated ice, western Arctic coast, Canada. *Permafrost and Periglacial Processes*, **1**: 31-43.
- French, H.M. and Pollard, W.H. 1986. Ground-ice investigations, Klondike District, Yukon Territory. *Canadian Journal of Earth Sciences*, **23**: 550-560.
- French, H.M., Harry, D.G. and Clark, M.J. 1982. Ground ice stratigraphy and late Quaternary events, south-west Banks Island, Canadian Arctic. *In* Proceedings of the Fourth Canadian Permafrost Conference, Calgary, National Research Council of Canada, Ottawa, pp. 81-
- Froese, D.G. 1997. Sedimentology and paleogeography of Klondike River valley terrace gravel, Yukon Territory. Unpublished M.Sc. thesis, University of Calgary.
- Froese, D.G. and Hein, F.J. 1996. Sedimentology of a high-level terrace placer gold deposit, Klondike valley, Yukon. *Yukon Quaternary Geology*, **1**: 13-26.
- Fujino, K., Horiguchi, K., Shinbori, M. and Kato, K. 1983. Analyses and characteristics of cores from the massive ice body in the Mackenzie delta, NWT, Canada. Proceedings, Fourth International Conference on Permafrost, Fairbanks, National Academy Press.
- Gleeson, C.F. 1970. Heavy mineral studies in the Klondike area, Yukon Territory. Geological Survey of Canada, Bulletin 173.
- Gonfiantini, R. 1981. The δ -notation and the mass-spectrometric measurement techniques. *In* Stable isotope hydrology: deuterium and oxygen-18 in the water cycle. Gat, J.R. and Gonfiantini (eds.). International Atomic Energy Agency, Technical Reports Series 210, p. 35-84.
- Green, L.H. 1972. Geology of Nash Creek, Larsen Creek and Dawson map areas. Geological Survey of Canada, Memoir 364.
- Guthrie, R.D. 1990. Frozen fauna of the Mammoth Steppe: The story of Blue Babe. Chicago, Illinois, University of Chicago Press, 323 pp.

- Hamilton, T.D. and Brigham-Grette, J. 1991. The last interglaciation in Alaska: stratigraphy and paleoecology of potential sites. *Quaternary International*, **10-12**: 49-71.
- Harington, C.R. and Clulow, F.V. 1973. Pleistocene mammals from Gold Run Creek, Yukon Territory. *Canadian Journal of Earth Sciences*, **10**: 697-744.
- Hopkins, D.M. 1982. Aspects of the paleogeography of Beringia during the Late Pleistocene. *In* Hopkins, D.M., Matthews, J.V., Jr., Schweger, C.E., Young, S.B. (eds.). *Paleoecology of Beringia*. Academic press, pp. 3-28.
- Hughes, O.L. 1972. Surficial geology of northern Yukon Territory and northwestern District of Mackenzie, Northwest Territories. Geological Survey of Canada, Paper 69-36, 11 pp.
- Hughes, O.L. 1987. Quaternary Geology. *In* Guidebook to Quaternary Research in Yukon, Morison, S.R. and Smith, C.A.S. (eds.) XII INQUA Congress Ottawa, Canada. National Research Council of Canada.
- Hughes, O.L. 1989. Quaternary Chronology, Yukon and Western District of Mackenzie. *In* Late Cenozoic History of the Interior Basins of Alaska and the Yukon, United States Geological Survey Circular 1026: 25-29.
- 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, Field Excursion Guidebook A11, 59 pp.
- Hughes, O.L., Morison, S.R. and Hein, F.J. 1987. Placer gravels of Miller Creek, Sixtymile area, 116B-116C. *Yukon Geology*, **1**: 50-55.
- Hughes, O.L., Rutter, N.W. and Clague, J.J. 1989. Yukon Territory (Quaternary stratigraphy and history, Cordilleran Ice Sheet). *In* Quaternary Geology of Canada and Greenland. R.J. Fulton (ed.). Geological Survey of Canada, Geology of Canada, No.1.
- Hyatt, J.A. 1990. Reconstruction of Holocene periglacial environments in the Pangnirtung area based on ice wedge characteristics. *In* Proceedings of the Fifth Canadian Permafrost Conference. Université Laval, Centre d'Études Nordiques, Québec and National Research Council Canada, Ottawa, Collection Nordicana, no. 54, pp. 17-
- Jackson, L.E., Jr. and Harington, C.R. 1991. Middle Wisconsinan mammals, stratigraphy and sedimentology at the Ketz River site, Yukon Territory. *Géographie physique et Quaternaire*, **45**: 69-77.

- Johnson, S.J., Clausen, H.B., Dansgaard, W., Fuhrer, K., Gundestrup, N., Hammer, C.U., Iversen, P., Jouzel, J., Stauffer, B. and Steffensen, J.P. 1992. Irregular glacial interstadials recorded in a new Greenland ice core. *Nature* **359**: 311-313.
- Jouzel, J. and Souchez, R.A. 1982. Melting-refreezing at the glacier sole and the isotopic composition of the ice. *Journal of Glaciology*, **28**: 35-42.
- Katasonov, E.M. 1969. Composition and cryogenic structure of permafrost. *In* Permafrost investigations in the field. National Research Council of Canada, Ottawa, Technical Translation 1358, pp. 25-36.
- Kirshenbaum, I. 1951. Physical Properties and Analysis of Heavy Water. McGraw-Hill, New York.
- Kudriavstev, V.A. 1978. (ed.) General Permafrost Science, 2nd edition. Moscow University, Moscow. (In Russian)
- Lauriol, B., Dushesne, C. and Clark, I.D. 1995. Systématique du remplissage en eau des fentes de gel: les résultats d'une étude oxygène-18 et deutérium. *Permafrost and Periglacial Processes*, **6**: 47-55.
- Laxton, N.F., Burn, C.R. and Smith, C.A.S. 1996. Productivity of loessal grasslands in the Kluane Lake region, Yukon Territory, and the Beringian 'production paradox'. *Arctic*, **49**: 129-140.
- Lorrain, R.D. and Demeur, P. 1985. Isotopic evidence for relic Pleistocene glacier ice, Victoria Island, Canadian Arctic Archipelago. *Arctic and Alpine Research* **17**: 89-98.
- Mackay, J.R. 1962. Pingos of the Pleistocene Mackenzie Delta area. *Geographical Bulletin*, **18**: 21-63.
- Mackay, J.R. 1965. Gas-domed mounds in permafrost, Kendall Island, N.W.T. *Geographical Bulletin*, **7**: 105-115.
- Mackay, J.R. 1971. The origin of massive icy beds in permafrost, western Arctic coast, Canada. *Canadian Journal of Earth Sciences*, **8**: 397-422.
- Mackay, J.R. 1972. The world of underground ice. *Annals of the Association of American Geographers*, **62**: 1-22.
- Mackay, J.R. 1974. Ice-wedge cracks, Garry Island, Northwest Territories. *Canadian Journal of Earth Science*, **11**: 1366-1383.

- Mackay, J.R. 1983*a*. Oxygen isotope variations in permafrost, Tuktoyaktuk peninsula area, Northwest Territories. *In* Current Research, Part B, Geological Survey of Canada, Paper 83-1B. pp. 67-74.
- Mackay, J.R. 1983*b*. Downward water movement into frozen ground, western arctic coast, Canada. *Canadian Journal of Earth Sciences*, **20**: 120-134.
- Mackay, J.R. 1989. Massive ice: some field criteria for the identification of ice types. Current Research, Part G, Geological Survey of Canada, Paper 89-1G, p. 5-11.
- Mackay, J.R. 1990*a*. Some observations on the growth and deformation of epigenetic, syngenetic and anti-syngenetic ice wedges. *Permafrost and Periglacial Processes*, **1**: 15-29.
- Mackay, J.R. 1990*b*. Seasonal growth bands in pingo ice. *Canadian Journal of Earth Sciences*, **27**: 1115-1125.
- Mackay, J.R. 1995. Ice wedges on hillslopes and landform evolution in the late Quaternary, western Arctic coast, Canada. *Canadian Journal of Earth Sciences*, **32**: 1093-1105.
- Mackay, J.R. and Dallimore, S.R. 1992. Massive ice of the Tuktoyaktuk area, western Arctic coast, Canada. *Canadian Journal of Earth Sciences*, **29**: 1235-1249.
- Matthews, J.V. 1970. Quaternary environmental history of interior Alaska; pollen samples from organic colluvium and peats. *Arctic and Alpine Research*, **2**: 241-251.
- McConnell, R.G. 1890. Glacial features of parts of the Yukon and Mackenzie Basins. *Geological Society of America Bulletin*, **1**: 540-544.
- McConnell, R.G. 1905. Report on the Klondike Goldfields, Canada. *Geological Survey Annual Report 19-B*, 1-17.
- McConnell, R.G. 1907. Report on gold values in the Klondike high level gravels. *Geological Survey of Canada Publication 979*, 34 pp.
- Michel, F.A. 1982. Isotope investigations of permafrost waters in northern Canada. Ph.D. thesis, University of Waterloo, Waterloo, 424 pp.

- Michel, F.A. 1983. Isotope variations in permafrost waters along the Dempster highway pipeline corridor. *In* Proceedings of the Fourth International Conference on Permafrost, Alaska, National Academy Press, pp. 843-847.
- Michel, F.A. 1990. Isotopic composition of ice-wedge ice in northwestern Canada. *In* Proceedings of the Fifth Canadian Permafrost Conference. Université de Laval, Centre d'études nordiques, Québec and National Research Council Canada, Ottawa, Collection Nordicana, no. 54, pp. 5-9.
- Michel, F.A. and Fritz, P. 1982. Significance of isotope variations in permafrost waters at Illisarvik, N.W.T. *In* Roger E. Brown Memorial Volume, Proceedings of the Fourth Canadian Permafrost Conference, Calgary, National Research Council, Canada, H.M. French (ed.), pp. 173-181.
- Milner, M.W. 1976. Geomorphology of the Klondike placer goldfields. Preliminary to Ph.D. thesis. Final Report, contract OSV 5-0047, Department of Indian Affairs and Northern Development, Canada. 157 pp.
- Morison, S.R. 1985. Sedimentology of White Channel placer deposits, Klondike area, west-central Yukon. Unpublished M.Sc. thesis, University of Alberta.
- Morison, S.R. 1989. Placer deposits in Canada. ch. 11, Quaternary Resources in Canada. *In* Quaternary Geology of Canada and Greenland, Fulton, R.J. (ed.), Geological Survey of Canada, Geology of Canada, no. 1, pp. 687-697.
- Morison, S.R. and Smith, C.A.S. (eds.) 1987. Guidebook to Quaternary Research in Yukon, 12th INQUA Congress, Ottawa, Canada. National Research Council of Canada, Ottawa. 110 pp.
- Murton, J.B. and French, H.M. 1993. Sand wedges and permafrost history, Crumbling Point, Pleistocene Mackenzie Delta, Canada. *In* Proceedings of the 6th International Conference on Permafrost, vol. 1, Beijing, China, University of Technology Press, p. 482-487.
- Murton, J.B. and French, H.M. 1994. Cryostructures in permafrost, Tuktoyaktuk coastlands, western arctic Canada. *Canadian Journal of Earth Sciences*, **31**: 737-747.
- Naldrett, D.L. 1981. Aspects of the surficial geology and permafrost conditions, Klondike Goldfields and Dawson City, Yukon Territory. M.Sc. thesis, University of Ottawa, Ottawa. 150 pp.

- Nickling, W.G. 1978. Eolian sediment transport during dust storms: Slims River valley, Yukon Territory. *Canadian Journal of Earth Sciences* **15**: 1069-1084.
- Nikolayev, V.I. and Mikhalev, D.V. 1995. An oxygen-isotope paleothermometer from ice in Siberian permafrost. *Quaternary Research*, **43**: 14-21.
- Permafrost Subcommittee, Associate Committee on Geotechnical Research. 1998. Glossary of permafrost and related ground ice terms, National Research Council of Canada.
- Péwé, T.L. 1952. Geomorphology of the Fairbanks area, Alaska. Ph.D. thesis, Stanford University, Stanford. 220 pp.
- Péwé, T.L. 1955. Origin of the upland silt near Fairbanks, Alaska. *Geological Survey of America Bulletin*, **66**: 699-724.
- Péwé, T.L. 1966. Ice wedges in Alaska - classification, distribution, and climatic significance. *In* Proceedings, International Permafrost Conference, Lafayette, 1963. National Academy of Sciences, National Research Council Publications, no. 1287, p. 76-81.
- Péwé, T.L. 1975. Quaternary stratigraphic nomenclature in central Alaska. United States Geological Survey Professional Paper 862, 32 pp.
- Péwé, T.L. 1989. Quaternary stratigraphy of the Fairbanks area, Alaska. *In* Late Cenozoic history of the interior basins of Alaska and the Yukon, Carter, L.D., Hamilton, T.D. and Galloway, J.B. (eds.), United States Geological Survey Circular 1026, pp. 72-77.
- Péwé, T.L., Berger, G.W. and Westgate, J. A. 1995. Past major global warming in 3 Myr loess record, central Alaska, USA, indicated by three buried taiga forest beds: 2 Myr, 125 ka, 10 ka. Berlin, Germany, International Union for Quaternary Research (Abstract Volume), p.217.
- Péwé, T.L., Berger, G.W., Westgate, J.A., Brown, P.M. and Leavitt, S.W. 1997. Eva Interglaciation forest bed, unglaciated east-central Alaska: global warming 125,000 years ago. *Geological Society of America Special Paper* 319, 54 pp.
- Placer Mining Section. 1996. Yukon Placer Mining Industry 1994-1995. Mineral Development Division, Yukon, Indian and Northern Affairs, Canada.

- Pollard, W.H. and French, H.M. 1984. The groundwater hydraulics of seasonal frost mounds, North Fork Pass, Yukon Territory. *Canadian Journal of Earth Sciences*, **21**: 1073-1081.
- Richardson, J. 1861. *The polar regions*. A.&C. Black, Edinburgh.
- Roddick, J.A. 1967. Tintina Trench. *Journal of Geology*, **84**: 23-33.
- Romanovskij, N.N. 1973. Regularities in formation of frost-fissures and development of frost-fissure polygons. *Biuletyn Peryglacjalny*, **23**: 237-
- Rutter, N.W., Foscolos, A.E. and Hughes, O.L. 1978. Climatic trends during the Quaternary in central Yukon based upon pedological and geomorphical evidence. *In* Quaternary Soils, Mahaney, W.C. (ed.), Geo Abstracts, Norwich, England. p. 309-359.
- Schweger, C.E. and Matthews, J.V., Jr. 1985. Early and Middle Wisconsinan environments of eastern Beringia: stratigraphic and paleoecological implications of the Old Crow tephra. *Géographie physique et Quaternaire*, **39**: 275-290.
- Shumskii, P.A. 1964. Ground (subsurface) ice. National Research Council of Canada, Ottawa, Technical Translation 1130.
- Souchez, R.A. and Jouzel, J. 1984. On the isotopic composition in δD and $\delta^{18}O$ of water and ice during freezing. *Journal of Glaciology*, **30**: 369-372.
- Tarnocai, C. 1989. Paleosols of northwestern Canada. *In* Late Cenozoic History of the Interior Basins of Alaska and the Yukon, United States Geological Survey Circular 1026: 39-44.
- Tarnocai, C., Smith, C.A.S. and Hughes, O.L. 1985. Soil development on Quaternary deposits of various ages in central Yukon Territory. Current Research Part A. Geological Survey of Canada Paper 85-1A: 229-238.
- Taylor, K.C., Lamorey, G.W., Doyle, G.A., Alley, R.B., Grootes, P.M., Mayewski, P.A., White, J.W.C. and Barlow, L.K. 1993. The 'flickering switch' of late Pleistocene climate change. *Nature*, **361**: 432-436.
- Tempelman-Kluit, D.J. 1972. Evidence for timing and magnitude of movement along the Tintina Trench, (abs.). Geological Association of Canada, Cordilleran Section Annual Meeting.

- Tempelman-Kluit, D.J. 1980. Evolution of physiography and drainage in Southern Yukon. *Canadian Journal of Earth Sciences*, **17**: 1189-1203.
- Tuck, R. 1940. Origin of the muck-silt deposits at Fairbanks, Alaska. *Geological Society of America Bulletin*, **51**: 1295-1310.
- Tyrell, J.B. 1917. Frozen muck in the Klondike District. *Transcripts of the Royal Society of Canada*, **4**: 39-46.
- Van Everdingen, R.O. 1978. Frost mounds at Bear Rock, near Fort Norman, Northwest Territories, 1975-1976. *Canadian Journal of Earth Sciences*, **15**: 263-276.
- Vasil'chuk, Y.K. and Vasil'chuk, A.C. 1995. Ice wedge formation in northern Asia during the Holocene. *Permafrost and Periglacial Processes*, **6**: 273-279.
- Williams, P.J. 1967. Properties and behaviour of freezing soils. Norwegian Geotechnical Institute, Paper 72, 120 pp.

Appendix I: Section illustrations
(all scales are in metres above creek gravel)

LEGEND TO ACCOMPANY ILLUSTRATIONS

HOLOCENE		Organic unit: 1 - 4 m, colluvial and in place; dark brown to black.
	Colluvium: pebble to cobble diamicton; yellow to brown
LATE PLEISTOCENE		Erosional
		Dago Hill Fm:
	Silt, light brown
	Thaw
	Colluviated silt + organics, 3 to 12 m thick; abundant grass and twigs, light to dark brown.
	Erosional
	Tephra: 0 - 20 cm thick, continuous to discontinuous, grey to cream.
	Quartz Creek Fm: grassy silt, 0 - 10 m; beige-brown with rhizomes and grass beds.
	Erosional
		Last Chance Creek Fm:
TERTIARY TO QUATERNARY		Cryoturbated bed: up to 75 cm; contains tree material, >40 ka.
	Thaw
	Silt, 1m; light beige, ice lens-rich
	Erosional
	Creek gravel: gold-bearing pebble cobble gravel; grey to yellow.
	Graphitic schist
	Ice: wedges and lenses
	Mammal bones
	Grass beds and rhizomes
	Tree twigs, branches

Appendix II: Complete results of isotope geochemistry

	delta 18-O (+/- 0.2) permille SMOW	delta 2-H (+/- 2) permille SMOW
DC2-IW2-01	-21.2	-164
IW1-3-01	-22.2	-176
DC2-IW-02	-22.8	-182
IW2-C-01	-23.2	-179
DC2-IW2-03	-23.3	-184
IW1-C-01	-23.4	-187
DC2-IW2-02	-23.6	-187
IW1-2-01	-23.8	-186
DC2-IW-03	-24.1	-192
IW1-1-01	-24.6	-189
QC-IW-03	-26.9	-214
LCC-PW-2	-27.5	-225
QC-IW-02	-27.5	-216
FV-IW-2	-27.7	-219
QC-IW-04	-27.7	-220
QC-IW-05	-27.8	-222
FV-IW-1	-28.1	-220
LCC-PW-1	-28.3	-220
QC-IW2-01	-30.7	-273
L1-01	-20.3	-164
QC-720	-25.3	-206
QC-770	-25.7	-210
QC-450	-26.0	-208
LCC-02	-26.3	-210
LCC-01	-26.5	-204
QC-490	-26.8	-201
QC-620	-26.8	-202
LCC-03	-27.2	-209
LCC-04	-27.5	-216
LCC-05	-27.7	-214
LCC-07	-28.2	-216
LCC-06	-28.3	-214
QC-400	-29.3	-234
QC-L-01	-29.8	-239
QC-250	-30.9	-248
QC-0	-31.2	-256
QC-50	-31.2	-256
QC-150	-31.7	-254
QC-300	-31.9	-248
QC-200	-31.9	-255
QC-100	-31.9	-257
MI1-01	-22.3	-178
MI1-02	-22.8	-181
MI1-10	-24.9	-203
DC1-GI-01	-25.7	-206
MI1-05	-25.7	-211
MI1-09	-25.8	-213
MI1-03	-26.6	-214
MI1-08	-26.6	-218
MI1-04	-26.7	-214
MI1-06	-27.0	-217
MI1-07	-27.5	-225
IW1-1-02	-23.6	n/a

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