A late glacial and holocene chronology of the Castner Glacier, Delta River Valley, Alaska

Michael W. Howley

University of New Hampshire, Durham

Follow this and additional works at: https://scholars.unh.edu/thesis

Recommended Citation

https://scholars.unh.edu/thesis/421

This Thesis is brought to you for free and open access by the Student Scholarship at University of New Hampshire Scholars' Repository. It has been accepted for inclusion in Master's Theses and Capstones by an authorized administrator of University of New Hampshire Scholars' Repository. For more information, please contact nicole.hentz@unh.edu.
A LATE GLACIAL AND HOLOCENE CHRONOLOGY OF THE CASTNER
GLACIER, DELTA RIVER VALLEY, ALASKA

BY

MICHAEL W. HOWLEY
B.S., University of New Hampshire, 2006

THESIS

Submitted to the University of New Hampshire
in Partial Fulfillment of
the Requirements for the Degree of

Master of Science

in

Earth Sciences: Geology

December, 2008
INFORMATION TO USERS

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 bleed-through, substandard margins, and improper alignment can adversely affect reproduction.

In the unlikely event that the author did not send 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.

UMI Microform 1463225
Copyright 2009 by ProQuest LLC.
All rights reserved. This microform edition is protected against unauthorized copying under Title 17, United States Code.

ProQuest LLC
789 E. Eisenhower Parkway
PO Box 1346
Ann Arbor, MI 48106-1346
DEDICATION

This thesis is dedicated to the memory of Dr. Anthony J. DiMauro

December 24, 1922 - June 20, 2007

You never stopped believing in my abilities,
and were always so proud of my accomplishments.

Thank you Grampa.
ACKNOWLEDGEMENTS

It was with the support and hard work of many people that that this project was completed. My sincere thanks go first to Joe Licciardi for his encouragement, advice and support of this project from inception to completion. I am eternally grateful to Joe for allowing me such a great deal of freedom in designing this project from the ground up based on scientific curiosity and a desire to undertake research in remote Alaska. His assistance in both the field and laboratory portions of this project were essential. I thank Joe for always having an open door, even when his schedule was busy. Without his tireless attention to detail and unwavering encouragement during the writing process, this project would lack the integrity that it now has. My thanks are also extended to my committee members Jason Briner and Wally Bothner for introducing me to the wonders of glacial geology in Alaska and for teaching this former drill rig operator to map and interpret geology in the field from a new perspective.

Special thanks are extended to my parents and grandmother for never doubting my abilities and always believing in me. Thank you to my friends and family for always encouraging me to follow my dreams. Thanks to Jenn Sell, Ryan Cassotto and Krista Reichert for providing invaluable assistance in the field in an often challenging environment, and to Jenny Locke for always being there for me.

Funding for this project was generously provided by the John T. Dillon Alaska Research Award from the Geological Society of America, the University of New Hampshire, Department of Earth Sciences, and a grant from Sigma Xi.
# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>DEDICATION</th>
<th>iii</th>
</tr>
</thead>
<tbody>
<tr>
<td>ACKNOWLEDGEMENTS</td>
<td>iv</td>
</tr>
<tr>
<td>LIST OF TABLES</td>
<td>vii</td>
</tr>
<tr>
<td>LIST OF FIGURES</td>
<td>viii</td>
</tr>
<tr>
<td>ABSTRACT</td>
<td>ix</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>CHAPTER</th>
<th>PAGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>I. INTRODUCTION</td>
<td>1</td>
</tr>
<tr>
<td>1.1 Project overview</td>
<td></td>
</tr>
<tr>
<td>1.2 Study location</td>
<td></td>
</tr>
<tr>
<td>1.3 Geologic setting</td>
<td></td>
</tr>
<tr>
<td>1.4 Climate and glacial history of Alaska</td>
<td>7</td>
</tr>
<tr>
<td>1.5 Glacial and climate history of the Delta River Valley</td>
<td>11</td>
</tr>
<tr>
<td>II. METHODS</td>
<td>14</td>
</tr>
<tr>
<td>2.1 Field mapping and sample sites</td>
<td>14</td>
</tr>
<tr>
<td>2.2 Dating methods of surficial deposits</td>
<td>15</td>
</tr>
<tr>
<td>2.2.1 Cosmogenic $^{10}$Be surface exposure dating</td>
<td>16</td>
</tr>
<tr>
<td>2.2.2 Lichen age determination (Rhizocarpon sp.)</td>
<td>19</td>
</tr>
<tr>
<td>2.3 Glacier modeling and paleoclimate inferences</td>
<td>21</td>
</tr>
</tbody>
</table>
## LIST OF TABLES

<table>
<thead>
<tr>
<th>TABLE</th>
<th>PAGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.1 Climate observations from meteorological stations</td>
<td>24</td>
</tr>
<tr>
<td>3.1 Beryllium sample data</td>
<td>28</td>
</tr>
<tr>
<td>3.2 Sample $^{10}$Be ages</td>
<td>28</td>
</tr>
<tr>
<td>3.3 Lichenometry data and corresponding ages</td>
<td>30</td>
</tr>
<tr>
<td>B.1 Snow depth and density observations</td>
<td>70</td>
</tr>
<tr>
<td>B.2 Mean summer temperature</td>
<td>71</td>
</tr>
<tr>
<td>B.3 Total annual accumulation</td>
<td>71</td>
</tr>
<tr>
<td>C.1 $^{10}$Be concentrations from AMS</td>
<td>73</td>
</tr>
<tr>
<td>C.2 $^{10}$Be ages using multiple scaling schemes</td>
<td>73</td>
</tr>
<tr>
<td>C.3 Snow shielding calculations</td>
<td>75</td>
</tr>
</tbody>
</table>
LIST OF FIGURES

<table>
<thead>
<tr>
<th>FIGURE</th>
<th>PAGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1 Location map of Alaska</td>
<td>3</td>
</tr>
<tr>
<td>1.2 Elevation model of the Delta River Valley</td>
<td>5</td>
</tr>
<tr>
<td>1.3 Little Ice Age glacier activity in Alaska</td>
<td>10</td>
</tr>
<tr>
<td>2.1 Lichen growth curve for the central Alaska Range</td>
<td>20</td>
</tr>
<tr>
<td>2.2 Temperature- and precipitation-elevation relationships</td>
<td>25</td>
</tr>
<tr>
<td>3.1 Geomorphic map of the Castner Glacier terminus</td>
<td>27</td>
</tr>
<tr>
<td>3.2 Unit QPg $^{10}$Be ages</td>
<td>29</td>
</tr>
<tr>
<td>3.3 South lateral moraine of Castner Glacier</td>
<td>32</td>
</tr>
<tr>
<td>3.4 Modern and past total area and ELA of the Castner Glacier</td>
<td>36</td>
</tr>
<tr>
<td>3.5 JJA and annual accumulation relationships of 70 alpine glaciers</td>
<td>37</td>
</tr>
<tr>
<td>4.1 Age of unit QPg with NGIP and GISP2 ice core $\delta^{18}$O records</td>
<td>42</td>
</tr>
<tr>
<td>4.2 Age of units QHg1 and QHg2 with regional and global proxy records</td>
<td>47</td>
</tr>
<tr>
<td>4.3 Temperature observations from 1943 to 2007 in Alaska</td>
<td>50</td>
</tr>
<tr>
<td>E.1 Previous surficial map of Canwell and Castner Glaciers</td>
<td>78</td>
</tr>
</tbody>
</table>
ABSTRACT

A LATE GLACIAL AND HOLOCENE CHRONOLOGY OF THE CASTNER GLACIER, DELTA RIVER VALLEY, ALASKA

by

Michael W. Howley

University of New Hampshire, December, 2008

New field mapping of surficial deposits near the Castner Glacier, Alaska has identified three distinct moraine complexes beyond the current glacier margin, denoting at least three separate intervals of glacial advance or stillstand. The timing of moraine stabilization and ice retreat was determined by cosmogenic $^{10}$Be surface exposure dating for the two older moraines along with lichenometric measurements for all three moraines. Surface exposure ages indicate the timing of ice retreat after the late Wisconsin maximum at $14.7 \pm 0.7 \text{ ka} (n = 4, \text{k-years before present})$, and identify the intermediate moraine as corresponding to the Little Ice Age (LIA) at $1627 \pm 32 \text{ years AD} (n = 2)$. Lichen measurements from the youngest moraine indicate an age of $1842 \text{ AD}$, corresponding to the late LIA. The lichen data agree to within 50 years of exposure dating results for the older LIA moraine complex, but greatly underestimate the timing of Late Pleistocene ice retreat, indicating a limit of the useful time range for lichen dating using $Rhizocarpon (sp.)$ in this region. There is no evidence for a Younger Dryas re-
advance of the Castner Glacier, indicating that the glacial response to this climate fluctuation was less pronounced than that caused by LIA cooling. The increase in elevation of the equilibrium line altitude (ELA) of the Castner Glacier from the late LIA to the present is $120 \pm 20$ m. The climate change that could force this rise in ELA is reconstructed to be either a 0.68 to 0.81 °C increase in JJA temperature, or a decrease of 8.9 to 12.5 cm H$_2$O annual accumulation.
CHAPTER I

INTRODUCTION

1.1 Project overview

Numerous reconstructions of Late Pleistocene and Holocene climate variability in Alaska indicate the occurrence of large climate shifts on millennial to decadal time scales. Details regarding the timing, magnitude, and geographic extent of these climate fluctuations are well known in only a few key localities in the interior of the state. The dynamic responses of Alaska’s glacier systems to climate variability and the corresponding geologic deposits provide an important proxy record of past environmental conditions. An improved understanding of these climatic and glaciological parameters is essential for predicting the effects of future climate warming in Arctic and Subarctic regions. (IPCC 2007, Contributions of Working Group 4).

An important record of climate variability in Alaska that spans the Late Pleistocene and Holocene is based on regional records of glacier activity (e.g. Denton and Karlén, 1973a; Péwé, 1975; Calkin, 1988; Hamilton, 1994; Calkin et al., 2001; Briner et al., 2002, 2005; Kaufman and Manley, 2004; Briner and Kaufman, 2008; Wiles et al., 2008). These studies are supported by proxy records developed from many other sources such as lake level reconstructions (Abbott et al., 2000), δ^{18}O from inorganic and biogenic silica in lake cores (Anderson et al., 2001; Hu et al., 2001, 2003, 2006; Kaufman et al.,
2003; MacKay et al., 2008) and pollen assemblages from lake cores (Bigelow and Edwards, 2001). Climate reconstructions based on these glacial and proxy records show fluctuations that correspond to events recognized at many locations in the Northern Hemisphere, including the global Last Glacial Maximum (LGM, 21 ± 2 ka), Bølling-Allerød warm period (14.7 to 12.9 ka), Younger Dryas cooling (12.9 to 11.6 ka), Neoglacial cooling (3.5 to 2.4 ka), First Millennium AD cooling (FMA, 1.7 to 1.2 ka), Medieval warm period (1.1 to 0.8 ka) and Little Ice Age cooling (1500 to 1900 AD) (Mayewski et al., 2004).

At present, the climate information contained in records of glacial extents and volumes in Alaska has been only partially deciphered. The vast size of Alaska and the contrasting climate regimes that dominate its various regions have resulted in asynchronous chronologies of glacial advance and recession. Additionally, records and maps of glacial extents are of limited use for climate reconstruction without accurate knowledge of the timing of deposition. Glacier systems react to the first-order controls of temperature and precipitation and are sensitive indicators of climate change on sub-millennial timescales (Paterson, 1994; Oerlemans et al., 1998). Paleoclimatic inferences are commonly based on the extent of past mountain glaciers, hence evidence of glacier advance and recession constitute an independent method of corroborating various climate reconstructions (Oerlemans, 2005).

The research strategy employed in this study involves surficial mapping of glacial deposits at the Castner Glacier in conjunction with the establishment of precise age control. The resulting chronology of glacial fluctuations allows a test of whether or not the timing of moraine deposition correlates with the timing of regional and global climate
fluctuations that have occurred since the LGM (21 ± 2 ka).

1.2 Study location

This study is focused on the Castner Glacier in the Delta River Valley at 63.4°N, 147.7°W in the central Alaska Range of interior Alaska. The Alaska Range extends for 900 km from the Canadian border to the base of the Aleutian Island chain in southwestern Alaska. The north-flowing Delta River bisects the Alaska Range 240 km

Figure 1.1 Location map of Alaska showing the major mountain ranges and the location of the Delta River Valley (black box) in southeastern interior Alaska. Base map modified from Alaska State Geospatial Data Clearinghouse, Alaska Albers projection, NAD 27 datum.
west of the Canadian border (Figure 1.1). Mountains of the central Alaska Range form the climate transition zone between the moist maritime air masses that dominate southern Alaska and the dry continental climate of interior Alaska (Calkin, 1988). The study area along the Richardson Highway and Trans-Alaska Pipeline route is located at 740 meters above sea level (m asl) with the surrounding mountain peaks ranging from 1700 to 3200 m asl. Castner Glacier is the northernmost of three valley glaciers, including the Fels and Canwell Glaciers, which occupy east-west trending valleys and flow west from higher elevations east of the valley floor. The Castner Glacier is presently a 59.1 km² dendritic valley glacier that occupies an ice-scoured trench north of the Denali Fault (Figure 1.2) and exists between elevations of 790 m and 2560 m asl. The terminal area is an extensive ice-cored ablation moraine, indicating that near-terminus ice is stagnant. Castner Glacier was historically observed to be at its most recent end moraine and thinning vertically until at least 1950 AD (Mendenhall, 1900; Nielsen and Post, 1953).

The Castner Glacier was chosen as the focus of this study for both its climatically strategic location and the lack of conclusive data constraining the timing of its fluctuations. A well-constrained history of glacial activity in this climatically sensitive region will allow comparison to records developed at glaciers in other regions of Alaska dominated by contrasting climate regimes. Comparisons between these glacial records will provide insights into past atmospheric circulation patterns and temperature variations that have dominated the region. Additionally, it is notable that the Castner Glacier is one of Alaska’s most accessible glaciers owing to its location near one of Alaska’s few major highways (Nielsen and Post, 1953). Results from this project present an opportunity for
developing the study site as a roadside attraction that conveys geological information to tourists interested in the glacial history of the Castner Glacier and Alaska.

Figure 1.2 Digital elevation model of the Delta River Valley in the vicinity of the Castner Glacier (blow-up of the region in the box in Figure 1). The locations of the Castner, Fels, Canwell and Black Rapids Glaciers are shown in blue and the trace of the right-lateral Denali Fault is shown in red. WR and YT indicate the tectonic provinces of Wrangellia and Yukon-Tanana, respectively, that are found south and north of the Denali Fault.

1.3 Geologic Setting

The study area is located at the margin of two major tectonic provinces in Alaska: the Yukon Tanana terrane (YT) to the north and the Wrangellia terrane (WT) to the
south. The YT in this region is characterized by Precambrian pelitic schist with localized quartz veins and intrusions of Mesozoic andesite and dacite. Lithologies in this terrane are severely folded and have been subjected to multiple metamorphic events (Nokelberg et al., 1992). The rocks that characterize the WT are early Mesozoic basalt, andesite and marine sedimentary rocks intruded by late Cretaceous granite. The WT is thought to be the remnant of an island arc that was sutured onto North America during the late Mesozoic (Nokelberg et al., 1994). The Denali Fault is an recently active right-lateral strike-slip fault that bisects the upper Delta River Valley 2 km south of the present terminus of the Castner Glacier and forms the boundary between these two terranes (Figure 1.2). Along strike of the fault, a ~2 km wide, near-vertical stock of quartz diorite is intruded between the YT and WT.

The Denali Fault accommodates a fraction of the oblique collision of the Yakutat block into the southern Alaska margin. The Alaska Range is adjacent to the Denali Fault, and the high topography of the range may be related to thrust faults that merge into the Denali Fault at depth (Nokelberg et al., 1994). Evidence suggests 38 km offset of plutons in the past 38 my, with 3 to 4 km offsets along major glacial valleys during the last 3.5 my. The modern (Holocene) slip rate has been measured at $12.1 \pm 1.7 \text{ mm yr}^{-1}$ by $^{10}$Be exposure dating of offset moraines (Matmon et al., 2006). An earthquake (M=7.9) occurred on 3 November, 2002 that caused 340 km of surface rupture, including 5.3 m of lateral offset and 1.5 m vertical offset (north side up) in the Delta River Valley (Eberhart-Phillips et al., 2003). The bedrock geology of the region is relevant to this glacial study because the boulder and bedrock lithologies provide quartz for $^{10}$Be exposure dating and stable substrates for lichen growth. The location of the Denali Fault forms a structural
trench that controls the straight trend of the Canwell Valley. The vertical component of Holocene motion along the Denali Fault has preserved glacial deposits that would have otherwise been modified by fluvial action of the Delta River.

1.4 Climate and glacial history of Alaska

Alaska encompasses the largest contiguous high latitude land area in North America that was not glaciated during the Cenozoic. The greatest concentration of glaciers in Alaska presently, as in the Pleistocene, is in the southern part of the state closest to moisture sources in the North Pacific. Glaciers also occur in the high peaks of the Brooks Range in the interior of Arctic Alaska. Altogether, glaciers once covered about 1,200,000 km$^2$ of Alaska and its adjacent continental shelf; during the late Wisconsin (MIS 2), the total glacier area was 727,800 km$^2$. At present 74,700 km$^2$ of Alaska is covered by ice, or 4.9% of the state (Kaufman and Manley, 2004). During global glacial intervals when eustatic sea level was lower, vast areas of the continental shelf were exposed. The regressed shoreline reduced the maritime influence that dominates the state today, and increased the distance between glaciers and their moisture sources by as much as 800 km. Increased sea-ice cover from lower atmospheric temperatures would have further reduced the availability of moisture (Hamilton, 1994; Kaufman and Manley, 2004). This increase in continentality explains why the late Wisconsin (MIS 2) glacial deposits in Alaska are less extensive than deposits from the penultimate (MIS 6 or MIS 4/3) glaciations (Kaufman and Manley, 2004; Briner et al., 2005; Briner and Kaufman, 2008). The timing of late Wisconsin moraine stabilization has been determined by a combination of $^{14}$C ages and cosmogenic $^{10}$Be exposure dates
from six regions in Alaska. Glaciers retreated from their local late Wisconsin maxima by: c. 24 to 27 ka, Kokrines Hills (west-interior Alaska); c. 24 to 26 ka, northeastern Brooks Range (NE Alaska); c. 21 to 23 ka, Yukon Tanana Upland (east-interior Alaska); c. 22 ka, Ahklun Mountains (SW Alaska); c. 20 ka, western Alaska Range (central Alaska); and c. 16 to 18 ka, Chuilnuk Mountains (SW Alaska) (Briner et al., 2005). The timing of glacier retreat from these late Wisconsin maximum positions varies by region, with the Yukon Tanana Uplands, Alaska Range and Ahklun Mountains concurrent with the peak of the LGM (21 ± 2 ka).

The Mount Waskey moraine in the Ahklun Mountains of southwest Alaska is the only well-constrained glacial deposit outside the Alaska Range attributed to the Younger Dryas (YD) period. The age of this moraine is between 11.0 and 12.4 ka as determined by $^{10}$Be and $^{26}$Al exposure dating (Briner et al., 2002). The McKinley Park IV moraine documented in the Alaska Range by Ten Brink and Waythomas (1985) is also attributed to the YD, but only the maximum age of this advance is constrained by $^{14}$C dates. Climate reconstructions from lake sediments and pollen assemblages indicate cooler and drier conditions during this period (Hu et al., 1995; Bigelow and Edwards, 2001; Kaufman et al., 2003), but the corresponding glacial record from Alaska is incomplete.

Multiple Holocene glacier advances have been described from various mountain ranges in Alaska (e.g., Denton and Karlén, 1973a; Péwé, 1975; Calkin, 1988; Hamilton, 1994; Calkin et al., 2001; Wiles et al., 2002, 2008). Denton and Karlén (1973a) were among the first to systematically investigate Holocene climate fluctuations in both North America and Europe. Their survey of the advance and recession of 51 North American and European alpine glaciers indicated three major intervals of Holocene glacial
expansion: 3400 to 2400 years BP (Neoglaciation), 1700 to 1100 years BP (FMA) and 1500 to 1900 AD (LIA). Studies conducted over the past 35 years in many locations in both Alaska and throughout the Northern Hemisphere yield glacial chronologies that fit in well with Denton and Karlén's framework with minor local timing and magnitude variations (Calkin, 1988; Mayewski et al., 2004; Wiles et al., 2008).

General trends in Holocene climate reconstructions in Alaska indicate warming in the early Holocene and possibly a complete disappearance of all glaciers in Alaska during the Holocene Thermal Optimum (11 to 9 ka) (Kaufman et al., 2004). The occurrence of this warmer and drier climate is supported by pollen assemblages, lake level reconstructions, $\delta^{18}O$ from lake cores and a lack of glacial deposits from this period (Abbott et al., 2000; Bigelow and Edwards, 2001; Calkin, 1988; Hu et al., 2003; Mayewski et al., 2004). Beginning around 5000 to 4000 years BP, large-scale cooling and increases in precipitation preceded the advance or re-growth of glaciers in many regions in Alaska. Neoglacial advances have been documented as the re-growth of high altitude mountain glaciers by researchers in the coastal Kenai and Chugach Mountains between 3600 and 3000 years BP (Calkin et al., 2001), the Wrangell and northern St. Elias Mountains between 2700 and 2000 years BP (Wiles et al., 2002), Icy Bay near the southern St. Elias Mountains between 3750 and 3250 years BP (Barclay et al., 2006), and the Arctic Brooks Range between 4900 and 2900 years BP (Calkin, 1988). Climate warming is inferred after 2300 years BP (Mayewski et al., 2004), followed by FMA glacial advances recorded in the Kenai and Chugach Mountains between 1500 and 1300 BP, the Wrangell Mountains between 1650 and 1500 years BP, the St. Elias Mountains between 1230 and 1050 years BP, Icy Bay between 1525 and 900 years BP (Reyes et al.,
Figure 1.3 Summary of documented Little Ice Age glacier activity from the major mountain ranges in Alaska. Modified after Wiles et al., 2004, with individual glacial chronologies from Calkin, 1988 (Brooks Range); Denton and Karlen, 1973a and Wiles et al., 2002 (Wrangell Mountains); and Calkin et al., 2001 (Coastal Ranges).

2006 and references therein), and the Brooks Range between 1700 and 1100 years BP (Calkin, 1988). The FMA glacial expansion was largely concluded by ~1000 years BP and followed by the Medieval Warm Period, which is observed in many lake-core proxy records of temperature in central Alaska (e.g. Hu et al., 2001). Glacial advances during the LIA (Figure 1.3) began as early as 1400 AD in the Brooks Range and by 1600 AD in the Wrangell Mountains and Coastal Ranges.
1.5 Glacial and climate history of the Delta River Valley

The glacial and climate history of the Delta River Valley has been the subject of many studies, beginning with pioneering work in the 1950's (Péwé, 1953; Nielsen and Post, 1953). During the late Pleistocene, glaciers flowed north and south from an ice divide centered over the Alaska Range (Hamilton and Thorson, 1983). Southward flowing ice coalesced with the Cordilleran Ice Sheet and extended into the Gulf of Alaska. Northward-flowing ice formed lobes in separate topographic lows that terminated relatively short distances (10 - 20 km) beyond the Alaska Range front. The maximum limit of glaciers (penultimate) and the more restricted MIS 2 (late-Wisconsin) extents along the northern Alaska Range in the Delta River Valley were first mapped by Péwé (1953) as the Delta and Donnelly Glaciations, respectively, and published in the map compiled by Coulter et al. (1965). The age of the Donnelly Glaciation, which is thought to be equivalent to other named late-Wisconsin (MIS 2) glaciations around the state, is constrained by \(^{14}\text{C}\) and \(^{10}\text{Be}\) ages that suggest the terminal moraines were deposited around 21 to 20 ka (Porter et al., 1983; Hamilton, 1994; Briner and Kaufman, 2008).

Climate in the Alaska Range during the late glacial period has been reconstructed from a relatively small number of records spanning this time. Lake cores from two locations in the Alaska Range have yielded high resolution records that extend back to \(~15\) ka in the form of lake level fluctuations at Birch Lake (110 km NNW of Castner Glacier, Abbott et al., 2000) and pollen reconstructions from Windmill Lake (150 km WNW of Castner Glacier, Bigelow and Edwards, 2001). These records indicate cool and dry conditions prior to 15.0 to 13.7 ka, followed by warmer and wetter conditions. A
return to cooler and drier conditions is reconstructed from both of these studies during the Younger Dryas chronozone, followed by warmer conditions in the early Holocene (Abbott et al., 2000; Bigelow and Edwards, 2001). Few moraines have been documented between the late Wisconsin glacial maximum positions and the culminating Holocene extents in the Alaska Range. Moraines located upvalley of late Wisconsin maximum positions that were offset by the Denali Fault in five valleys of the central Alaska Range were recently exposure-dated using cosmogenic $^{10}\text{Be}$ to determine slip rates (Matmon et al., 2006). One of the five moraines studied is within 2 km of the modern glacier terminus is dated to $12.0 \pm 1.3$ ka ($n = 7$) (site DFCR of Matmon et al., 2006). The McKinley Park IV glaciation in Denali National Park, 100 km west of the study area in the Alaska Range, has also been dated to 12.3 to 11.0 ka by radiocarbon dating (Ten Brink and Waythomas, 1985). Both of these moraines may represent Younger Dryas re-advances in the Alaska Range, but morainal evidence of Younger Dryas glacier oscillations is lacking at other key locations in this region.

Among all the valley glaciers in the upper Delta River Valley, the Holocene chronology of the Black Rapids Glacier was previously the best constrained (Figure 1.2). Efforts to date Holocene moraines at the Black Rapids, Castner and Canwell glaciers with dendrochronology, lichenometry and radiocarbon dating have yielded somewhat inconsistent results (Pévé, 1961; Reger, 1968; Reger and Pévé, 1969; Pévé and Reger, 1991; Reger et al., 1993). Three Holocene advances are documented from the Black Rapids Glacier, with the earliest Holocene advance occurring sometime after 3,360 years BP and before 1,710 years BP as indicated by radiocarbon dates on blocks of peat and a tree stump contained in till (Pévé and Reger, 1983; Reger et al., 1993). The Black
Rapids Glacier advanced twice more during the Holocene, and these later advances have been described as coeval with the two recognized Holocene advances of the Canwell and Castner Glaciers (Reger and Péwé, 1991). The intermediate Holocene moraine at Black Rapids Glacier and the outermost Holocene moraine of the Canwell and Castner Glaciers (hereafter QHg1) are likely correlative based on similar ages determined by radiocarbon evidence (Reger and Péwé, 1983; Péwé, 1987) and were built by an advance during the past 400 years (see Appendix E). On the basis of tree-ring counting of living trees and the soil development on Black Rapids moraines, Péwé estimated the age of the older Holocene moraine at 330 years BP (~1620 AD) and correlates it with coeval moraines at the Castner and Canwell glaciers (Péwé, 1961; Péwé and Reger, 1983; Péwé, 1987). Both the Black Rapids and Canwell Glaciers have been previously described as surging glaciers in the literature (Calkin, 1988). At present, there is no age control on the intermediate moraine at the Castner Glacier, aside from this tentative correlation.

The age of the innermost moraines of Black Rapids and Canwell Glaciers is constrained by radiocarbon dates and tree-ring counting, and is reported to have stabilized at about 120 years BP (1830 AD) (Reger and Péwé, 1991; Reger et al., 1993). Prior to this study, the age of the innermost Castner Glacier moraine (hereafter QHg2) was inferred to be of the same age (1830 AD) based on historical observations of glacier position (Mendenhall, 1900; Nielsen and Post, 1953) and correlation with mapped moraines at the nearby glaciers (Péwé and Reger, 1983; Reger and Péwé, 1991).
CHAPTER II

METHODS

2.1 Field mapping and sample sites

A field campaign was undertaken at the Castner Glacier during the summers of 2007 and 2008 with the dual purposes of mapping the spatial distribution of glacial deposits and determining the ages of these deposits. The chronology was developed by collecting boulder and bedrock samples for cosmogenic $^{10}$Be surface exposure dating analyses (Gosse and Phillips, 2001) along with lichen measurements. Mapping of moraines and outwash surfaces was accomplished using a hand-held GPS unit, a 1:16,000 scale orthorectified satellite photo, and a 1:63,360 scale topographical map (U.S. Geological Survey - Alaska Series, Mt. Hayes B-4). The elevation of each field site was recorded from GPS and checked by triangulation using a compass, inclinometer and known benchmarks. At each field site, information was recorded about the position relative to the modern glacier, the morphological characteristics of the feature being described such as moraine crest height and slope angle, the depth of weathering and soil development, degree of weathering and lithology of clasts and boulders, and vegetation species present. These criteria were used to distinguish older from younger units and identify the boundaries between those units in the field. Different glacial units are composed of tills of different degrees of weathering which aided in field identification.
Unit boundaries are generally sharp between younger deposits and more gradational for older deposits. Field sites and sample locations were transferred to the digitized base map by uploading recorded GPS locations and displaying these points on both the orthorectified satellite photo and elevation model generated from the U.S. Geological Survey topographical map. Data recorded in the field were used in the construction of a map (see section 3.1) showing the location and morphostratigraphic positions of surficial units.

2.2 Dating methods of surficial deposits

In order to precisely determine the ages of the surficial units mapped at the Castner Glacier, a combination of dating methods was applied. The application of multiple independent dating methods has the advantage of allowing greater confidence in the unit ages. All ages yielded from all dating methods applied in this study are in years before sample collection (2007). Previous studies report ages derived from different methods in different formats ($^{10}$Be years for $^{10}$Be surface exposure ages, years BP (1950) for $^{14}$C ages and tephrochronology, and L-years for lichen-based ages). This complicates direct comparison of these age data. For this study, we report new $^{10}$Be surface exposure ages older than 10,000 years in thousands of years (ka), given that typical uncertainty in ages $>$10 ka are much greater than the difference between ages reported with respect to 1950 AD or 2008 AD. For deposits and surfaces younger than 10,000 years, we report both newly determined and previously published ages in Roman calendar year to allow comparison to other high resolution data sets and to accurately report the more precise younger ages determined by $^{10}$Be surface exposure dating (Wolff, 2007).
2.2.1 Cosmogenic $^{10}$Be surface exposure dating

The recent advent of cosmogenic exposure dating methods has revolutionized dating of Quaternary deposits and surfaces, and has permitted previously unattainable insights into rates and styles of surficial processes (Gosse and Phillips, 2001). Cosmogenic $^{10}$Be surface exposure dating (hereafter “surface exposure dating”) has been applied in many glaciated settings to directly determine the age of moraines and other features deposited or exposed by glaciers (e.g., Gosse et al., 1995; Ivy-Ochs et al., 1999; Briner et al., 2002, 2005; Licciardi et al., 2004; Licciardi and Pierce, 2008).

Six samples were collected with hammer and chisel from boulder tops and bedrock outcrops in the study area for the purpose of determining the age of moraine stabilization and timing of deglaciation for bedrock outcrops. Samples were selected based on their position on moraine crests, boulder shape and height. Samples MH07-22, -28, -29, and -30 were collected from boulders composed of mica schist with localized quartz veins. Samples MH07-24 and -26 were chiseled from quartz diorite bedrock. Beryllium was extracted from these rock samples to produce purified BeO target material for accelerator mass spectrometry (AMS) analysis, following a modified version of procedures developed by Licciardi (2000). Samples were crushed to a 250–600 μm size fraction, and then leached first in 5% HCl and then in 5% HNO₃ for pre-cleaning. Quartz was then isolated by repeated leaching in a 2%-HF/1%-HNO₃ solution. $^9$Be carrier (~0.2 mg) was added to ~40 g of purified quartz from each sample before dissolution in concentrated HF. Beryllium was then extracted from the samples using ion-exchange chromatography, selective precipitation, and oxidation to BeO. $^{10}$Be/$^9$Be ratios were measured at the Center for Accelerator Mass Spectrometry at the Lawrence Livermore
National Laboratory (LLNL) (Davis et al., 1990). The $^{10}\text{Be}/^{9}\text{Be}$ ratios reported by LLNL were corrected for blank activity levels and converted to atoms of $^{10}\text{Be}$ per grams of quartz.

All age calculations were derived using the CRONUS Earth online exposure age calculator available online at http://hess.ess.washington.edu/math (Balco et al., 2008). Factors such as topographic shielding, snow shielding, and surface erosion were evaluated according to previously published methods explained below (see Gosse and Phillips, 2001). The sea-level high-latitude ($\geq 60^\circ$) spallogenic $^{10}\text{Be}$ production rate of $4.96 \pm 0.43$ atoms g$^{-1}$ yr$^{-1}$ SiO$_2$ ($\pm$1$\sigma$), and muonic $^{10}\text{Be}$ production according to Heisinger et al. (2002a, b) were used in the age calculations (Balco et al, 2008). The site specific $^{10}\text{Be}$ production rates of 9.91 to 11.85 atoms g$^{-1}$ yr$^{-1}$ were determined for the study area using the altitude, latitude and atmospheric scaling equations of Stone (2000) following Lal (1991). Adjustments for topographic shielding were determined using measurements recorded at each sample site and the CRONUS Earth online topographic shielding calculator (Balco et al., 2008). The orientation of major topographical features and elevation-to-horizon angle were recorded by compass and inclinometer at the time of sample collection at all sample sites.

The effect of possible shielding by snow cover was evaluated based on snow depth and density recorded at the nearby Trims Camp meteorological station. Snow depth and snow water equivalent (SWE) measurements are available at weekly intervals between 1953 and 1979 at this location 2.6 km northwest of the Castner Glacier terminus. The Trims Camp station is within 150 m of the elevation of all sample sites in this study and is assumed to provide a reliable constraint on snow cover at the field sites. Snow
depth was measured to be greater than 0.5 m for 5 months of the year and the net effect of snow shielding in this vicinity is estimated at ~8% (Appendix B). This empirically-derived snow shielding adjustment is twice the estimated shielding effect of 4 months of snow cover 1 m deep, corresponding to a 4% increase in exposure ages that has been used by other studies in Alaska (e.g. Briner et al., 2002). The calculated 8% age adjustment likely overestimates the snow shielding due to the effect of wind clearing snow from protruding boulders and bedrock outcrops sampled in this study, which range in height from 0.7 to 4.0 m. To account for this, we used an age adjustment of 5% as an intermediate value for snow shielding. This is based on the maximum possible amount of snow shielding on the shortest boulder or bedrock outcrop sampled in this study (0.7 m). The exposure ages discussed in the text indicate whether or not an adjustment was made for snow shielding.

Based on observed mineral etching and surface erosion of sampled boulder and bedrock surfaces, we used an erosion rate of 3 mm ka\(^{-1}\) after Briner et al. (2002). This value is less than the maximum erosion rate of 4.8 mm ka\(^{-1}\) determined for the Yukon-Tanana Uplands as suggested by measurements of both \(^{10}\)Be and \(^{26}\)Al in boulders (Briner et al., 2005). The net age adjustments due to the effect of 3 mm ka\(^{-1}\) erosion increase exposure ages by up to 3.7%, according to the CRONUS Earth online exposure age calculator (Balco et al., 2008).

Temporal changes in geomagnetic field intensity, the position of the geomagnetic dipole axis, and solar modulation affect the intensity of cosmic-ray flux at the Earth's surface and thus modulate cosmogenic nuclide production rates (Gosse and Phillips, 2001). To compensate for these variations, we report ages determined in this study based
on the constant production rate scaling scheme of Lal (1991)/Stone (2000), as well as the
time-varying production rates of Desilets et al. (2003), Dunai (2001), Lifton et al. (2006)
are based on the commonly-used Lal (1991) / Stone (2000) scaling unless otherwise
specified. Uncertainties of individual surface exposure ages, reported at 1σ, represent
analytical uncertainties in accelerator mass spectrometry (AMS) measurements and blank
corrections for $^{10}\text{Be}$, but do not include the $\sim10\%$ uncertainty in the $^{10}\text{Be}$ production rate.

### 2.2.2 Lichen age determination (Rhizocarpon sp.)

Lichen measurements have been used as a method for dating surficial deposits in
numerous regions of Alaska since the 1960's (Reger and Péwé, 1969; Denton and Karlen,
1973b; Ten Brink, 1983; Solomina and Calkin, 2003). *Rhizocarpon geographicum* is a
yellow-green crustose lichen species commonly found in mountainous areas with low
levels of air pollution. Each lichen colony or thallus is bordered by a black line of spores
forming a boundary that distinguishes individuals and enables accurate measurement of a
single thallus. *Rhizocarpon geographicum* is difficult to positively identify to the species
level in the field, and hereafter will be referred to as *Rhizocarpon sp.* to account for the
probable mis-identification or inclusion of other lichen species. The calibration curves
used in this study are also based on *Rhizocarpon sp.* and not necessarily on
*geographicum* (Solomina and Calkin, 2003). Lichenometry is based on the assumption
that the largest lichen growing on a feature is the oldest individual. In regions of Alaska
where the lichen growth rate is calibrated, equations have been developed to allow
conversion of a lichen measurement to an age (Denton and Karlén, 1973b; Begét, 1994; Solomina and Calkin, 2003).

Figure 2.1 Lichen growth curves for the central Alaska Range with the calibration points (blue diamonds) reported by Solomina and Calkin (2003). The logarithmic curve (shown in red) was modified by Begét (1994) for the central Alaska Range from the curve constructed by Denton and Karlén (1973b) for the Wrangell-St. Elias Mountains. The compound linear curve (shown in black) was published by Solomina and Calkin (2003) but neither model is supported by any data between -300 to 3700 years. The equations for the two different calibrations were reported by Solomina and Calkin (2003). Although both curves are shown, ages reported in this study are calculated using the logarithmic curve.

A total of 326 lichen measurements were recorded from 26 separate lichen measurement stations on three separate glacial deposits in the terminus region of the Castner Glacier. At each measurement station, the long axes of the 10 to 15 largest circular or semi-circular thalli were measured on all stable boulders and exposed bedrock surfaces. The size of the largest lichen found was used as an age indicator for the site.
(Beschel, 1961; Jomelli et al., 2007). A calibration curve of lichen growth rate for this region was originally developed using tree-ring counting (Péwé, 1961) and later refined with $^{14}$C dates (Péwé and Reger, 1983). This curve was later abandoned by Péwé and Reger (1983) in favor of the curve developed by Denton and Karlén (1973b) and modified by Begét (1994) by incorporating a control point based on a layer of Jarvis Creek Ash buried in morainal material (Begét, 1994). The most recent compilation of lichen studies in Alaska and Siberia (Solomina and Calkin; 2003) used measurements of lichens growing on historically dated surfaces and mine tailings of known age to refine the youngest part of the growth curve in the central Alaska Range. Two different curves are available for the central Alaska Range, a logarithmic curve and a compound linear curve (Figure 2.1). Both of these calibrations are based on the single largest lichen method. For this study, we report all lichenometric ages in Roman calendar year according to the logarithmic calibration curve compiled by Solomina and Calkin (2003). Solomina and Calkin (2003) also proposed a compound linear calibration curve for this region, but this is not supported by results found in this study or in other lichen studies from the Delta River Valley.

2.3 Glacier modeling and paleoclimate reconstruction

2.3.1 Reconstruction of past glacier equilibrium line altitudes

Paleoclimatic reconstructions based on the limits of former glaciers commonly make use of estimates of the associated equilibrium line altitude (ELA). Variations in the ELA are strongly linked with local climate, particularly annual accumulation and summer temperatures (Benn and Lehmkuhl, 2000). This relationship enables the calculation of
these climate parameters from a reconstructed ELA position.

The innermost moraine associated with the most recent stillstand of the Castner Glacier is exceptionally well preserved, with lateral and terminal moraines clearly marking the extent of ice at time of deposition. This provides the basis for modeling the steady-state ELA of Castner Glacier at the time of deposition of this moraine. ELA positions associated with the older stillstands of the Castner Glacier were not reconstructed because the terminal zones of the Castner, Fels and Canwell Glaciers coalesced at the culmination of these advances. Two methods were applied to reconstruct the elevation of the former ELA: accumulation area ratio (AAR) and maximum elevation of lateral moraines (MELM).

The AAR method relies on the assumption that the accumulation area of the glacier (i.e., the area above the ELA) occupies a fixed proportion of the total glacier area (Meier and Post, 1962; Porter, 1975; Benn and Lehmkuhl, 2000). The AAR used in this study for the Castner Glacier is 0.61 (accumulation area / total area), a value which has been determined from ongoing mass-balance studies conducted by the U.S. Geological Survey at the nearby Gulkana Glacier between 1966 and 2007 (Josberger et al., 2007). The Gulkana Glacier has a debris covered terminal region and nearly identical climate parameters as the Castner Glacier and provides a reasonable constraint on this value..

The MELM associated with the most recent advance of the Castner Glacier was determined using GPS-based field mapping as well as remote mapping on a high-resolution satellite photo. The MELM method is based on the fact that the flow of a glacier from its accumulation zone to its ablation zone results in a net transfer of debris from the accumulation zone, along the center line of the glacier and then out towards the
margins of the glacier in the ablation zone. The result of this debris transport path is that lateral moraines are deposited only below the ELA (Andrews, 1975; Benn and Evans, 1998). Lateral moraines deposited during the youngest stillstand were distinguished from recessional moraines deposited later at higher elevations.

2.3.2 Paleoclimate Inferences

Reconstructions of climate change provide insight into the response time and sensitivity of a glacier system to climate forcing. Ohmura et al. (1992) investigated climate conditions at the ELA of 70 modern mid- and high-latitude glaciers and found that glaciers exist within narrow boundary conditions of mean summer temperature (JJA temperature) and annual accumulation (summer precipitation + winter accumulation in cm H$_2$O) at their ELA. Climate conditions at the modern ELA of the Castner glacier were determined using temperature and precipitation lapse rate equations derived from meteorological data. The shift in ELA since deposition of the innermost moraine was used to infer the change in temperature and/or precipitation that would force the ELA to rise by the reconstructed amount. The magnitude of the climate change that would force a change in position of the ELA is determined by assuming that temperature and precipitation gradients have remained the same over time.

JJA temperature at the modern and past ELA was determined as a function of elevation using an equation describing the temperature lapse rate in this region. This equation was developed from mean monthly temperature observation recorded at three meteorological stations at various elevations in the region (Table 2.1). The JJA
temperature lapse rate was determined by fitting a linear regression line to the data points (Figure 2.2). The relationship between JJA temperature and elevation in this region is described by the equation: $\text{JJA Temperature (°C)} = -6.214 \times (\text{elevation (km)}) + 14.848$. The $r^2$ value of this fit is 0.96 which indicates a strong linear relationship between temperature and elevation. The temperature lapse rate in this region approximates the global adiabatic lapse rate of 6.5 °C km$^{-1}$. Using this equation, I extrapolated the mean JJA temperature upwards in elevation to the present (2007 AD) and past ELA. The difference in JJA temperature at these two elevations was assumed to be the minimum increase necessary to force the raising of the ELA to the 2007 AD position due to shifting climate conditions.

### Table 2.1
Mean ablation season temperature and annual precipitation by observation station

<table>
<thead>
<tr>
<th>Observation Station</th>
<th>Elevation (m asl)</th>
<th>Years used in data set</th>
<th>Mean JJA Temperature (°C)</th>
<th>Mean Annual Accumulation (cm H$_2$O)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Trims Camp</td>
<td>732</td>
<td>1968 - 1979</td>
<td>9.95</td>
<td>70.3</td>
</tr>
<tr>
<td>Summit</td>
<td>985</td>
<td>1968 - 1979</td>
<td>9.25</td>
<td>----</td>
</tr>
<tr>
<td>Gulkana</td>
<td>1480</td>
<td>1968 - 1979</td>
<td>5.47</td>
<td>136.9</td>
</tr>
</tbody>
</table>

Observation station data: Trims Camp - 2.6 km northwest of modern terminus, 732 m asl, records from 1953-1979; Summit - 28.3 km south of modern terminus, 985 m asl, records from 1968-1998; Gulkana Glacier - 15.8 km southeast of modern terminus, 1485 m asl, records from 1966-2007.

Annual accumulation (cm H$_2$O) at the modern and past ELA was determined as a function of elevation using a precipitation lapse rate equation derived from empirical observations recorded at Trims Camp and Gulkana Glacier between 1968 and 1979. The mean annual accumulation measured at Gulkana Glacier during this period (99.2 cm H$_2$O) represents 62% of actual annual accumulation due to a systematic instrument inaccuracy (Kennedy et al., 1997). To account for this instrument bias, the annual accumulation used to determine the precipitation lapse rate was adjusted to 38% greater
Figure 2.2 a. Temperature-elevation relationship determined for the Castner Glacier area based on meteorological observations of mean summer temperature between 1968 and 1979. b. Precipitation-elevation relationship determined for the Castner Glacier. Data from Trims Camp, AK and Summit, AK were obtained from Western Regional Climate Center, and data from Gulkana Glacier are from U.S. Geological Survey glacier monitoring project.

than the instrument measurement. The lapse rate equation was determined from the slope of the line connecting these two points. Precipitation data are available for the Summit, Alaska observation station for this period, but the greater distance and setting (28.3 km south, outside the orographic influence of Alaska Range) precluded use of these data. Less confidence is placed in the accumulation lapse rate equation than the JJA temperature lapse rate equation due to the aforementioned concerns.
CHAPTER III

RESULTS AND IMPLICATIONS

3.1 Glacial chronology of the Castner Glacier

The field mapping and geochronological work completed in this study has resulted in the identification of three glacial units at the Castner Glacier. Two of these units comprise terminal and lateral moraine complexes denoting distinct glacial advances or stillstands of progressively more limited extent. The third unit is composed of landforms deposited and exposed by glacial retreat after the MIS 2 (late-Wisconsin) stillstand. The geomorphic character, stratigraphic relationships and age control for each unit are described below, from oldest to youngest.

3.1.1 Late-Pleistocene Glacial Deposits (OPg)

The oldest glacial unit in the vicinity of the Castner Glacier is a discontinuous gray till, weathered brown within 1 m of the surface, and containing weathered mica schist cobbles that form a ground moraine and lateral moraines. Unit QPg is best preserved above elevations of 775 m asl in areas that were not modified by later glacial advances or affected by fluvial modification from the Delta River, Castner and Lower Miller Creeks. The best preserved examples of these deposits are found on the medial
Figure 3.1 Geomorphic map of the Castner Glacier terminal area indicating unit names and boundaries, as well as $^{10}$Be exposure dating sample locations. Base map is orthorectified satellite photo taken 9-15-2006. See text or Appendix D for full unit descriptions.

moraine preserved at the former confluence of the Castner and Fels Glaciers (Figure 3.1). This moraine was previously mapped as being coeval with the older Holocene advances of the Canwell and Black Rapids Glaciers (Reger and Péwé, 1991), however new evidence found in this study disputes that interpretation.

Surface exposure ages determined from the four samples obtained from unit QPg range in age from $12.8 \pm 0.3$ to $14.2 \pm 0.3$ ka (no adjustments for effects of erosion or snow shielding, $\pm 1\sigma$ analytical uncertainty). The two samples recovered from boulders
located on the south-lateral moraine (MH07-22 and -28) yield unadjusted ages of 12.8 ± 0.3 and 13.9 ± 0.3 ka. Two samples recovered from quartz diorite bedrock knobs exposed outside the terminal moraines of the Castner Glacier (MH07-24 and -26) yield unadjusted ages of 14.2 ± 0.3 and 13.6 ± 0.3 ka (Table 3.2).

Table 3.1
Beryllium data and sample information

<table>
<thead>
<tr>
<th>Sample</th>
<th>Quartz (g)</th>
<th>Elev. (m)</th>
<th>Height (cm)</th>
<th>Thick (cm)</th>
<th>Latitude (°N)</th>
<th>Longitude (°W)</th>
<th>Shielding Factor</th>
<th>$^{10}$Be$/^{9}$Be (AMS ratio)</th>
<th>$^{10}$Be (10$^6$ atoms g$^{-1}$)</th>
<th>Blank %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Unit QPg samples</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MH07-22</td>
<td>40.4495</td>
<td>884</td>
<td>160</td>
<td>1.80</td>
<td>63.39599</td>
<td>145.70778</td>
<td>0.9868</td>
<td>4.241E-13</td>
<td>15.36 ± 0.36</td>
<td>2.30%</td>
</tr>
<tr>
<td>MH07-24</td>
<td>40.1936</td>
<td>818</td>
<td>400</td>
<td>1.75</td>
<td>63.39190</td>
<td>145.74134</td>
<td>0.9943</td>
<td>4.578E-13</td>
<td>16.21 ± 0.33</td>
<td>2.13%</td>
</tr>
<tr>
<td>MH07-26</td>
<td>39.3996</td>
<td>814</td>
<td>150</td>
<td>2.50</td>
<td>63.39233</td>
<td>145.74364</td>
<td>0.9983</td>
<td>4.184E-13</td>
<td>15.41 ± 0.37</td>
<td>2.33%</td>
</tr>
<tr>
<td>MH07-28</td>
<td>40.2366</td>
<td>888</td>
<td>160</td>
<td>1.90</td>
<td>63.39586</td>
<td>145.70746</td>
<td>0.9254</td>
<td>4.264E-13</td>
<td>15.66 ± 0.38</td>
<td>2.29%</td>
</tr>
<tr>
<td>BeBLK-21</td>
<td>0.0</td>
<td>-----</td>
<td>-----</td>
<td>-----</td>
<td>-----</td>
<td>-----</td>
<td>-----</td>
<td>2.30%</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Unit QHgl samples</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MH07-29</td>
<td>42.4628</td>
<td>875</td>
<td>70</td>
<td>1.75</td>
<td>63.39688</td>
<td>145.70871</td>
<td>0.9892</td>
<td>1.401E-14</td>
<td>0.4367 ± 0.04</td>
<td>15.07%</td>
</tr>
<tr>
<td>MH07-30</td>
<td>42.1694</td>
<td>773</td>
<td>190</td>
<td>2.15</td>
<td>63.40154</td>
<td>145.73396</td>
<td>0.9979</td>
<td>7.542E-15</td>
<td>0.2518 ± 0.04</td>
<td>27.99%</td>
</tr>
<tr>
<td>BeBLK-22</td>
<td>0.0</td>
<td>-----</td>
<td>-----</td>
<td>-----</td>
<td>-----</td>
<td>-----</td>
<td>-----</td>
<td>2.111E-15</td>
<td>-----</td>
<td>-----</td>
</tr>
</tbody>
</table>

All $^{10}$Be concentrations were determined at the Center for Accelerator Mass Spectrometry at LLNL. The $^{10}$Be data are normalized with respect to a LLNL secondary $^{10}$Be/$^{9}$Be standard (KNSTD3110) calibrated against an ICN $^{10}$Be standard. Sample thickness corrections assume an attenuation coefficient of 160 g cm$^{-2}$ and a rock density of 2.7 g cm$^{-3}$. LDEO Beryl $^{9}$Be carrier was used for samples -22, -24, -26, -28 and blank -21, and LDEO BeSO$_4$ carrier was used for samples -29, -30 and blank -22. Measurement uncertainties reflect 1 σ analytical error only.

Table 3.2
Sample and mean $^{10}$Be ages showing effects of erosion and snow shielding.

| Sample  | Unadjusted Lal (1990) / Stone (2000) | 3 mm kyr$^{-1}$ Erosion Only | Snow Shielding Only | Both Snow Shielding and 3 mm kyr$^{-1}$ Erosion |
|---------|---------------------------------|----------------------------|-----------------|---------------------------------|-----------------|
| Sample  | Age ± 1σ | Age ± 1σ | Age ± 1σ | Age ± 1σ | Age ± 1σ |
| Unit QPg samples | | | | | |
| MH07-22 | 12.8 ± 0.3 ka | 13.2 ± 0.3 ka | 13.4 ± 0.3 ka | 13.8 ± 0.3 ka |
| MH07-24 | 14.2 ± 0.3 ka | 14.7 ± 0.3 ka | 14.9 ± 0.3 ka | 15.5 ± 0.3 ka |
| MH07-26 | 13.6 ± 0.3 ka | 14.0 ± 0.4 ka | 14.3 ± 0.3 ka | 14.7 ± 0.4 ka |
| MH07-28 | 13.9 ± 0.3 ka | 14.3 ± 0.4 ka | 14.6 ± 0.4 ka | 15.0 ± 0.4 ka |
| Mean Age | 13.6 ± 0.6 ka | 14.0 ± 0.7 ka | 14.3 ± 0.6 ka | 14.8 ± 0.7 ka |
| Unit QHgl samples | | | | | |
| MH07-29 | 364 ± 30 years | 364 ± 30 years | 381 ± 32 years | 381 ± 32 years |
| MH07-30 | 229 ± 35 years | 229 ± 35 years | 240 ± 37 years | 240 ± 37 years |
| Mean Age | 297 ± 95 years | 297 ± 95 years | 311 ± 99 years | 311 ± 99 years |

Ages were determined using Lal (1991)/Stone (2000) constant production scaling scheme and the $^{10}$Be sea-level, high-latitude (≥ 60°) spallation production rate of 4.96 ± 0.43 atoms g$^{-1}$ yr$^{-1}$ (Balco et al., 2008). Unit QHgl sample ages are reported with respect to date of collection (2007 AD).
Figure 3.2 Ages ± 1σ analytical uncertainty determined for samples from unit QPg in 10^3 years, un-adjusted ages (red squares) were determined using Lal (1991) / Stone (2000) scaling for $^{10}$Be production. The effects of modeled 3 mm ka$^{-1}$ erosion snow shielding (blue triangles) are also indicated.

The coherence between $^{10}$Be ages from widely separated parts of this unit and between samples from both boulders and scoured bedrock increases confidence in the determined timing of deglaciation of this area (Table 3.2). Samples from this unit appear to be minimally affected by prior exposure or burial which would influence the $^{10}$Be exposure ages. The small corrections for $^{10}$Be background levels (Table 3.1) enable high precision of the determined exposure ages for these surfaces and lend confidence that there are no meteoric $^{10}$Be contamination issues.

Modeling the effects of snow shielding based on historical data and the effect of an assumed 3 mm ka$^{-1}$ erosion rate increases the exposure ages by 5% and 3.7%,
respectively. The snow- and erosion-adjusted mean age of deglaciation is 14.8 ± 0.7 ka (Figure 3.2, Table 3.2). Using the oldest-age method (Briner et al., 2005; Licciardi and Pierce, 2008), the age of stabilization of unit QPg would be 15.5 ± 0.3 ka. Given the high coherence (within 2σ uncertainty) between all samples, however, the preferred estimate for the age of unit QPg determined by ¹⁰Be exposure dating is 14.8 ± 0.7 ka.

### Table 3.3 Lichenometry data and the corresponding lichen-age

<table>
<thead>
<tr>
<th>Unit</th>
<th>Single largest lichen Diameter (mm)</th>
<th>N</th>
<th>Single largest lichen Age (year AD)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Unit QHg2</td>
<td>51</td>
<td>195</td>
<td>1842</td>
</tr>
<tr>
<td>Unit QHg1</td>
<td>75</td>
<td>30</td>
<td>1692</td>
</tr>
<tr>
<td>Unit QPg</td>
<td>161</td>
<td>101</td>
<td>- 1171 (3.2 ka)</td>
</tr>
</tbody>
</table>

The late Wisconsin age of this surface as determined by ¹⁰Be cosmogenic exposure dating is not consistent with lichen measurements recorded by this and previous studies (Péwé and Reger, 1983; Reger and Péwé, 1991). Lichen measurements recorded from the same features sampled for TCN exposure dating indicate a much younger age for these surfaces. The largest lichen colony measured on stable surfaces associated with this deposit is 161 mm and the mean ± sd of the 5 largest lichens is 158 ± 2.9 mm (Table 3.3). Using the largest lichen present as an indicator of duration of exposure and the logarithmic calibration curve of Solomina and Calkin (2003), the corresponding lichen-age of unit QPg would be 3.2 ka. Using the mean of the 5 largest lichens found, the lichen-age is 2.9 ka. This older part of the growth curve for lichens is very flat and poorly constrained which creates large age uncertainties.
3.1.2 Holocene Glacial Deposit 1 (QHg1)

This discontinuous terminal and lateral moraine complex is composed of a gray till with unweathered mica schist boulders. Relief of moraine ridges in this unit ranges from 4 m for the end moraine crest to 6 m for the lateral moraines plastered on valley walls. This unit is preserved primarily near the moraine crests, as lower elevations of this deposit have been overridden by later glacial advances. The end moraine of this deposit is the oldest moraine crest preserved beyond the modern Castner Glacier. The glacial stillstand that formed this unit implies a coalescence of the Castner, Fels and Canwell Glaciers as indicated by the continuity of deposits. The moraine is mostly stabilized and shows little evidence of mass wasting and slope instability, with boreal forest vegetation containing upright spruce trees with basal diameters < 1.5 m. Most of the areas mapped as QHg1 in this study were previously designated as coeval with the older Holocene advance of the Canwell Glacier (Reger and Péwé, 1991, see Appendix E).

The timing of stabilization for the end and lateral moraines of unit QHg1 was determined by a combination of surface exposure dating analyses and lichenometry measurements collected from boulders on the lateral and terminal moraine crests of this unit (Figure 3.1). Sample MH07-30 was recovered from a quartz vein in a mica schist boulder located on a discontinuous end moraine and sample MH07-29 was recovered from a quartz vein in a mica schist boulder on the south lateral moraine (Figure 3.3). The exposure ages of samples MH07-29 and -30 (no adjustments for erosion or snow shielding, ± 1σ analytical uncertainty) were measured to be 1643 ± 30 years AD and 1778 ± 35 years AD, respectively (Table 3.2). The effect of 3 mm ka⁻¹
Figure 3.3 View looking east of the south lateral moraines of the Castner Glacier. The unit boundaries from Figure 3.1 are indicated along with sample locations and $^{10}$Be exposure ages (adjusted for erosion and snow cover). This sequence shows all of the late Wisconsin through present glacial deposits mapped in this study. The modern Castner Glacier is behind the ridge of QHg2 on the left side of the photo. Photo by M. Howley July, 2008.

Surface erosion is negligible for the ages of these samples. The estimated effect of snow shielding increases the ages of samples MH07-29 and -30 by ~5% to 1626 ± 32 years AD and 1767 ± 37 years AD, respectively. The $^{10}$Be/$^9$Be ratio of the process blank (BeBLK-22) prepared with these samples is 2.11 x 10$^{-15}$ (Table 3.1). The blank corrections for these samples (15% and 28%, respectively) are considerably higher than those determined for older QPg surfaces, likely due to the low $^{10}$Be concentration in the much younger QHg1 surfaces (Table 3.1).
Sample MH07-30 comes from a boulder situated on a discontinuous moraine crest located ~600 m inboard of the maximum mapped extent of unit QHg1. This discontinuous moraine crest is interpreted as a recessional moraine deposited after the maximum QHg1 advance. This spatial relationship is the preferred explanation for the younger age of sample MH07-30 as compared to sample MH07-29 which is close to the outer QHg1 contact (see Figure 3.1). In light of this relationship, the preferred age for the maximum extent of unit QHg1 as determined by cosmogenic exposure dating is the (snow adjusted) age of sample MH07-29 at 1626 ± 32 years AD (1594 to 1658 AD). Abandonment of this moraine position is constrained by the age of sample MH07-30 at 1767 ± 37 years AD (1730 to 1804 AD).

The surface exposure ages from unit QHg1 are supported by lichen measurements from stable surfaces associated with this moraine complex. Of the 30 lichens measured from this unit, the largest lichen present is 75 mm and the mean ± sd of the 5 largest lichens present is 72 ± 2.7 mm. Using the largest lichen method, the lichen-based age of unit QHg1 is 1692 AD, and using the mean of the 5 largest the age is 1717 AD (Table 3.3). This result is between the ages indicated by surface exposure dating. A previous study by Pévé (1987) utilized radiocarbon dating to determine the age of the glacial advance that formed the outer Holocene moraine at the Canwell Glacier. A 15 cm diameter spruce log collected from beneath till of this moraine sampled for $^{14}$C dating yields a calibrated age range of 1437 to 1681 AD (391 ± 122 years BP) (sample GX-10982, 2σ error, Pévé 1987). This gives an estimate of the maximum age of the glacial advance that formed this unit at the Canwell Glacier, which has previously been mapped as coeval to the QHg1 unit at Castner Glacier (Pévé and Reger, 1983). The surface
exposure ages and lichenometry data reported here constrain the timing of stabilization and retreat from the moraine crest. The advance that deposited the QHg1 moraine began after 1559 ± 122 years AD, was at its end moraine by 1626 ± 32 years AD, and was receding by 1767 ± 37 years AD.

3.1.3 Holocene Glacial Deposit 2 (QHg2)

The innermost moraine complex at the Castner Glacier, unit QHg2, is composed of a mostly continuous, largely unweathered light gray till containing mica schist cobbles. These deposits form a partially ice-cored, sharp crested moraine bisected by Castner Creek. The moraine crest shows evidence of active mass wasting and slope instability, and ranges in height from 5 to 20 m above the valley floor. This unit occupies the zone between unit QHg1 and the edge of the modern glacier and contains at least four sharp-crested, low relief (< 4 m high) recessional moraines (Figure 3.1). This moraine represents the most recent advance or stillstand of the Castner Glacier as well as the only moraine complex preserved where no glacier ice coalesced with the nearby Fels and Canwell Glaciers. Vegetation is likely first generation and composed of mostly shrubs with a few spruce trees with basal diameters up to 0.2 m.

The age of this moraine complex was determined by lichen measurements collected in this study, combined with previous tree-ring counting by Péwé (1961). 195 lichen diameters were recorded from stable boulder surfaces found at 16 different field stations on unit QHg2. The largest lichen found on boulder surfaces on this moraine had a diameter of 51 mm, and the mean ± sd of the 5 largest lichens was 49 ± 1.5 mm (Table 3.3). The corresponding age of this surface based on the largest lichen present is 1842
AD. This age is supported by tree-ring counting by Péwé, who in 1951 counted 102 rings from a first-generation spruce tree growing on the terminal moraine crest of this unit (Péwé and Reger, 1983). Accounting for the ecesis time of tree colonization (+15 years, Péwé and Reger, 1983, their Table 4), the age of this feature according to Péwé’s tree-ring counting is 1834 AD. The age of moraine stabilization as determined by a combination of these two methods is between 1834 and 1842 AD.

The advance that deposited the QHg2 moraine complex began after recession of the glacier from the QHgl extent ($^{10}$Be exposure age, 1767 AD). This age gives an estimate of the maximum timing of the advance, and the lichenometry and tree-ring data reported here constrain the timing of deposition and stabilization of the moraine crest. Results indicate that the Castner Glacier deposited its youngest moraine between 1834 and 1842 AD. This is supported by historical observations of W.C. Mendenhall in 1898 who describes the terminus of the Castner Glacier as an extensive ice-cored ablation moraine that extended from active glacier ice to its end moraine (Mendenhall, 1900).

### 3.2 ELA and climate variations in the Delta River Valley

The position of the modern ELA of the Castner Glacier determined by the AAR method is 1750 ± 10 m asl (Figure 3.4a). Using the temperature lapse rate equation developed in this study (Figure 2.4a and b), the equilibrium ablation season temperature (JJA temp) and annual accumulation at the ELA was extrapolated as a function of elevation $x$ using the lapse rate equations: JJA temperature ($^\circ$C) = -6.2143 x (km) +14.848 and annual accumulation (cm H$_2$O) = 88.979 x (km) + 5.2106. The JJA temperature at the modern ELA of the Castner Glacier is calculated to be 4.0 °C. The annual
Figure 3.4 a. The modern (2007 AD) area occupied by the Castner Glacier (59.1 km$^2$) with an accumulation area of 35.5 km$^2$ (blue) and ablation area of 23.6 km$^2$ (red) and the determined ELA at 1750 ± 10 m asl. b. The areal extent of the Castner Glacier at the height of the advance that deposited the QHg2 moraine (total area = 70.2 km$^2$). The accumulation area (blue) occupied the same basin area but has a greater total area of 42.1 km$^2$ due to a lower ELA at 1630 ± 10 m asl. The ablation area (red) is reconstructed from the extent of QHg2 deposits that once occupied the entire valley bottom with an area of 28.1 km$^2$. Base map U. S. Geological Survey digital elevation model. See text for further explanation.
accumulation at this ELA is calculated to be 160.9 cm H$_2$O. These climate parameters at the modern ELA of the Castner Glacier fall within the field of JJA temperature and annual accumulation combinations (Figure 3.5) observed for ELAs of 70 modern mid- and high-latitude glaciers (Ohmura et al., 1992).

The paleo-ELA of the Castner Glacier during the height of the last QHg2 advance was determined to be 1630 ± 10 m asl based on the MELM method. This value agrees with the AAR-based ELA elevation during this period (Figure 3.4b). This implies an ELA rise of 120 ± 20 m since height of the QHg2 advance (1834 to 1842 AD). If the
raising of the ELA by 120 m was forced entirely by an increase in temperature, the magnitude of the increase would be at least 0.68 to 0.81 °C. If the change in ELA position were due entirely to a decrease in annual accumulation, the corresponding annual decrease would be 8.9 to 12.5 cm H$_2$O. The magnitude of these modeled climate changes assume that the modern glacier is at equilibrium, which is unlikely (see section 4.5).
CHAPTER IV

DISCUSSION

4.1 Deglaciation of the Delta River Valley

The terminal moraine of the Donnelly Glaciation (Pévé, 1953), located 42 km north of Castner Valley, is the maximum extent of the ice stream that occupied the Delta River Valley during MIS 2. The age of stabilization for this moraine has previously been reported at 21 to 20 ka (Porter et al., 1983; Hamilton, 1994; Kaufman and Manley, 2004; Briner and Kaufman, 2008). The cosmogenic $^{10}$Be surface exposure ages obtained in this study for unit QPg provide the first available constraints on the timing of ice recession from the Donnelly Glaciation terminal position to the Castner Glacier region following the MIS 2 maximum. Ice retreat from the Donnelly Moraine likely began well before deposition of unit QPg in the Castner Valley. The $^{10}$Be ages determined on surfaces from this unit indicate that ice abandonment occurred at 14.8 ± 0.7 ka from the upper Delta River Valley (Table 3.2). The rate of ice recession over the 42 km distance between the Donnelly Moraine and the Castner Valley is approximately 2.8 m yr$^{-1}$.

Lichen measurements collected from the same surfaces sampled for cosmogenic $^{10}$Be exposure dating indicate a much younger age of 3.2 ka for unit QPg (Tables 3.2 and 3.3). The cosmogenic $^{10}$Be ages reported in this study indicate a minimum exposure duration for these surfaces that would unlikely be explained as any younger by any post-
depositional processes. The > 10 ka discrepancy between lichen-based and \(^{10}\text{Be}\) ages therefore indicates that the lichen age is not an accurate measure of deposition age. This discrepancy suggests major limitations in the lichen-age determination method, at least for this moraine. The lichens measured on these surfaces range in diameter from 130 mm to 161 mm, approaching the largest diameter observed in this region for a single colony (170 mm at Gulkana Glacier, Begét, 1994). The calibration curve delineates an extremely slow growth rate of 0.016 mm yr\(^{-1}\) for lichens > 150 mm diameter (Figure 2.3). Ages determined using lichen measurements corresponding to the flatter regions of the logarithmic calibration curve are therefore highly uncertain and must be tested using independent age determinations. An alternative explanation for the anomalously young lichen ages reported here is that the measured lichens may not represent first-generation growth on this surface, and therefore yield an erroneous surface age, or that lichen growth is being inhibited by crowding out from adjacent lichens.

The highest resolution late glacial climate reconstructions that exist for the central Alaska Range are derived from lake level fluctuations and pollen records at Birch Lake (Abbott et al., 2000) and Windmill Lake (Bigelow and Edwards, 2001). Both of these studies identified cooler and drier conditions prior to 15 ka (12.7 \(^{14}\text{C}\) ka) shown by seasonally dry or greatly decreased lake levels. Pollen assemblages at both sites from this time indicate sparse herb and grass-dominated tundra vegetation, also suggestive of a cold and arid environment. Between 15 ka and 13.7 ka (12.8 and 11.8 \(^{14}\text{C}\) ka) lake levels increased at Birch Lake by 18 m and at Windmill Lake by 4 m, signifying an increase in effective moisture (precipitation minus evaporation) in the region. The dominant vegetation, as indicated by pollen abundances, shifted from sparse herb and grass-
dominated tundra to birch-dominated tundra. This transition is recognized in other pollen studies throughout the region and likely is the result of a warmer and wetter growing season (Abbott et al., 2000; Bigelow and Edwards, 2001). The timing of the climate shift from cool, arid conditions to warmer and wetter conditions between 15 and 13.7 ka coincides with the age of deglaciation reported here of 14.8 ± 0.7 ka.

The timing of deglaciation of the upper Delta River Valley also corresponds well with Greenland ice core records of the onset of the North Atlantic Bolling warm period between 15.0 and 14.5 ka in the North Atlantic region (Figure 4.1) (Grootes et al., 1993; Severinghaus and Brook, 1999; Alley et al., 2003; Rasmussen et al., 2006). The Bolling transition is postulated to have occurred over a period of as few as 50 years centered on 14.7 ka, as indicated by high resolution sampling in the NGRIP ice core (Steffensen et al., 2008). In addition, evidence of rapid sea-surface temperature (SST) warming in subarctic and mid-latitude North Pacific regions has been reconstructed in marine sediment cores from the Gulf of Alaska (de Vernal and Pedersen, 1997) and the Santa Barbara Basin (Hendy and Kennett, 2000). These records show abrupt SST warming occurred in these regions between 15.0 and 14.5 ka. Records from many other mid-ocean Pacific sites indicate that gradual SST warming began much earlier (c. 19 ka) than in these subarctic and mid-latitude North Pacific locations (Kiefer and Kienast, 2005).

The coincident timing of the Bolling transition at 14.7 ka with the deglaciation of the upper Delta River Valley may suggest dynamic responses of alpine valley glaciers in the Alaska Range to abrupt climate shifts recorded in both proximal and distal locations. The timing of deglaciation during the Bolling transition following the termination of iceberg rafting event H1 and the concurrent rapid warming of SST in Atlantic and
marginal North Pacific regions may indicate climate teleconnections between the northern Pacific rim and the North Atlantic basin. The existence of major teleconnections between these regions during this time has previously been shown by sudden synchronous shifts recorded in multiple proxy records during both the Bølling warming and the Younger Dryas chronozone (e.g. Stuiver et al., 1993; Briner et al., 2002; Kaufman et al., 2003; Briner et al., 2005; Hu et al., 2006; Kokorowski et al., 2008).
4.2 Younger Dryas

A significant finding of this study is the absence of a moraine of the Castner Glacier dating to the Younger Dryas chronozone (YD) (12.9 to 11.6 ka; Alley, 2000) that was more extensive than the Little Ice Age (LIA) margin. The YD has been hypothesized to be a global climate deterioration in response to an abrupt shift in North Atlantic Ocean circulation patterns (Broecker and Denton, 1989; Alley et al., 1993). Records from widespread Northern Hemisphere locations, including Alaska, indicate climatic fluctuations during the YD chronozone (e.g. Hu et al., 1993, 1995; Bigelow and Edwards, 2001; Bigelow and Powers, 2001; Mann et al, 2002). No intermediate moraines are found between the late glacial unit QPg (14.7 ka) and the mid-Little Ice Age moraine QHg1 (324 years BP) at the Castner Glacier (see Figure 3.3). This indicates that the glacial response to climate forcing associated with the 1,300-year duration YD event in this part of the Alaska Range was apparently less pronounced than the glacier response to the 400 year duration LIA cool period.

Climate proxies developed from lake-core studies elsewhere in the Alaska Range indicate cooler and drier conditions during the YD chronozone (Abbot et al. 2000, Bigelow and Edwards, 2001). The absence of a YD moraine associated with the Castner Glacier is also contrary to the findings of a recent study in southwest Alaska that established the age of the Waskey Moraine in the Ahklun Mountains between 12.4 and 11.0 ka, sometime during or after the YD event (Briner et al., 2002). The McKinley Park IV glaciation in the western Alaska Range (Ten Brink and Waythomas, 1985) is also attributed to the YD event. Numerous surface exposure dating studies have dated moraines attributed to glacier re-advances during the YD in the western U.S. (e.g., Gosse
et al., 1995; Licciardi and Pierce, 2008), the European Alps and the Southern Alps of New Zealand (Ivy-Ochs et al., 1999), and many other locations. A situation analogous to that found in Alaska has been documented in the Sierra Nevada Mountains of California where analysis of lake-core records indicate cooling during the YD chronozone, but no moraines were deposited in the region between LGM moraines and LIA moraines (Clark and Gillespie, 1997; MacDonald et al., 2008).

At present, the Waskey Moraine in southwest Alaska and the McKinley Park IV advance of the western Alaska Range remain the only well-constrained glacial deposits in the state attributed to the YD (Ten Brink and Waythomas, 1985; Briner et al., 2002). The lack of a YD moraine at the Castner Glacier demonstrates spatial variability in the climate forcing and/or glacier response to this event. The reconstructed decrease in effective precipitation during the YD in the central Alaska Range (Abbot et al. 2000, Bigelow and Edwards, 2001) indicates that the glaciers in this region may have been moisture deprived, which could explain the lack of preserved moraines corresponding to the YD.

The lack of a moraine corresponding with the YD at the Castner Glacier does not support the inferred teleconnections between climate changes in the North Atlantic Basin and interior Alaska described in section 4.1. The coincident Bolling warming and ice recession from the upper Delta River Valley support the existence of these teleconnections, but the lack of a YD moraine at the Castner Glacier indicates a spatially complex regional pattern of climatic response to this event. This apparent discrepancy is not fully resolved at this time. A recent evaluation of the climatic patterns of the YD in Beringia (Alaska and Eastern Asia) by Kokorowski et al. (2008) shows ambiguous
support for widespread cooling over land. Sites investigated south of the Alaska Range do indicate cooling during the YD event, but sites north of the Alaska Range show little or no cooling, and possibly warmer and wetter conditions. These results suggest that the mountains of the Alaska Range played a role in preventing the climatic signal of the YD from extending further northward (Kokorowski et al., 2008). Cooler SST during the Younger Dryas in the North Pacific would also result in less precipitation for inland regions, which may have limited the magnitude of the glacial response to this climate event (Peteet et al., 1997).

4.3 Neoglaciation and First Millennium AD

Glacial records from many regions throughout Alaska include moraines associated with Neoglacial and FMA cooling (Calkin et al., 2001; Wiles et al., 2002; Reyes et al., 2006). Again, there is no evidence in the form of moraines at the Castner Glacier for these climate events between the late glacial and the LIA (see Figure 3.3). Reger and Péwé (1991) indicated that the south lateral moraine of the adjacent Canwell Glacier dated to the Neoglacial period (3600 to 3000 years BP), relying exclusively on ages determined from lichen measurements. Given the previously discussed uncertainty associated with lichen ages of this antiquity (section 4.1), the age of this moraine is in question. The south lateral moraine of the Castner Glacier, from which samples MH07-22 and -28 were recovered, was previously mapped by Reger and Péwé (1991) as continuous with this Neoglacial deposit of the Canwell Glacier. The exposure age determined in this study for two boulders at this location (13.8 and 15.0 ka) refutes the Neoglacial age previously assigned to this advance. There is no evidence for a glacial
advance during the FMA climate event at Castner Glacier.

Both the Neoglacial and FMA cool periods are well represented at glaciers located closer to moisture sources in the Gulf of Alaska (Wrangell Mountains and Coastal Ranges) and at the Black Rapids Glacier in the central Alaska Range (Reger et al., 1993), but no corresponding moraine is present at the Castner Glacier. The climatic explanation for the lack of a preserved moraine at the Castner Glacier associated with the Neoglacial and FMA is not presently known. A decrease in effective precipitation in the region, similar to that hypothesized for the YD period, is unlikely given that the Black Rapids Glacier was advancing during the Neoglacial period. Black Rapids Glacier has previously been documented to surge in response to dynamic inputs aside from precipitation and temperature, which may explain this apparent paradox.

4.4 Little Ice Age activity of the Castner Glacier

The end and lateral moraine complexes preserved at the Castner Glacier are attributed to two glacial advances or stillstands that occurred over the last 400 years. The timing of moraine stabilization, established by cosmogenic $^{10}\text{Be}$ exposure ages for the older complex and previously reported tree-ring record from the younger (Péwé, 1959), is reported here at 1626 ± 32 years AD for QHg1 and 1834 AD for QHg2. Ages based on lichen measurements on stable surfaces for these units are 1692 AD and 1842 AD, respectively. These lichen results are consistent (within 2σ) with the unit stabilization ages determined by other methods, which is expected for this well-constrained part of the calibration curve with lichen growth rates of 3.3 to 4.4 mm yr$^{-1}$ (Figure 2.3). The timing of moraine stabilization reported for these two complexes is well within the defined LIA
Figure 4.2 a. Time-distance diagram showing the timing and relative position of LIA moraines deposited at the Castner Glacier. The ages determined for stabilization of units QHgl at 1626 ± 32 years AD, and QHg2 at 1834 to 1842 AD are shown. b. The glacier expansion index compiled from well-dated glacier records in the Brooks Range, Wrangell Mountains and Coastal Ranges (Wiles et al., 2004). c. Northern Hemisphere temperature anomaly (compared to 1961-1990 mean) reconstructed from tree-ring widths, lacustrine cores and marine cores at many locations with a 25 year smoothing applied (Esper et al., 2002). Note that the scale is reversed so that cooler temperatures are to the right. d. $^{14}$C productivity as a proxy for variations in solar irradiance during the last 1000 years (Bond et al., 2001). Named intervals of decreased solar irradiance (indicated by gray bars) correspond to higher production rates of $^{14}$C due to magnetic field weakening.

cold event in Alaska (1500 to 1900 AD) (Denton and Karlén, 1973a; Péwé, 1975; Calkin, 1988; Wiles et al, 2008). The Castner Glacier LIA maximum positions established in this study (Figure 4.2a) correlate well with the reconstructed mean Northern Hemisphere temperature record of Esper et al. (2002). This reconstruction is based on a compilation of tree-ring records, and is supported by lake and marine sediment core data compiled from a multitude of sites distributed throughout the Northern Hemisphere. This record is biased towards summer temperature which exerts a strong influence on tree-ring width
variations (Moberg et al., 2005) as well as glacial advances and recessions (Patterson, 1994). The coldest time during the LIA indicated by this reconstruction is just prior to 1600 AD with another brief cold period indicated at ~1820 AD (Figure 4.2 c) (Esper et al., 2002). The timing of these coolings coincides with the stabilization dates of QHg1 and QHg2 (within 2σ) for both events.

The two separate LIA advances of the Castner Glacier during the LIA period are analogous to multiple LIA moraines documented in other regions of Alaska (Péwé, 1975; Calkin, 1988; Wiles et al., 2008). The timing of moraine stabilization during the LIA established here for the Castner Glacier does not appear to match well with the individual records of LIA advances developed for glaciers in the Brooks Range, Wrangell Mountains or Coastal Ranges (Calkin, 1988; Wiles et al., 2002; Calkin et al., 2001). The same is true for the composite glacier expansion index (GEI) for all major mountain ranges in Alaska (Figure 4.2 b) (Wiles et al., 2004). Periods of significant glacier expansions in Alaska during the LIA as indicated by the GEI have been attributed to episodes of solar variability (Hu et al., 2001; Wiles et al., 2004) (Figure 4.2 d). The LIA chronology determined for the Castner Glacier appears to be shifted 50 years ahead of both the composite GEI and periods of solar minima indicated by increases in \(^{14}\)C productivity (Bond et al., 2001; Wiles et al., 2004) (Figure 4.2, b and d). This suggests that factors other than solar variability were controlling the timing of fluctuations of the Castner Glacier during the LIA. The timing established here for the advances and recessions of the Castner Glacier during the LIA show a temperature dependence that is not discernible from glacial records reconstructed in other mountain ranges of southern Alaska. The correlation between the timing of moraine deposition at the Castner Glacier
and Northern Hemisphere temperature fluctuations may indicate that this glacier is being influenced by different controlling factors than glaciers in other regions (Brooks, Wrangell and Coastal Ranges). However, the relatively few ages developed in this study for these moraines do not provide sufficient age control to establish phase relationships. The correlation with the Northern Hemisphere temperature reconstruction of Moberg et al. (2005) possibly indicates that the Castner Glacier is controlled more by variations in JJA temperature rather than variations in solar activity.

4.5 Reconstructed post-LIA climate changes

The late Holocene ELA rise of 120 ± 20 m at the Castner Glacier is most likely in response to warming climate conditions since the late LIA (~1840 AD). The amount of ELA rise between ~1840 and 2007 AD for the Castner Glacier is similar to the 100 to 150 m rise determined for land-terminating glaciers in the southern Kenai Mountains between 1850 AD and 2000 AD (Calkin et al., 2001) and the 100 to 200 m rise inferred for the Brooks Range between 1850 AD and 1988 AD (Calkin, 1988).

The annual ELA of a glacier is equivalent to the steady state ELA (i.e., that determined using the AAR method) only when the glacier is in equilibrium with prevailing climate (Benn and Lehmluhl, 2000). This situation may have existed at the culmination of the LIA advance that formed the QHg2 moraine as evidenced by the high-relief moraine crests comprising this deposit. The modern Castner Glacier is presently experiencing both vertical and lateral recession that has been ongoing since the late LIA, indicating that the glacier is currently out of equilibrium with modern climate. Accompanying an increase in summer temperature, there is usually an increase in
precipitation due to enhanced evaporation. If the annual accumulation at the ELA has increased rather than decreased since the LIA, the JJA temperature increase reconstructed here would further underestimate this temperature change.

Figure 4.3 JJA temperature observations from 1943 to 2007 for Big Delta, AK and Gulkana Airport, AK indicating mean JJA temperature increases of 1.1 °C and 0.6 °C, respectively. Data from the Alaska Climate Research Center at the University of Alaska, Fairbanks.

Given these considerations, the reconstructed JJA temperature increase of 0.68 to 0.81 °C are likely minimum values for the changes in climate that have occurred at the ELA. These reconstructed minimum changes are supported by data from two long-term meteorological stations north and south of the Alaska Range: Big Delta, Alaska (85 km north) and Gulkana Airport, Alaska (140 km south). The JJA temperature increases recorded between 1943 and 2007 at these weather monitoring stations are 0.6 and 1.1 °C, respectively (Figure 4.3). These empirical observations indicate that the glacier will likely continue to recede until equilibrium is reached with the present and future climate.
CHAPTER V

CONCLUSIONS

Detailed mapping of glacial deposits at the Castner Glacier in the central Alaska Range defines three distinct moraine complexes beyond the current ice margin. Analysis of cosmogenic $^{10}$Be exposure dating samples collected from the two outer moraine units indicate that these moraines were deposited or bedrock surfaces exposed at 14.7 ± 0.7 ka and 1626 ± 32 years AD, respectively. New lichen-age measurements combined with previously published tree-ring data from the innermost moraine indicate an age of 1834 to 1842 AD. The late glacial age determined for the oldest unit constrains the timing of ice recession in this region following the late Wisconsin glacial maximum stillstand (21 to 20 ka). The unit age of 14.7 ka coincides with the onset of warming indicated by regional pollen assemblage shifts and lake-level variations (Bigelow and Edwards, 2001; Abbott et al., 2000) as well as the sudden onset of the Bølling warm period recorded in Greenland ice cores (Alley et al., 2005; Steffansen et al., 2008).

Ages determined for the two younger moraine complexes indicate deposition by two separate glacial advances during the LIA cool period. The timing of moraine stabilization at 1626 ± 32 and 1834 to 1842 AD show broad consistency with multiple LIA advances in other records from Alaska. The LIA chronology of the Castner Glacier correlates well with the Northern Hemisphere temperature trend reconstructed from tree-
ring records, along with lake and marine sediment cores (Moberg et al., 2005). The timing of the coldest interval of the LIA indicated by this reconstruction (1600 AD) is coincident with the age of stabilization of the older LIA advance (1626 AD). The LIA chronology of the Castner Glacier does not correspond well with the timing of glacial advances in the Brooks Range, Wrangell Mountains and Coastal Ranges coinciding with periods of solar minima (Hu et al., 2001; Wiles et al., 2002).

ELA reconstructions for the 2007 AD and late LIA (~1840 AD) extents of the Castner Glacier indicate an ELA rise of 120 ± 20 m since deposition of the youngest moraine complex. If the change in ELA position is attributed entirely to a temperature change (precipitation constant), the minimum JJA temperature increase at the ELA is 0.68 to 0.81 °C. If attributed to a decrease in annual accumulation (JJA temperature constant), the decrease is at least 8.9 to 12.5 cm H2O. These are minimum estimates based on a modern and past steady state ELA for the Castner Glacier, that are unlikely given the observed recession of this glacier in the last century that indicates the Castner Glacier is not at equilibrium with its climate. In addition, if precipitation has increased rather than decreased since the LIA, additional JJA temperature increase above that reconstructed here would be required to raise the ELA to the 2007 AD position.

There is no evidence at the Castner Glacier for a Younger Dryas re-advance. The lack of a moraine associated with this climate event indicates that the magnitude of glacial response to the YD in the central Alaska Range was less than the response to LIA cooling. This paradox is not unique to this location; there are also no moraines found between LGM and LIA maximum positions in the Sierra Nevada Mountains in California (Clark and Gillespie, 1997; MacDonald et al., 2008). Observations from both ranges
suggest that the glacial response to the Younger Dryas in the Pacific Rim region was variable (Licciardi et al., 2004), possibly moisture-limited in the Alaska Range, and of lesser magnitude than the LIA.
LIST OF REFERENCES


Alaska Climate Research Center, Geophysical Institute, University of Alaska Fairbanks, http://climate.gi.alaska.edu/


Dyuygerov, M. B., and Meier, M. F., 2000, Twentieth century climate change: Evidence from small glaciers. PNAS, 97, 1406-1411.


56


Western Regional Climate Center, 2008. Western Regional Climate Center data set of historical climate data for Big Delta, AK, Trims Camp, AK and Summitt, AK.

APPENDICES
APPENDIX A

SAMPLE LOCATIONS

Sample MH07-22

Date Collected: 07-29-2007

Location: 63.39599° N, 145.70778° W

Elevation: 884 m asl

Lithology: Mica schist boulder with 0.1 to 0.2 m thick quartz veins. Glacial polish preserved on exposed quartz veins.

Surface Strike/Dip: 160° / 25° SW

Sample Thickness: 1.8 cm

Horizon Shielding: 018°/12°, 030°/12°, 055°/8°, 095°/17°, 136°/5°, 145°/6°, 170°/7°
(azimuth/elevation) 180°/3°, 205°/3°, 230°/12°, 270°/5°, 280°/7°, 295°/3°, 305°/4°,
320°/2°, 325°/3°, 347°/8°
**Sample MH07-24**

Date Collected: 08-03-2007  
Location: 63.39190° N, 145.74134° W  
Elevation: 818 m asl  
Lithology: Brown-weathered, white and black colored granodiorite bisected by mafic dikes between 0.5 and 3.0 m thick. Surfaces of this exposure are ice-sculpted, but do not preserve glacial polish.  
Surface Strike/Dip: 028° / 18° E  
Sample Thickness: 1.75 cm  
Horizon Shielding: 026°/12°, 056°/6°, 086°/11°, 111°/4°, 151°/8°, 166°/2°, 181°/5°, 211°/13°, 246°/9°, 276°/7°, 296°/7°, 321°/3°, 336°/2°, 351°/5°
**Sample MH07-26**

Date Collected: 08-03-2007

Location: 63.39233° N, 145.74364° W

Elevation: 814 m asl

Lithology: Brown-weathered, white and black colored granodiorite bisected by mafic dikes between 0.5 and 3.0 m thick. Surfaces of this exposure are ice-sculpted, but do not preserve glacial polish.

Surface Strike/Dip: 000° / 0°

Sample Thickness: 2.5 cm

Horizon Shielding: 026°/12°, 056°/6°, 086°/11°, 111°/4°, 151°/8°, 166°/2°, 181°/5°, 211°/13°, 246°/9°, 276°/7°, 296°/7°, 321°/3°, 336°/2°, 351°/5°
Sample MH07-28

Date Collected: 08-03-2007
Location: 63.39586° N, 145.70746° W
Elevation: 888 m asl
Lithology: Mica schist boulder with 0.1 to 0.2 m thick quartz veins. Glacial polish preserved on exposed quartz veins.
Surface Strike/Dip: 164° / 44° SW
Sample Thickness: 1.9 cm
Horizon Shielding: 018°/12°, 030°/12°, 055°/8°, 095°/17°, 136°/5°, 145°/6°, 170°/7°
(azimuth/elevation) 180°/3°, 205°/3°, 230°/12°, 270°/5°, 280°/7°, 295°/3°, 305°/4°,
320°/2°, 325°/3°, 347°/8°
Sample MH07-29

Date Collected: 08-03-2007

Location: 63.39688° N, 145.70871° W

Elevation: 875 m asl

Lithology: Mica schist boulder with 0.1 to 0.2 m thick quartz veins. Glacial polish preserved on most surfaces.

Surface Strike/Dip: 265° / 22° NW

Sample Thickness: 1.75 cm

Horizon Shielding: 018°/12°, 030°/12°, 055°/8°, 095°/17°, 105°/18°, 120°/9°, 150°/10°, 180°/3°, 205°/3°, 230°/12°, 270°/5°, 280°/7°, 295°/3°, 305°/4°, 320°/2°, 325°/3°, 347°/8°
Sample MH07-30

Date Collected: 08-04-2007

Location: 63.40154° N, 145.73396° W

Elevation: 773 m asl

Lithology: Mica schist boulder with 0.1 to 0.2 m thick quartz veins. Glacial polish preserved on most surfaces.

Surface Strike/Dip: 246° / 12° NW

Sample Thickness: 2.15 cm

Horizon Shielding: 038°/10°, 066°/5°, 102°/9°, 136°/3°, 146°/7°, 161°/9°, 211°/9°, 246°/8°, 291°/8°, 311°/2°, 331°/2°, 341°/8°, 351°/9°
APPENDIX B

CLIMATE DATA - CENTRAL ALASKA RANGE

Table B.1
Mean monthly snow depth and density (determined from snow water equivalent) at Trims Camp, Alaska recorded between 1953 and 1979 (Western Regional Climate Center data).

<table>
<thead>
<tr>
<th></th>
<th>$Z_{\text{snow}}$ (cm)</th>
<th>$\rho_{\text{snow}}$ (g cm$^{-3}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>71.12</td>
<td>0.257</td>
</tr>
<tr>
<td>Feb</td>
<td>111.76</td>
<td>0.286</td>
</tr>
<tr>
<td>Mar</td>
<td>139.70</td>
<td>0.305</td>
</tr>
<tr>
<td>Apr</td>
<td>116.84</td>
<td>0.346</td>
</tr>
<tr>
<td>May</td>
<td>27.94</td>
<td>0.327</td>
</tr>
<tr>
<td>Jun</td>
<td>0.00</td>
<td>0.000</td>
</tr>
<tr>
<td>Jul</td>
<td>0.00</td>
<td>0.000</td>
</tr>
<tr>
<td>Aug</td>
<td>0.00</td>
<td>0.000</td>
</tr>
<tr>
<td>Sep</td>
<td>0.00</td>
<td>0.000</td>
</tr>
<tr>
<td>Oct</td>
<td>20.32</td>
<td>0.175</td>
</tr>
<tr>
<td>Nov</td>
<td>35.56</td>
<td>0.214</td>
</tr>
<tr>
<td>Dec</td>
<td>63.50</td>
<td>0.244</td>
</tr>
</tbody>
</table>
Table B.2 Mean summer (JJA) temperature for 3 stations in the Alaska Range

<table>
<thead>
<tr>
<th>Elevation</th>
<th>732m</th>
<th>985m</th>
<th>1480m</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Trims Camp, AK</td>
<td>Summit, AK</td>
<td>Gulkana Glacier</td>
</tr>
<tr>
<td></td>
<td>JJA temp (°C)</td>
<td>JJA temp (°C)</td>
<td>JJA temp (°C)</td>
</tr>
<tr>
<td>Year</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1968</td>
<td>10.941</td>
<td>9.935</td>
<td>6.867</td>
</tr>
<tr>
<td>1970</td>
<td>8.689</td>
<td>8.069</td>
<td>3.867</td>
</tr>
<tr>
<td>1972</td>
<td>11.002</td>
<td>9.876</td>
<td>5.100</td>
</tr>
<tr>
<td>1973</td>
<td>10.078</td>
<td></td>
<td>3.200</td>
</tr>
<tr>
<td>1974</td>
<td>10.585</td>
<td></td>
<td>6.000</td>
</tr>
<tr>
<td>1975</td>
<td>10.256</td>
<td></td>
<td>5.200</td>
</tr>
<tr>
<td>1976</td>
<td>11.941</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1977</td>
<td>11.735</td>
<td></td>
<td>6.467</td>
</tr>
<tr>
<td>1978</td>
<td>10.322</td>
<td></td>
<td>4.967</td>
</tr>
</tbody>
</table>

Table B.3 Total annual accumulation at 2 stations in the Alaska Range

<table>
<thead>
<tr>
<th>Year</th>
<th>Trims Camp Total Annual Accumulation (cm H₂O)</th>
<th>Gulkana* Total Annual Accumulation (cm H₂O)</th>
<th>Adjusted Gulkana Total Annual Accumulation (cm H₂O)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1969</td>
<td>47.7</td>
<td>58.7</td>
<td>81.0</td>
</tr>
<tr>
<td>1970</td>
<td>92.9</td>
<td>124.5</td>
<td>171.8</td>
</tr>
<tr>
<td>1971</td>
<td>114.1</td>
<td>136.4</td>
<td>188.2</td>
</tr>
<tr>
<td>1972</td>
<td>119.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1973</td>
<td>64.0</td>
<td>86.5</td>
<td>119.4</td>
</tr>
<tr>
<td>1974</td>
<td>75.3</td>
<td>91.9</td>
<td>126.8</td>
</tr>
<tr>
<td>1975</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1976</td>
<td>55.2</td>
<td>84.8</td>
<td>117.0</td>
</tr>
<tr>
<td>1977</td>
<td></td>
<td>91.0</td>
<td></td>
</tr>
<tr>
<td>1978</td>
<td>43.1</td>
<td>111.5</td>
<td>153.9</td>
</tr>
</tbody>
</table>

Mean Annual Accumulation (cm H₂O)

<table>
<thead>
<tr>
<th></th>
<th>Trims Camp</th>
<th>Gulkana*</th>
<th>Adjusted Gulkana</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>70.3</td>
<td>99.2</td>
<td>136.9</td>
</tr>
</tbody>
</table>

* The precipitation catch basin at Gulkana Glacier meteorological station records 62% of annual accumulation (snow + liquid precipitation). The reported totals were increased by 38% to compensate for this (Kennedy et al., 1997).
APPENDIX C

10BE COSMOGENIC EXPOSURE DATING CALCULATIONS

Cosmogenic nuclides are formed by the interaction of 'target' atoms with cosmic radiation. Such nuclides are formed in space, in the atmosphere (e.g. 14C and 10Be), and in situ within minerals at or near the Earth’s surface (e.g. 10Be, 26Al, and 21Ne). The accumulation of cosmogenic nuclides in minerals at or near the Earth’s surface provides a basis for exposure dating of landforms, the quantification of erosion rates, and other geologic applications (Bierman, 1994; Gosse and Phillips, 2001). Evidence strongly suggests that production rates of these nuclides have remained nearly constant, enabling their use in geochronometry. Cosmogenic exposure dating using 10Be has been applied primarily to landforms containing quartz-rich lithologies. 10Be is a radionuclide with a half-life of 1.5 Ma, and is primarily produced by spallation from O, Mg, Si, and Fe (Gosse and Phillips, 2001).

The blank-corrected 10Be/9Be ratios reported by LLNL were converted to atoms of 10Be per grams of quartz using the equation:

\[ N_{10} = \frac{R_{10/9} \cdot M_c \cdot 6.022 \times 10^{23}}{9.012 \cdot M_q} \]

where \( N_{10} \) is the number of atoms 10Be per grams of quartz, \( M_q \) is the mass of quartz (g), \( R_{10/9} \) is the blank corrected 10Be/9Be ratio, and \( M_c \) is the mass of Be added as carrier (g) (Balco et al., 2008).
Table C.1
Blank-corrected $^{10}$Be concentrations from AMS analyses.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Quartz (g)</th>
<th>Carrier Used</th>
<th>Carrier (mg)</th>
<th>$^{10}$Be/$^9$Be (AMS ratio)</th>
<th>1 $\sigma$</th>
</tr>
</thead>
<tbody>
<tr>
<td>MH07-22</td>
<td>40.4495</td>
<td>LDEO BeSO$_4$</td>
<td>0.2192</td>
<td>4.241E-13</td>
<td>2.3%</td>
</tr>
<tr>
<td>MH07-24</td>
<td>40.1936</td>
<td>LDEO BeSO$_4$</td>
<td>0.2129</td>
<td>4.578E-13</td>
<td>2.0%</td>
</tr>
<tr>
<td>MH07-26</td>
<td>39.3996</td>
<td>LDEO BeSO$_4$</td>
<td>0.2171</td>
<td>4.184E-13</td>
<td>2.4%</td>
</tr>
<tr>
<td>MH07-28</td>
<td>40.2366</td>
<td>LDEO BeSO$_4$</td>
<td>0.2212</td>
<td>4.264E-13</td>
<td>2.4%</td>
</tr>
<tr>
<td>MH07-29</td>
<td>42.4628</td>
<td>LDEO Beryl</td>
<td>0.1981</td>
<td>1.401E-14</td>
<td>8.3%</td>
</tr>
<tr>
<td>MH07-30</td>
<td>42.1694</td>
<td>LDEO Beryl</td>
<td>0.2107</td>
<td>7.542E-15</td>
<td>15.1%</td>
</tr>
<tr>
<td>BeBLK-21</td>
<td>0.0000</td>
<td>LDEO BeSO$_4$</td>
<td>0.2210</td>
<td>9.766E-15</td>
<td>16.5%</td>
</tr>
<tr>
<td>BeBLK-22</td>
<td>0.0000</td>
<td>LDEO Beryl</td>
<td>0.2101</td>
<td>2.111E-15</td>
<td>24.4%</td>
</tr>
</tbody>
</table>

Table C.2
Sample and mean unit ages ± 1$\sigma$ (ka BP) using the scaling schemes indicated. No adjustments for erosion or snow shielding.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Unit QPg</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MH07-22</td>
<td>12.79 ± 1.15</td>
<td>13.69 ± 1.61</td>
<td>13.59 ± 1.59</td>
<td>13.08 ± 1.30</td>
<td>13.10 ± 1.16</td>
</tr>
<tr>
<td>MH07-24</td>
<td>14.20 ± 1.27</td>
<td>15.21 ± 1.78</td>
<td>15.09 ± 1.76</td>
<td>14.52 ± 1.44</td>
<td>14.54 ± 1.27</td>
</tr>
<tr>
<td>MH07-26</td>
<td>13.57 ± 1.23</td>
<td>14.54 ± 1.71</td>
<td>14.43 ± 1.69</td>
<td>13.89 ± 1.39</td>
<td>13.90 ± 1.23</td>
</tr>
<tr>
<td>Mean Age</td>
<td>13.60 ± 0.60</td>
<td>14.57 ± 0.64</td>
<td>14.46 ± 0.64</td>
<td>13.91 ± 0.61</td>
<td>13.93 ± 0.61</td>
</tr>
<tr>
<td>Unit QHg1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MH07-29</td>
<td>0.307 ± 0.044</td>
<td>0.333 ± 0.055</td>
<td>0.328 ± 0.054</td>
<td>0.309 ± 0.046</td>
<td>0.316 ± 0.044</td>
</tr>
<tr>
<td>MH07-30</td>
<td>0.172 ± 0.040</td>
<td>0.188 ± 0.047</td>
<td>0.185 ± 0.046</td>
<td>0.173 ± 0.041</td>
<td>0.157 ± 0.041</td>
</tr>
<tr>
<td>Mean Age</td>
<td>0.240 ± 0.095</td>
<td>0.263 ± 0.099</td>
<td>0.258 ± 0.106</td>
<td>0.243 ± 0.099</td>
<td>0.243 ± 0.099</td>
</tr>
</tbody>
</table>
Topographic Shielding

Shielding of sample MH07-22
Strike / Dip of surface: 160°/25°W
Shielding factor = 0.9868

Shielding of sample MH07-24
Strike / Dip of surface: 028°/18°E
Shielding factor = 0.9943

Shielding of sample MH07-26
Strike / Dip of surface: 0°/0°
Shielding factor = 0.9983

Shielding of sample MH07-28
Strike / Dip of surface: 164°/44°W
Shielding factor = 0.9254

Shielding of sample MH07-29
Strike / Dip of surface: 265°/22°N
Shielding factor = 0.9892

Shielding of sample MH07-30
Strike / Dip of surface: 246°/12°N
Shielding factor = 0.9979

Figure C.1 Horizon plots produced by the CRONUS Earth topographic shielding calculator (Balco et al., 2008). Horizon plot of sample MH07-26 is entirely due to angle-to-horizon because the surface sampled is horizontal. The horizon of sample MH07-28 is greatly influenced by the strike / dip of the surface sampled, in this case the s/d is 164°/44° W.
Snow Shielding

The snow shielding $S_{\text{snow}}$ was calculated based on the mean monthly snow depth and density (from SWE) according to the equation:

\[
S_{\text{snow}} = \frac{1}{12} \sum_{i} e^{-(Z_{\text{snow},i} \cdot \rho_{\text{snow},i})/\Lambda_{\text{fe}}}
\]

Equation C.2:

where $Z_{\text{snow},i}$ is the monthly average snow depth (cm), $\rho_{\text{snow},i}$ is the average monthly snow density, $\Lambda_{\text{fe}}$ is the attenuation length (160 g/cm$^2$), and $i$ is the month of the year (Gosse and Phillips, 2001). The shielding coefficient due to snow cover calculated from this data is 0.9198 which results in an age increase of 8.02%.

<table>
<thead>
<tr>
<th>$i$</th>
<th>$Z_{\text{snow}}$ (cm)</th>
<th>$\rho_{\text{snow}}$ (g cm$^{-3}$)</th>
<th>$\Lambda_{\text{fe}}$ (g cm$^{-2}$)</th>
<th>Shielding Effect $S_{\text{snow}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>71.12</td>
<td>0.257</td>
<td>160</td>
<td>0.8920</td>
</tr>
<tr>
<td>Feb</td>
<td>111.76</td>
<td>0.286</td>
<td>160</td>
<td>0.8189</td>
</tr>
<tr>
<td>Mar</td>
<td>139.70</td>
<td>0.305</td>
<td>160</td>
<td>0.7662</td>
</tr>
<tr>
<td>Apr</td>
<td>116.84</td>
<td>0.346</td>
<td>160</td>
<td>0.7767</td>
</tr>
<tr>
<td>May</td>
<td>27.94</td>
<td>0.327</td>
<td>160</td>
<td>0.9445</td>
</tr>
<tr>
<td>Jun</td>
<td>0.00</td>
<td>0.000</td>
<td>160</td>
<td>1.0000</td>
</tr>
<tr>
<td>Jul</td>
<td>0.00</td>
<td>0.000</td>
<td>160</td>
<td>1.0000</td>
</tr>
<tr>
<td>Aug</td>
<td>0.00</td>
<td>0.000</td>
<td>160</td>
<td>1.0000</td>
</tr>
<tr>
<td>Sep</td>
<td>0.00</td>
<td>0.000</td>
<td>160</td>
<td>1.0000</td>
</tr>
<tr>
<td>Oct</td>
<td>20.32</td>
<td>0.175</td>
<td>160</td>
<td>0.9780</td>
</tr>
<tr>
<td>Nov</td>
<td>35.56</td>
<td>0.214</td>
<td>160</td>
<td>0.9536</td>
</tr>
<tr>
<td>Dec</td>
<td>63.50</td>
<td>0.244</td>
<td>160</td>
<td>0.9077</td>
</tr>
</tbody>
</table>

$S_{\text{snow}} = 0.9198$
APPENDIX D

KEY TO MAP UNITS

The map produced in this study is the result of field mapping completed during the summers of 2007 and 2008. Unit boundaries of the four glacial units found at the Castner Glacier are indicated by solid lines. The inferred past extent of the advance that deposited unit QHg1 is indicated by a dashed line. From youngest to oldest the surficial units present are:

Modern Castner Glacial Ice:

The terminal zone of the modern Castner Glacier consists of near-stagnant ice covered by an extensive ablation moraine from 0.5 to 1 m thick. Two drainage streams emanate from beneath the ice and surmaglacial ponds form in ice-collapse structures. The modern glacier surface at the ice-margin is 50 to 75 m lower in elevation than the trimlines and lateral moraines of the last major advance (QHg2) indicating ongoing downwasting and lateral recession.

Qal - Alluvial Deposits:

Composed of the continuous modern braid plains and terraces of the Delta River, Castner Creek and Lower Miller Creek. Rounded to well-rounded pebble to boulder sized clasts that are stratified and deposited by water. These surfaces are seasonally inundated by meltwater from snow and glacier melt water and subject to constant modification by migration of channels.

Qt - Terrace Deposits:

Discontinuous, flat-topped pods of well-rounded gravel and sand with large boulders (up to 3 m³) present. These deposits are associated with paleo-braid plains and outwash surfaces of both Castner Creek and Delta River. Older (higher elevation) deposits are vegetated by small basal diameter (< 0.20 m) spruce trees. The ages of different terrace levels are not distinguished on the map.
QHg2 - Holocene Glacial Deposit 2:

Mostly continuous, unweathered, light gray till and unweathered erratics forming a partially ice-cored sharp-crested moraine that is bisected by Castner Creek. The moraine crest ranges in height from 5 to 30 m above the valley floor and shows evidence of ongoing mass wasting and slope instability. This unit is considered continuous from the boundary with QHg1 to the edge of the modern glacier and contains at least four sharp-crested low relief (< 4 m high) recessional moraines. This moraine represents both the most recent advance of the Castner Glacier, and the only advance preserved indicating that no glacier ice coalesced with the nearby Fells and Canwell Glaciers. Vegetation is first-generation and composed of spruce with basal diameters up to 0.4 m.

QHg1 - Holocene Glacial Deposit 1:

Partially continuous, tan-weathered, gray till with dominant unweathered erratics forming a broadly crested moraine that has been bisected by Castner and Lower Miller Creeks. Relief of this unit ranges from 3 m (terminal moraine) to 100 m above the valley floor (along lateral moraines). This unit is primarily preserved only at or near the moraine crest because most of these deposits have been overrun by subsequent glacial advances. The terminal moraine of this deposit is the oldest continuous unit preserved in the terminal zone of the Castner Glacier. The glacial advance that formed this unit represents coalescence of the Castner, Fels and Canwell Glaciers. The stabilized moraine shows no evidence of mass wasting and slope instability with well developed boreal forest type vegetation containing spruce trees with basal diameters > 1 m. Portions of this moraine system have previously been mapped as coeval with the older Holocene advance of the Canwell Glacier (Reger and Péwé, 1991).

QPg - Late-Pleistocene Glacial Deposits:

Discontinuous, brown-weathered, gray till with dominant weathered schistose erratics forming a ground moraine and lateral moraines. This unit is best preserved above elevations of 775 m asl in areas that were not modified by later glacial advances. Below this elevation, deposits have been affected through fluvial modification by the Delta River, Castner and Lower Miller Creeks. The best preserved examples of these deposits are found on the medial moraine preserved at the former confluence of the Castner and the Fels Glaciers. This locale has multiple large (> 1 m³) erratics overlying >2 m of till with well-developed tundra-type vegetation.
PREVIOUS SURFICIAL MAP OF CASTNER GLACIER

Figure E.1 Surficial map of the Castner and Canwell Glaciers from Reger and Péwé, 1991.