LATE QUATERNARY GLACIER FLUCTUATIONS, ITSI RANGE, YUKON TERRITORY

by

Caleb Andrew Mathias

B.Sc., University of Tennessee, 2019

THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF MASTER OF SCIENCE IN NATURAL RESOURCES AND ENVIRONMENTAL STUDIES

UNIVERSITY OF NORTHERN BRITISH COLUMBIA

April 2025

© Caleb Andrew Mathias, 2025

Abstract

Understanding past glacier fluctuations is critical for understanding Holocene paleoclimate, yet glacier chronologies in the northeastern Canadian Cordillera remain sparse and are frequently obscured by the occurrence of geologic scatter. This thesis reconstructs late Pleistocene and Holocene glacier changes in the Itsi Range, Yukon Territory, and examines processes driving regionally consistent scatter in moraine ages. Using terrestrial *in situ* ¹⁰Be surface exposure dating, I identify two Little Ice Age advances at the Itsi Glacier among the Itsi Range: 0.96 ka \pm 0.36 (outer moraine) and 0.20 ka \pm 0.05 (inner moraine). Erratics dated to 11.61 ka \pm 0.54 indicate Cordilleran Ice Sheet retreat concurrent with the Younger Dryas termination. To contextualize the Itsi Glacier chronology with other regional advances, I compiled ¹⁰Be dated moraine records across western North America moraines organized by moraine morphostratigraphic position as denoted by end groups 1 - 3. The Itsi Glacier inner moraine aligns closely with end group 1 (0.13 ka \pm 0.10), while outer moraine ages reflect significant scatter, mirroring the poorly constrained end group 2 record (0.46 ka \pm 0.35). Distinctions among end group 1 and 2 moraine ages across latitudes suggest that northeastern Canadian Cordillera glaciers emplaced their moraines centuries earlier than those in the southwest. I assess the western North America moraine record with relevant multicentennialscale climate proxy records to deduce what factors influence glacier growth and retreat. Late Holocene advances were driven by orbital forcing, with additional contributions from temperature, precipitation, greenhouse gas emissions, and solar irradiance anomalies coherent with end group 1 culminations. End group 2 moraine scatter complicates its climatic interpretation. A parsimonious explanation for the inconsistencies among end group 2 moraine ages point to the result of amalgamated moraines, heterogeneous mixtures of previously exposed

ii

and reincorporated moraine boulders, deposited between 3.54 - 2.50 ka during the Tiedemann-Peyto Advance. Further research must be carried out to quantify geomorphic uncertainties inherent to techniques of cosmogenic nuclide dating.

Table of Contents

Abstract ii
Table of Contents iv
List of Tables
List of Figures
Acknowledgements xi
1. Introduction
1.1 Motivations and objectives
1.2 Document structure
2. Holocene activity among the Itsi Range
2.1 Climate and the cryosphere
2.1.1 Spatial and temporal climate scales
2.1.2 Glacier chronology as climate proxy
2.1.3 Modern cryospheric conditions7
2.1.4 Canadian Cordillera cryospheric activity
2.2 Terrestrial <i>in situ</i> cosmogenic nuclide dating theory
2.3 Study site
2.4 Methodology17
2.4.1 ¹⁰ Be dating
2.4.2 DEM creation and geodetic mass balance

2.4.3	Termini tracking	. 26
2.5 R	lesults	. 28
2.5.1	Itsi Glacier ¹⁰ Be chronology	. 28
2.5.2	Termini tracking	. 33
2.5.3	Geodetic mass balance	. 33
2.6 D	Discussion	. 34
2.6.1	Geologic scatter	. 34
2.6.2	Inner moraine	. 39
2.6.3	Outer moraine	. 40
2.6.4	Erratics	. 41
2.6.5	Geodetic mass balance and termini retreat	. 42
2.7 C	Conclusion	. 46
3. Comp	arison with western North America moraine records	. 47
3.1 In	ntroduction	. 47
3.2 N	1ethodology	. 48
3.3 E	Discussion of moraine records	. 51
3.3.1	Review of western North America Neoglacial advances	. 51
3.3.2	Radiocarbon, dendrochronology, and glaciolacustrine records	. 53
3.3.3	¹⁰ Be moraine ages	. 55
3.3.4	¹⁰ Be erratic ages	. 61

3.4	Limitations	2
3.5	Conclusion	5
4. Con	nparison of moraine record with climate proxy records	7
4.1	Introduction	7
4.2	Methodology	8
4.3	Results and discussion7	3
4.3.	1 Moraine end groups	4
4.3.	2 Anthropogenic warming	6
4.4	Limitations	7
4.5	Conclusion	9
5. Con	nclusions	0
5.1	Summary of advances	0
5.2	Research limitations	2
5.3	Research recommendations	3
Appendi	x	5
Referenc	9 yes	1

List of Tables

Table 2.1: Itsi Glacier field data collected on 13 August 2022.	18
Table 2.2: Aerial photographs used for glacier change analysis.	22
Table 2.3: NMAD errors before and after coregistration for the DEM of each year	24
Table 2.4: Itsi Glacier ¹⁰ Be sample information.	32
Table 2.5: Statistics derived from the GTT tool and area analysis among all glaciers of the Itsi Range.	33
Table 2.6: Mass budget of Itsi Range glaciers (1949 – 2022).	34
Table 2.7: Itsi Glacier moraine and erratic ages with accompanying reduced chi-squared (X2) and Mean Absolute Deviation (MAD) statistics.	36
Table 3.1: Description for each morphostratigraphic position used for WNA comparisons	49
Table 3.2: ¹⁰ Be dated WNA moraines grouped by morphostratigraphic position with associated statistics (Dortch et al., 2010; Hawkins et al., 2023, 2021; Howley, 2008; Menounos et al., 2013a; Pendleton et al., 2017; Young et al., 2009)	d . 57

List of Figures

Figure 2.1: July solar insolation at 65° N (W/m ²) (Berger and Loutre, 1991) and NH 50 year ensemble mean temperature anomalies (Erb et al., 2022) throughout the Holocene
Figure 2.2: (a) Regional view of the study site in relation to adjacent physiographic provinces. (b) Itsi Range study area and glaciers overlayed on Sentinel 2 10 m true colour imagery captured on 11 August 2022
Figure 2.3: Three-dimensional visualization of the 2004 orthomosaic of the Itsi Range, optimized using CLAHE. No vertical exaggeration is applied. Glaciers are outlined in red, outer moraine highlighted in yellow, and inner moraine highlighted in blue. Scale is approximate. Created in QGIS using the Qgis2threejs plugin
Figure 2.4: Itsi Glacier and moraines. Aerial photo and Sentinel-2 derived glacier extents are overlayed for 1949, 2004, and 2022. DEM basemap was generated using stereoscopic aerial image pairs from 2004. Individually ¹⁰ Be dated boulder ages from each moraine are included and reported with propagated error as LSDn exposure age BP
Figure 2.5: ¹⁰ Be ages from two neoglacial moraines emplaced by the Itsi Glacier reported in ka BP. Grey curves show ages of individual boulders rendered as Gaussian distributions. The coloured lines show the sum of each individual Gaussian distribution for the ¹⁰ Be ages sorted by morphostratigraphic position. Box-and-whisker plots corresponding to each moraine denote the median \pm IQR, minimum, and maximum surface exposure ages
Figure 2.6: ¹⁰ Be ages from two erratic boulders outboard the Itsi Glacier moraines reported in ka BP. The pink line shows the sum of each individual Gaussian distribution. Box-and-whisker plot denotes the median \pm IQR, minimum, and maximum surface exposure ages of the two erratics.31
Figure 2.7: A regional overview of permafrost zone coverage in relation to the Itsi Range field site from Obu et al. (2019)
Figure 2.8: An oblique angle of the Itsi Glacier forefield highlighting the inner and outer moraines as well as an elongated ridge of glacially abraded rock proximal to the ice margin and parallel to the moraine axes
Figure 2.9: ERA5-Land values for the Itsi Range (62.9°, -130.2°) plotted to represent boreal summer melt (June, July, August (JJA)) and winter accumulation (December, January, February (DJF)) seasonal means. Values are plotted as five-year moving averages with associated trendlines
Figure 2.10: ERA5-Land 2023 MAT anomalies for the YT using reference period 1950-1980. Itsi Range is highlighted in green
Figure 2.11: Observed mean discharge values at Ross River (61.99, -132.41; Station ID: 09BA001) within the Upper Pelly sub-sub-drainage area for May and June. Extracted from the Environment and Climate Change Canada Historical Hydrometric Data web site (https://wateroffice.ec.gc.ca/mainmenu/historical_data_index_e.html)
Figure 3.1: Regional compilation of glacial advance intervals from 4.0 ka to present based on regional reviews using evidence from radiocarbon dating, dendrochronology, and glaciolacustrine sedimentation records (Barclay et al., 2009; Hawkins et al., 2021; Menounos et

Figure 4.1: Neoglacial climate forcings and indicators. (a) Global volcanic eruptions per decade (30-year moving average) and major VSSI event magnitudes (Tg S) (Sigl et al., 2021; Toohey and Sigl, 2019). (b) NPI at Mt. Logan (750-year spline) (Osterberg et al., 2014). (c) TSI anomalies ($\Delta W m^{-2}$) plotted as 50-year moving averages (light line) and smoothed using a Gaussian filter (dark line) (Steinhilber et al., 2009). (d) A spline smoothed dataset of atmospheric GHGs CO₂ and CH₄ plotted as radiative forcing ($\Delta W m^{-2}$) (Köhler et al., 2017) Increases in

Acknowledgements

I acknowledge that the University of Northern British Columbia resides on the unceded territory of the Lheidli T'enneh, the First Nation where the two rivers flow together. I thank the Yukon Kaska Dena First Nations of the Ross River Dena Council and Liard First Nation for allowing the sampling of boulders on their land. Thank you to Gerald Osborne and Rachel Pagdin for accompanying Brian and me during field sampling.

I extend my deepest gratitude to my supervisor, Brian Menounos, for providing not only the opportunity for this research but also what became the first vertiginous steps toward a new life abroad. You offered invaluable guidance and provided opportunities that enriched my research skills throughout this journey. Thanks to Adam Hawkins for his insights as both a friend and collaborator during the development of my research, and to Charlie Bourque for keeping me accountable through to the end. Lastly, I thank my partner, Parker Ahlstrom, for providing a warm home and daily support, which were essential to completing my work.

This research was made possible by funding from NSERC and the Hakai Institute.

1. Introduction

The principal goal of this thesis is to reconstruct paleoclimate behaviour during the Holocene in the northeastern Canadian Cordillera, specifically among the Selwyn Mountains of the Yukon Territory (YT). In this region, glacier chronologies remain sparse due to the remote nature of the region and a lack of available organic materials for radiocarbon dating.

The timing of mountain glacier advances in the northeastern Canadian Cordillera, which encompasses mountain ranges in the eastern YT and western Northwest Territories (NWT), in response to climate events throughout the Holocene remains poorly understood (Hawkins et al., 2021; Menounos et al., 2009). Refining glacier chronologies in this area will allow for improved understanding of regional Holocene climate drivers and the complex morphological processes acting upon moraines (Goehring et al., 2022; Solomina et al., 2015). A widely utilized method for developing a glacier chronology is surface exposure dating through the measurement of Terrestrial *in situ* Cosmogenic Nuclides (TCNs) (Balco, 2020a, 2011; Gosse and Phillips, 2001; Schaefer et al., 2022; Solomina et al., 2015). TCN Dating (TCND) is a method to determine how long a rock has been exposed to the flux of cosmic rays emitted primarily from within the Milky Way galaxy (Gosse and Phillips, 2001). Certain minerals react to these cosmic rays with the production of rare radioactive isotopes, the concentration of which is proportional to surface exposure duration (Balco, 2011; Darvill, 2013; Gosse and Phillips, 2001)

Few applications of TCND have been carried out resolving Little Ice Age (LIA) Holocene mountain glacier fluctuations in the northeastern Canadian Cordillera (Menounos et al., 2009; Solomina et al., 2015). As glacier-fed catchments in this region face declines in latesummer streamflow and heightened risks to geomorphic stability and water quality, understanding the response of alpine glaciers to both long- and short-term climate variability is crucial for effective environmental resource planning and hazard mitigation (Bonsal et al., 2020; Moore et al., 2009). Water resource management must adapt in communities downstream from alpine glaciers to larger influxes of winter and early spring streamflow and account for long-term implications of water availability (Marshall, 2011; Menounos et al., 2020).

1.1 Motivations and objectives

The primary objective of my research is to contribute to our understanding of past glacier fluctuations in western Canada to help inform broader-scale climatic interpretations (Balco, 2020a; Menounos et al., 2009). To accomplish this, I developed a new glacier chronology for glaciers among the Itsi Range in the Selwyn Mountains of the northeastern Canadian Cordillera. Additionally, I aim to address the methodological uncertainties of geologic scatter that underpin TCND research in this region. My research seeks to answer the following questions:

- 1. How have western Canadian glaciers responded to late Quaternary climate change?
- 2. Is there a plausible cause for observed geologic scatter in this region apart from TCN inheritance or post-depositional disturbance?
- 3. How does the newly derived glacier chronology correlate with existing moraine records, regional advances, and climate proxy data to enhance our understanding of regional Holocene climate drivers?

1.2 Document structure

Chapter 2 presents new numerical ages of cirque glacier moraines in the northeastern Canadian Cordillera using TCND methods and derive geodetic mass balance estimates among glaciers of the Itsi Range since 1949. Chapter 3 compares the new numerical ages and positions of moraines with the western Canada moraine record and discusses observations of geologic scatter. Chapter 4 presents further analysis of the new moraine record with relevant climate proxy records. Chapter 5 recounts the significance of my findings in the context of paleoclimate research, explores remaining knowledge gaps, and the methodological advancements required to improve the accuracy and correlative potential of TCND derived surface exposure ages.

2. Holocene activity among the Itsi Range

Chapter 2 considers how glaciers in the Itsi Range fluctuated during the late Holocene. I developed my glacier chronology by using terrestrial cosmogenic ¹⁰Be surface exposure dating techniques.

2.1 Climate and the cryosphere

2.1.1 Spatial and temporal climate scales

The cryosphere encompasses all frozen forms of water on Earth and is highly sensitive to changes in mean global temperature (Marshall, 2011; Nye, 1960). Climate controls the precipitation and temperature conditions that regulate glacier formation. Glaciers are perennial masses of ice that flow due to gravitational deformation under their own weight (Cogley et al., 2011; Marshall, 2011; Roe and O'Neal, 2009). Glaciers form at localities where snow accumulation equals or exceeds ablation multi-annually, leading to the compression of snow into dense glacier ice. Glacier retreat occurs when ablation exceeds accumulation (Marshall, 2011). Mass balance refers to this equilibrium between accumulation and ablation (Cogley et al., 2011).

Holocene climate drivers operate at several different spatial and temporal scales within western Canada; explaining the behaviour of glacier fluctuations requires considering these scales (Clague et al., 2009; Menounos et al., 2009). Three dominant frequencies of climate change characterize the Holocene: millennial, decadal-to-centennial, and sub-decadal (Clague et al., 2009). The reduction of Northern Hemisphere (NH) summer solar insolation is responsible for millennial-scale periods of successive glacier expansion throughout the Holocene (Clague et al., 2009; Wanner et al., 2011) (Figure 2.1). Changes in solar activity, volcanism, and teleconnection patterns drive decadal-to-centennial glacier fluctuations (Porter, 1986; Wiles et al., 2004). Non-climatic factors such as topography, elevation, and glacier hypsometry affect sub-decadal responses (Clague et al., 2009; Menounos et al., 2013b; Wanner et al., 2011).



Figure 2.1: July solar insolation at 65° N (W/m²) (Berger and Loutre, 1991) and NH 50 year ensemble mean temperature anomalies (Erb et al., 2022) throughout the Holocene.

Additionally, climate broadly operates on three spatial scales: planetary, continental, and regional. On a planetary scale, incident radiation from the sun reaches Earth's surface at lower angles towards the poles, resulting in a cooler climate at higher latitudes from the reduced heat flux and vice versa at lower latitudes (Berger, 1988). At a continental scale, ocean temperatures strongly influence land temperatures due to water's high capacity for storing solar energy (Dommenget, 2009; Menounos et al., 2020). Topography affects climate at a local-to-regional scale; the troposphere cools with increases of elevation, and rain shadows cause desert climates on the leeward side of mountain ranges (Clague et al., 2009).

2.1.2 Glacier chronology as climate proxy

One of the greatest resources for understanding the natural variability of past climate is reconstructing past glacier behavior through the development of glacier chronologies (Balco, 2020a, 2011; Menounos et al., 2009). Mountain glacier mass balance in the NH is most sensitive to variations in boreal summer (June through August) temperatures and, to a lesser extent precipitation, responding with geometric adjustments of thickness, length, and width (Balco, 2020a; Cogley et al., 2011; Hodge et al., 1998; Menounos et al., 2009; Nye, 1960; Solomina et al., 2015). Due to lateral constraints, many valley and cirque glaciers exhibit fluctuations of length as their most common responses to long-term changes in climate (Menounos et al., 2009). Although long records of glacier mass balance offer the most direct way to infer how climate affects glacier health, ample evidence of glacier length change exists and offers the ability to glean information about climatic change (Oerlemans, 1994).

Non-climatic effects such as response time or how glaciers may respond to random, highfrequency weather events complicate insight into past climate (Roe and O'Neal, 2009). Glacier response, in terms of fluctuating geometry, to changes in climate lags at a timescale ranging from several years to centuries (Jóhannesson et al., 1989). This response may indicate either the natural interannual variability of a stationary climate (i.e., one that contains a constant mean and constant standard deviation), which does not reflect a true change in climate forcing, or a statistically significant change in the expected response to natural climate variability (Roe and O'Neal, 2009).

The null hypothesis, that any change in glacier length must be temporally and spatially greater than those expected to be driven by the observed record of interannual variability in a constant climate, must be rejected to demonstrate a glacier chronology as explicitly driven by climate change (Roe and O'Neal, 2009). The signal of anthropogenic warming evidenced by a globally synchronous retreat of glaciers since the late-20th century has been demonstrated as statistically significant, exceeding the natural interannual variability of the climate system; however, even the LIA advances cannot be demonstrated to exceed this natural variability (Reichert et al., 2002; Roe and O'Neal, 2009). Roe and O'Neal (2009) advocate that late-Holocene moraine chronologies merely represent the random noise of a complex climate system and that the inherently stochastic glacier response to climate prevents asserting any significant correlations with paleoclimate drivers (Kirkbride and Winkler, 2012; Roe and O'Neal, 2009). Although this perspective may be provocative and circumspect, the criticism offers a valid rationale for examining quantified glacier response time lags before interpretation of dated terminal and recessional moraines.

Bracketing ages of a given glacier advance remains an important prerequisite to closely link climate drivers with glacier response as geologically evident glacier fluctuations can preserve decadal, centennial, and millennial scale climatic signals indicating regional and, when viewed synoptically, global Holocene climate trends (Hawkins et al., 2021; Kaplan et al., 2016; Menounos et al., 2009; Schaefer et al., 2009; Schimmelpfennig et al., 2022; Solomina et al., 2015).

Glacier response time operates as a function of mean thickness, terminus ablation rate, hypsometry, and mass balance gradient, therefore highly variable among different environments and geomorphological settings (e.g., high-relief coastal and low-relief continental interior glaciers will have respectively fast and slow response times) (Kirkbride and Winkler, 2012). This variability of response time implies that individual glaciers will respond differently, as interpreted by their geomorphic record, to the same climatic history, rendering moraine

6

formation scale-dependent wherein large glaciers filter high-frequency climate signals (i.e., overriding of previously deposited moraines) and small glaciers respond more readily (Kirkbride and Winkler, 2012).

2.1.3 Modern cryospheric conditions

Currently, perennial ice covers 10.8% of the Earth's land surface with an additional 15.4% underlain by permafrost – frozen stores of soil at the Earth's veneer (Marshall, 2011). Net radiation from the sun governs temperatures on Earth; however, because the change in solar radiation on Earth over time is quite small, the greatest contributing factor to planetary climate is global albedo (i.e., the planet's reflectivity) which largely depends upon snow and ice cover (Marshall, 2011). Due to its high albedo, the cryosphere reflects shortwave radiation from the sun at high latitudes and mountainous regions back into space which helps keep Earth's surface temperature stable (Groisman et al., 1994). This albedo reduces incoming net solar radiation which, with sufficient cryospheric cover, contributes to a reduced global mean annual temperature (Groisman et al., 1994; Marshall, 2011). Cryospheric albedo also affects regional-to-local climate due to its influence on atmospheric circulation (Marshall, 2011).

Global temperature variability on the order of 1 degree Celsius has driven Holocene fluctuations of alpine glaciers (Marshall, 2011). Recent warming since the late 19th century is not in line with expected orbital cooling trends (IPCC, 2021; Kaufman and Broadman, 2023). Increased atmospheric Greenhouse Gas (GHG) emissions, as well as solar and volcanic activity, have primarily driven trend divergence (IPCC, 2021; Solomina et al., 2015). Arctic warming increases at over two times the average rate of global surface temperature warming, making this a vulnerable region to climate warming (IPCC, 2021; Marshall, 2011). With unprecedented increases in GHG emissions since the late 20th century, tropospheric warming has destabilized the cryosphere, impacting hemispheric circulation, sea level rise, and ice-albedo feedbacks leading to a warming climate (IPCC, 2021). Additionally, the potential for a dramatic increase of carbon emissions from permafrost stores has risen as permafrost melt leads to a sustained release of GHGs CO₂ and CH₄ due to positive feedback effects from climate warming (IPCC, 2021; Marshall, 2011). Onset of anthropogenic climate change, resulting in the global reduction of glacier mass, diminishes the capacity for glacier and snow melt to support regional energy demands, ecosystems, fisheries, and agriculture of western Canada and increases the likelihood of associated glacier hazards like outburst floods (IPCC, 2022; Milner et al., 2017; Rounce et al., 2023). Global projections reveal that glaciers will experience mass loss ranging from $28 \pm 9\%$ (SSP1-2.6) to $44 \pm 20\%$ (SSP5-8.5) by the year 2100 relative to 2015 (Rounce et al., 2023).

2.1.4 Canadian Cordillera cryospheric activity

The most recent glaciation, the Last Glacial Maximum (LGM), occurred from 23 – 19 ka and covered a significant portion of the NH with ice sheets (IPCC, 2021). The Cordilleran and Laurentide Ice Sheets (CIS/LIS) merged around 22 ka, covering much of modern Canada and the northern United States (Dalton et al., 2023; Darvill et al., 2018). The CIS attained its maximum extent circa 18.5 ka with subsequent detachment and retreat of the two ice sheets beginning at 17.5 ka (Dalton et al., 2023). Following the LGM towards the end of the Pleistocene, global climate fluctuated between glacial and interglacial conditions (Menounos et al., 2017). A release of warm deep-ocean waters in the North Atlantic that had built up over time reignited the Atlantic meridional overturning circulation causing the Bølling-Allerød (BA), an abrupt warming event at 14.6 ka (Thiagarajan et al., 2014). BA warming induced CIS mass thinning which subsequently allowed valley and cirgue moraines in the western Canadian Cordillera to advance into newly deglaciated alpine areas as early as 14.0 ka (Menounos et al., 2017). The BA represents the final interstadial of the last glacial period which terminated at 12.9 ka with the onset of the Younger Dryas (YD), the Pleistocene's final stadial (Broecker et al., 2010). The CIS disappeared by the end of the YD which terminated at 11.7 ka (Menounos et al., 2017; Rasmussen et al., 2006). A return to consistent interglacial conditions after 11.7 ka marked the beginning of the Holocene (Rasmussen et al., 2006). The Holocene Thermal Maximum (HTM) was a period during the early Holocene from approximately 11-5 ka (Renssen et al., 2009; Schimmelpfennig et al., 2022). The HTM signal is most strongly recorded in the NH, characterized by a relatively warm climate due to orbitally forced high summer insolation (Carlson et al., 2008; Renssen et al., 2009). Decreasing insolation in the NH during the boreal summer led to Neoglaciation, characterized as the period of time when NH glaciers advanced from early-to-mid Holocene, namely HTM, minimal extents and encompasses multiple smaller scale phases of advance and retreat (Luckman et al., 1993; Menounos et al., 2009; Porter and Denton, 1967; Wanner et al., 2011). This reduction of NH insolation culminated as globally recorded negative temperature anomalies that characterize the most recent cold relapse, the LIA, occurring between 0.7 – 0.15 ka (1200 – 1800 CE) (Berger, 1978; Wanner et al., 2011).

Glaciers within the Canadian Cordillera expanded throughout the Holocene with maximum downvalley positions of glaciers achieved in the last millennium during the LIA; because of this, scant morphological evidence for previous Neoglacial activity in the Canadian Cordillera remains as prior glacier activity, in the form of lateral and terminal moraines, is commonly overridden and erased through a process of erosion censoring called "obliterative overlap" (Gibbons et al., 1984; Kirkbride and Winkler, 2012; Luckman et al., 1993; Menounos et al., 2009; Reyes et al., 2006). Erosion censoring by means of obliterative overlap effectively erases former records of climate among geomorphological deposits. However, some highelevation pre-LIA cirque and valley moraines in the Yukon Territory (YT) and Northwest Territories (NWT) were discovered hundreds of meters to several kilometers beyond LIA extents, either forming shortly after or interacting with the decaying CIS (Lakeman et al., 2008; Menounos et al., 2017). Climatically, anthropogenic warming has defined the last two centuries, largely due to the introduction of GHGs into the atmosphere and has prematurely terminated influence of Little Ice Age (LIA) climate (IPCC, 2021; Wanner et al., 2011). Compared to baseline variability of centennial to millennial glacier fluctuations recorded in Holocene glacier chronologies, the anomalous retreat of glaciers throughout the last century in alpine regions worldwide provides evidence for anthropogenic climate change (IPCC, 2013; Solomina et al., 2015).

2.2 Terrestrial in situ cosmogenic nuclide dating theory

Most of the Galactic Cosmic Radiation (GCR) that contributes to TCN production originates from the Milky Way galaxy (Gosse and Phillips, 2001). While the GCR cascades from its primary source to the Earth, interactions between primary GCR and nuclei of atmospheric atoms create a series of reactions and particles with net energy eventually lost to the materials of Earth, such as rock (Darvill, 2013; Davis Jr and Schaeffer, 1955). Production of cosmogenic nuclides occurs first as primary cosmic rays, then as secondary cosmic rays, and finally as nuclides produced through nucleonic spallation and muogenic reactions on certain elements (Darvill, 2013; Schaefer et al., 2022). Various spatial and temporal factors such as latitude, altitude, GCR intensity and solar modulation, atmospheric shielding, and the shifting morphology of the landform surface through time determine the production rate of cosmogenic nuclides in rocks exposed at the surface generated by this cosmic radiation flux (Darvill, 2013; Lal, 1991; Schaefer et al., 2022). Terrestrial *in situ* cosmogenic nuclide (TCN) production rates assume a reference condition that consists of a horizontal surface with no obstructions at high latitude and at sea level. In practice, TCN samples collected for research cover a wide range of latitudes, elevations, and surface slopes that may be partially obstructed. To accurately record TCN exposure ages, implementation of a scaling model allows the determination of spallogenic and muonic nuclide production rates for any latitude, altitude, depth, and time period (Lal, 1991; Lal et al., 1960; Lifton et al., 2014).

After the discovery of cosmic radiation by physicist Victor Hess in 1912, geologists promptly recognized the potential utility for solving geological problems with the first attempt at TCN Dating (TCND) carried out in 1955 (Davis Jr and Schaeffer, 1955; Gosse and Phillips, 2001). When Accelerator Mass Spectrometry (AMS) was introduced in the 1980s, measurement of cosmogenic nuclides became broadly accessible, thus applications of TCND grew exponentially in the years to follow (Brown et al., 1982; Gosse and Phillips, 2001).

TCND has distinct advantages over other methods used to construct a chronology of glacier fluctuations prior to the 20th century. Other approaches such as lichenometry, dendrochronology, and radiocarbon dating of subfossil wood, rely on biological proxies. Lichenometry, often criticized as pseudoscience, has been scrutinized for producing unreliable ages and for lacking consistent methodological practices (Osborn et al., 2015). Although radiocarbon dating can be highly valuable, it requires environments that support vegetative growth and can only bracket an age range for glacier advance or retreat (Darvill, 2013; Gosse and Phillips, 2001). The geochemical method of tephrochronology requires the presence of

volcanic tephra deposits within moraine stratigraphy making it less suitable for late Quaternary research (Osborn et al., 2001; Preece et al., 2011).

The relevance of TCND for constraining glacier chronologies is based upon the assumption that fresh rock surfaces are created at the bed of glaciers which are concealed from exposure to the cosmic-ray flux (Balco, 2020a, 2011). Boulders eroded at the glacier bed are carried to the ice margin where they are exposed during ice retreat or deposited as a moraine (Balco, 2020a, 2011). Once emplaced, the rock becomes exposed to the cosmic-ray flux and build-up of previous cosmogenic nuclide concentrations are considered negligible; therefore, the duration of exposure for a relevant rock can be used as a direct proxy for the timing of glacier ice stabilization and retreat (Balco, 2020a, 2011). Accurately dating a glacial landform requires resolving two questions: 1) how long a sample has been exposed to the cosmic-ray flux and; 2) whether or not the sample's exposure duration can be linked to the presence of a glacier at that time (Balco, 2020a).

I focus on applications of ¹⁰Be for dating in this thesis due to its well constrained production rates (within the range of 3 – 4% uncertainty), abundance in the ubiquitous mineral quartz, smallest amount of uncertainty in terms of coefficients of variation, and ability for use at the centennial-scale throughout the Holocene (Balco, 2020a, 2011; Gosse and Phillips, 2001; Schaefer et al., 2022). Technical advances in the 21st century now allow analysis of LIA era ¹⁰Be ages producing low uncertainties and exposure ages consistent with historical records (Braumann et al., 2020; Schaefer et al., 2022, 2009).

Successfully carrying out TCND methods requires knowledge of TCN production at depth below the Earth's surface. Production by spallation, which occurs within the first 3 m of the Earth's surface, produces 96.4% of 10 Be; below 3 m the neutron flux diminishes to <1% and

12

muonic reactions dominate the remaining 3.6% of ¹⁰Be production (Darvill, 2013; Heisinger et al., 2002; Hippe, 2017). Quartz (SiO₂) is the primary target for TCND samples in this research context. High-energy neutron spallation on elements Si and O produce ¹⁰Be isotopes, specifically: beryllium with two additional neutrons (¹⁰Be) from oxygen [¹⁶O(n,4p3n)¹⁰Be] and from silicon [²⁸Si(n,x)¹⁰Be] (Gosse and Phillips, 2001). Exposure time depends on the concentration of ¹⁰Be:

$$N_{10} = \frac{P_{10}}{\lambda_{10}} \left[1 - e^{-\lambda_{10}t} \right] \tag{1}$$

Where N_{10} is the ¹⁰Be concentration in quartz (atoms g⁻¹), P_{10} is the production rate for ¹⁰Be (atoms g⁻¹ yr⁻¹), λ_{10} is the ¹⁰Be decay constant (4.99 x 10⁻⁷ yr⁻¹), and *t* is the exposure time (years) (Balco, 2011).

2.3 Study site

Evolution of the Canadian Cordillera began around 750 Ma with the opening of an ancestral ocean basin due to rifting of the Neoproterozoic supercontinent Rodinia (Monger and Price, 2002). A deep-water depositional environment formed at this time, known as the Selwyn Basin tectonic assemblage, which accumulated clastic sediments from the Late Neoproterozoic to the Triassic (Gordey, 2013; Smith and Bennett, 2004). Beginning with the Jurassic period, western offshore subduction zones converged with the North American continent wherein processes of accretion formed a new continental margin by the Late Cretaceous (Monger and Price, 2002). During the Mid-Cretaceous, a continental arc formed the Selwyn Plutonic Suite that intruded into the sedimentary deposits of the Selwyn Basin forming quartz-rich granitic massifs that host the alpine glacier and genetically related Neoglacial moraine deposits sampled for this research (Gordey, 2013; Smith and Bennett, 2004).

The Selwyn Mountains straddle the YT and NWT border from 61°N to 64°N and belong to the Taiga Cordillera ecozone (Smith and Bennett, 2004). High elevation peaks separated by broad valleys characterize the region. The highest point in this mountain range is Keele Peak at 2970 m, with local relief ranging between 900 – 1500 m. Mid-Cretaceous (98 – 93 Ma) quartz monzonite of the Mayo Suite characterize the dominant bedrock geology of the Itsi Range, matching the lithology of the moraine boulders and erratics chosen for sampling (Colpron, 2022).

Buried thrust faults encompass much of the Selwyn Mountains and the geographically adjacent Mackenzie Mountains have been documented to experience active thrust and strike-slip faulting generating high-magnitude earthquakes (Gordey, 2013; Hyndman et al., 2005). The earthquakes recorded in these northeastern Canadian Cordillera mountain fronts trigger rockslides along arêtes and cirque glacier headwalls (Smith and Bennett, 2004). Within a 100 km radius of the Itsi Range, 165 seismic events have been recorded by the National Earthquake DataBase (NEDB) since 1985; of these earthquakes, four range in magnitude from 4.6 to 5.2. The Selwyn Mountains contain the highest values of precipitation outside of YT's coastal ranges. Among the Pelly Mountains just southwest of the Selwyn Mountains, rock avalanches generating volumes of non-carbonate rock talus on the order of 1 to 5 x 10^6 m³ have been reported (Jackson and Isobe, 1990).

Itsi Mountain (~2500 m a.s.l.) and its neighbouring landforms comprise the Itsi Range, a massif located within the Selwyn Mountains of the eastern YT. The Itsi Range covers an area of approximately 150 km², corresponding to locally exposed contiguous pre-Quaternary bedrock (Jackson et al., 1993). The Itsi Range contains evidence of Quaternary glaciation through the presence of cirque glaciers, arêtes, Neoglacial moraines, and expansive U-shaped valleys.

14

Assessment of August 2022 Sentinel-2 imagery shows 26 active cirque and niche glaciers comprising a summed total area of 11.30 km² (Figure 2.2). Northern slope-facing cirque glaciers have retained a greater volume of ice than their southern-slope facing counterparts. Mean Annual air Temperature (MAT) and Mean Annual Precipitation (MAP) for the Itsi Range are interpolated from a raster grid as -8.5 °C and 519 mm yr⁻¹, respectively (Fick and Hijmans, 2017).

I completed an extensive review of potential study areas within the Selwyn and Mackenzie Mountain ranges of the YT and NWT before selecting a site among the Itsi Range (Figure 2.2). I first examined potential moraines using historic aerial stereoscopic imagery available at the Yukon Energy, Mines, and Resources Library in Whitehorse, YT. This led to the identification of a suitable pair of Holocene moraines and erratics down valley of the moraine pair which sit on the forefield of an active unnamed south-facing cirque glacier, herein referred to as the Itsi Glacier (Figure 2.4). The closely spaced moraines derive from quartz monzonite bedrock and lie several hundred meters down valley from cirque headwalls. ArcticDEM reveals older morphological evidence of the Itsi Glacier's activity, indicating the presence of latest Pleistocene lateral valley moraines. The Itsi Glacier likely constructed these moraines as it advanced soon after, or concurrently with, the decay of the CIS (Menounos et al., 2017).

The Itsi Glacier lies in an area of discontinuous permafrost (50 – 90% coverage; Figure 2.7) which places it in the middle of the basal thermal regime process-form continuum, with the ice-margin likely sensitive to decadal or seasonal climate forcings (Evans, 2009; Obu et al., 2019). The Itsi Massif lies in a setting of low- to moderate-relief, with an approximate vertical distance between valley floor and summit ridge of 1250 m and an average relief of 163 m between Itsi Glacier and its northern headwall.



Figure 2.2: (a) Regional view of the study site in relation to adjacent physiographic provinces. (b) Itsi Range study area and glaciers overlayed on Sentinel 2 10 m true colour imagery captured on 11 August 2022.

2.4 Methodology

2.4.1 ¹⁰Be dating

Given the expense required to access the Itsi Range, funding for TCN analysis was limited to 10 samples; an insufficient amount to attain a synoptic understanding of site-specific paleoclimate dynamics (Balco, 2020a). Consequently, I ascertained that augmenting ongoing research in the eastern YT/western NWT to establish a more robust regional chronology would be more beneficial than simply increasing sample numbers at previously studied sites.

We accessed the field site via a chartered Bell 407 helicopter during late summer (13 August 2022) to minimize the presence of seasonal snow cover. Based on criteria outlined by Darvill (2013) and Gosse and Phillips (2001), we chose ten boulder samples in the field: five boulders from the outer moraine, three boulders from the inner moraine, and two glacial erratics. Relevant qualitative site selection criteria include identifying a quartz-rich lithology, selecting the tallest boulders relative to the morainal ridge and closest to the moraine crest (avoids postdepositional shielding), selecting large boulders indicative of stable emplacement within the moraine matrix (avoids post-depositional exhumation), and ensuring sampled surfaces are sufficiently flat and horizontal showing minimal signs of exhumation and erosion (e.g., presence of glacial striations and a moderate amount of lichen cover present).

Tomkins et al. (2021) make a case that no conclusive pattern to the presence of outliers in a glacier chronology exists when comparing those sampled from three intra-landform locations: moraine crests, ice-proximal slopes, and ice-distal slopes. The most important geomorphic consideration to mitigate geologic scatter remains moraine sedimentology, with boulder-rich

17

matrix-poor moraines more likely to yield reliably accurate boulder ages than matrix-rich unconsolidated moraines (Tomkins et al., 2021). We followed recommendations from prior literature and practice by sampling from moraine crests as Itsi Glacier moraines exhibited a boulder-rich and matrix-poor composition (Gosse and Phillips, 2001; Hawkins et al., 2021).

We used a gas-powered rock saw to recover samples for TCN analysis. We made long, straight incisions along the length and width of the chosen surface, creating a grid of extractable rock 'brownies' with dimensions roughly equal to $5 \ge 3 \le 3 \le (L \ge W \ge H)$ pried out using a hammer and chisel. The process resulted in approximately ~1 kg of extractable material per boulder with samples stored in cloth bags labeled with site identifier, year, and sample number (Figure A.4). I recorded the following properties for each boulder sample: sample ID, strike/dip, boulder dimensions, lithology, amount of vegetation cover, and geomorphic context (Table 2.1). At each boulder's location, a Magellan eXplorist 210 GPS unit recorded latitude, longitude, and elevation. A Brunton transit compass with clinometer recorded the topography of the landscape at 45° intervals to account for topographic shielding of cosmic rays for each boulder. A roughly 15 x 20 cm surface of freshly cut rock remains after extracting each sample. We placed small rocks atop the modified boulder surface to minimize aesthetic disturbance which risks compromising the site's cultural and spiritual values (Schaefer et al., 2022).

Sample	Latitude	Longitude	Elevation	Strike°/	Shielding	L	W	Н
ID	(d. d.)	(d. d.)	(m)	Dip ^{oa}	Factor ^b	(m)	(m)	(m)
Outer moraine								
228B01	62.8904	-130.2817	1722	108/02	0.983	1.45	0.85	0.70
228B02	62.8900	-130.2803	1724	342/08	0.991	1.80	1.60	0.80
228B03	62.8901	-130.2799	1722	214/04	0.991	3.18	2.24	1.60
228B04	62.8927	-130.2905	1750	138/07	0.986	1.50	1.10	0.55
228B05	62.8928	-130.2903	1761	104/04	0.986	1.25	0.75	0.60
Inner moraine								

Table 2.1: Itsi Glacier field data collected on 13 August 2022.

228B06	62.8938	-130.2913	1796	160/12	0.987	2.40	2.20	0.90
228B07	62.8938	-130.2910	1798	290/15	0.986	2.20	1.00	0.70
228B08	62.8936	-130.2916	1789	260/13	0.986	2.40	2.10	1.10
Erratics								
228B09	62.8921	-130.2916	1757	014/05	0.982	1.50	1.45	0.65
228B10	62.8921	-130.2916	1757	326/59	0.821	2.05	1.00	1.00

^a Strike and dip of sampled surface (right-hand rule); values recorded with magnetic declination set to 0 degrees. World Magnetic Model (WMM) magnetic declination for the site is $19.73^{\circ} E \pm 0.55^{\circ}$. ^b Shielding factor calculated using the topographic shielding calculator found at https://stoneage.iced.org/math/skyline/skyline in.html (version info.: wrapper: 2.0; skyline: 2.0).

Sample 228B10, an erratic boulder immediately adjacent to 228B09 and displaying the same highly weathered and lichen covered characteristics, had its geometry sampled experimentally to see how an erratic displaying poor geomorphic characteristics in the context of ideal ¹⁰Be sampling would be affected when compared to 228B09, providing insight into the suitability of standard field sampling recommendations. 228B10's surface strike/dip of 326/59 and shielding factor of 0.821 augments the sample's geometric complexity, foreshortening its exposure to the cosmic-ray flux, i.e., abating the uniform cosmic-ray exposure history that a sufficiently flat and horizontal surface would have (Gosse and Phillips, 2001).

Upon collection, samples were sent to the Lamont-Doherty Earth Observatory (LDEO) in Palisades, New York for quartz extraction. Before the samples are subjected to the physical processes of crushing and sieving, they must be described in terms of average thickness and weight. A weighted average thickness of all pieces of an individual sample utilized for extracting quartz is then calculated and recorded. Samples were crushed using a hydraulic press, milled, and sieved to fractions varying between 63 and 710 µm. The rock crushing lab was thoroughly cleaned between samples to avoid ¹⁰Be contamination. We attempted the LDEO's (2012a) standard methods of quartz extraction from whole rock including techniques of froth flotation, heavy liquid mineral separation, and up to five leaches of 5% hydrofluoric acid (HF) per sample, but this was unsuccessful as the rock samples completely dissolved in HF yielding no extractable quartz (LDEO, 2012a). Subsequent XRD analysis of samples 228B01, 228B07, and 228B09 at the UNBC Northern Analytical Laboratory Services revealed a substantial percentage of quartz present in each rock sample; consequently, Rinterknecht's (2003) alternative procedures of Ophosphoric boiling were followed successfully. Sample 228B02 was not processed due to a lack of available material and time constraints.

Young samples of rock for use in cosmogenic surface exposure dating require significantly more quartz for ¹⁰Be measurement than older samples. As our sampled late Holocene moraines are particularly young, it was recommended to carry out an experimental procedure of sample preparation. All samples were spiked with an average of 107.13 µg ⁹Be and the remaining sample bulked with Fe instead of the typical $\sim 250 \ \mu g$ pure ⁹Be carrier. Using less ⁹Be carrier improves ¹⁰Be data precision when determining younger exposure ages. This approach allows for improved accuracy of ages, counting statistics, and requires ~40% less quartz compared to current standards (Braumann et al., 2021; Walker et al., 2022). Two low concentration process blanks were included for the nine samples (Hippe, 2017; LDEO, 2012b). Standard methods of Be extraction from quartz were used after ⁹Be spiking. This includes anion exchange columns for removal of Fe^{III} and cation exchange columns for the separation of Be from Al and Ti (LDEO, 2012b). Sample aliquots of extracted Be were shipped to Lawrence Livermore National Laboratory Center for Accelerator Mass Spectrometry (LLNL-CAMS) to carry out AMS measurement for beryllium-isotope ratio analyses relative to the ¹⁰Be/⁹Be standard 07KNSTD dilution series (Nishiizumi et al., 2007).

Surface exposure ages are reported and discussed as either years or thousands of years (ka) Before Present (BP) using 1950 CE as the reference year. Exposure ages were calculated using the online exposure age calculator formerly known as CRONUS-Earth v3

20

(https://hess.ess.washington.edu/) (Phillips et al., 2016). Ages are calculated using the Lifton-Sato-Dunai nuclide-dependent (LSDn) scaling framework with default Be atom reference production rates as no regional ¹⁰Be production rate calibration site for the YT is available (Borchers et al., 2016; Lifton, 2016; Lifton et al., 2014). Moraine ages and erratics are conservatively reported as the median \pm the interquartile range (IQR) rounded to the nearest decade. The median \pm IQR is less sensitive to extreme values and accounts for uncertainties such as geological scatter, providing a more conservative estimate of the moraine's true age (Darvill et al., 2022; Menounos et al., 2017). Measurement uncertainties for exposure age uncertainty (Table 2.4) is calculated as propagated error:

$$\sqrt{(\sigma_{int})^2 + (\sigma_{ext})^2} \tag{2}$$

Where σ_{int} is internal error and σ_{ext} is external error.

2.4.2 DEM creation and geodetic mass balance

Unlike geometric fluctuations in glacier length, which reflect long-term changes, mass balance records provide a more immediate response to changes in atmospheric conditions (Oerlemans, 1994; Zemp et al., 2009). Geodetic mass balance studies reflect changes in surface elevation based on assumed density values for water and ice (Beedle et al., 2014). To gain insight into glacier response to climate change over the past century, I estimated the geodetic mass balance for the Itsi Range using photogrammetric methods. To achieve this, I ordered aerial photo negative scans and associated fiducial calibration reports from the Natural Resources Canada (NRC) National Air Photo Library (NAPL) scanned at 2032 dpi and saved as TIF files (Table 2.2). These air photos provide nearly full spatial coverage of the Itsi Range on 8/24/1949 and 9/9/2004.

Year	Date	Roll	Photo	Scale	GCPs ^a	RMSE ^b (m)	RMSE (m)
		ID	Numbers			XY	Z
1949	24 Aug.	A12255	0367 - 0370, 0433 - 0436	1:40 000	34	4.780	5.853
2004	09 Sep.	A28519	0145 - 0148	1:40 000	20	3.640	4.156
Tatal number of Cround Control Doints (CCDs) used							

Table 2.2: Aerial photographs used for glacier change analysis.

^a Total number of Ground Control Points (GCPs) used.

^b Root Mean Square Error (RMSE) of GCP positions.

I created a ground control point geopackage in QGIS (v. 3.36.1) to georectify the aerial photos. I placed GCPs on stable, non-glaciated terrain using both a spaceborne DEM (ArcticDEM) and Planet imagery, both resampled to a 4 m resolution. The ArcticDEM mosaic is comprised of DEM strips taken between 9/12/2011 and 2/12/2021 (Porter et al., 2023). I aimed for 5 – 10 GCPs per photo, placing them on consistent morphological features among the suite of images such as pyramidal peaks, moraines, and distinctive bedrock features (Table 2.2). I saved GCP coordinates to CRS EPSG:4326 (WGS84) and exported them as a CSV file.

I first optimized the scanned air photos in Python using the Contrast Limited Adaptive Histogram Equalization (CLAHE) method from the OpenCV package to correct for film overexposure on glacier ice (Reza, 2004). I used Agisoft Metashape Professional Edition (v. 2.1.2) for DEM and orthomosaic construction from the air photos. I processed each roll of air photos separately, beginning with automated masking of scanned film borders and fiducial mark detection. Fiducial coordinates from the calibration report were entered in the camera calibration dialog box, with transformation type set to conformal. In the reference panel, the CRS was set to EPSG:4326 and marker accuracy set to 20 m. To generate a sparse cloud of tie points, I then aligned the photos at the lowest accuracy setting, set reference preselection to sequential, and applied masks to key points. Lower accuracy settings coarsen resolution but minimize voids in the DEM. After GCP import and assignment to each image, I built a point cloud at the lowest accuracy setting. I created the DEM from this point cloud and projected it to EPSG:3579 (NAD83(CSRS)/Yukon Albers). I then constructed an orthomosaic of the aerial photos based on the DEM (Figure 2.3).



Figure 2.3: Three-dimensional visualization of the 2004 orthomosaic of the Itsi Range, optimized using CLAHE. No vertical exaggeration is applied. Glaciers are outlined in red, outer moraine highlighted in yellow, and inner moraine highlighted in blue. Scale is approximate. Created in QGIS using the Qgis2threejs plugin.

To assess geodetic mass balance over the periods 1949 – 2004 and 2004 – 2022, I imported the newly georectified DEMs, the ArcticDEM mosaic, and an additional ArcticDEM strip captured on 9/19/2022 into a Python environment (v. 3.11.9) with the package xDEM (v. 0.0.19) (contributors, 2024; Porter et al., 2022). I identified fourteen glaciers in the Itsi Range present across all three rasters, and glacier outlines for each date were saved as vector polygon geopackages. I reprojected vectors and rasters to EPSG:3579 and resampled rasters to the lowest resolution DEM (photo A12255) to 8.11 m. I utilized the ArcticDEM mosaic as the reference for coregistration. The coregistration process required two non glacierized, stable ground inlier masks created from 1949 and 2004 glacier extents. The aerial photo-derived DEMs then undergo deramping by correcting directional biases using a 2nd order polynomial (Girod et al., 2017). I further processed the deramped DEMs and the 2022 DEM using a nonlinear blockwise coregistration approach, which divides each DEM into a 16 chunk grid (Nuth and Kääb, 2011). I assessed normalized median absolute deviation (NMAD) errors before and after coregistration to ensure an improved fit for each DEM (Table 2.3).

Table 2.3: NMAD errors before and after coregistration for the DEM of each year.

DEM year	NMAD before (m)	NMAD after (m)
1949	10.61	4.70
2004	4.70	2.08
2022	3.65	1.75

I created a difference-DEM (dDEM) for each period by subtracting the oldest year from the newest, resulting in dDEM 2004 – 1949 and dDEM 2022 – 2004 (Figure A.6). Outlier removal entailed using a distance filter for each dDEM with a radius of 1 pixel and mean outlier threshold of 50. An additional filter removed elevation difference values \geq 200 for dDEM 2004 – 1949 and values \geq 50 for dDEM 2022 – 2004. I divided the filtered dDEMs into 50 m mean elevation bins and fit to a third-order polynomial (Mukherjee et al., 2023). I used these fitted bins to derive geodetic mass balance (*B*) values in metres of water equivalent (m w.e.) using equation (3:

$$B = \frac{dV}{A} \cdot \frac{\rho_i}{\rho_w} \tag{3}$$

Where dV is total volume change, A is average area of glacier extents, and an assumed ice density conversion factor (ρ_i) of 850 kg m⁻³ and water density (ρ_w) of 1000 kg m⁻³ (Huss, 2013; Mukherjee et al., 2023).

Geodetic mass balance uncertainty is standardized using stable terrain as error proxy (Hugonnet et al., 2022). I calculated combined errors (E_{Δ}) , such as area and elevation change errors, by combining error terms (E_n) as shown in equation (4 (Tennant and Menounos, 2013).

$$E_{\Delta} = \sqrt{E_1^2 + E_2^2 + \dots + E_n^2}$$
(4)

Following methods by Menounos et al. (2019), I calculated random uncertainty on geodetic mass balance ($\sigma_{\Delta M,rdn}$) which derives from three independent sources of error: rate of elevation change uncertainty ($\sigma_{\Delta z}$), glacierized area uncertainty (σ_A), and volume to mass conversion uncertainty ($\sigma_{f\Delta V}$). After correcting for effective sample size, I derived elevation change uncertainty ($\sigma_{\Delta z}$) from uncertainty on stable terrain ($\sigma_{\Delta h}$):

$$\sigma_{\Delta z} = \sigma_{\Delta h} \cdot \sqrt{\frac{A_{cor}}{5A}}$$
(5)

Where A is glacier area and $A_{cor} = \pi L^2$. I assume a decorrelation length (L) of 500 m (Menounos et al., 2019). I derive glacierized area uncertainty (σ_A) using the summed perimeter of each glacier (P) and half the resolution of a pixel (dx). I set dx to match the resampled raster resolution of 8.11 m.

$$\sigma_A = P \cdot \left(\frac{dx}{2}\right) \tag{6}$$

I use these uncertainties to calculate non-systematic volume change uncertainty ($\sigma_{\Delta V}$):

$$\sigma_{\Delta V} = \sqrt{\left(\sigma_{\Delta z} \cdot \left(p_A + 5(1 - p_A)\right)A\right)^2 + (\sigma_A \cdot \Delta z)^2}$$
(7)
Where p_A represents the percentage of glacierized area after accounting for voids in the dDEM due to DEM construction and outlier filtering, and *A* is glacier area. Random uncertainty on geodetic mass balance ($\sigma_{\Delta M,rdn}$) is then derived using volume change uncertainty ($\sigma_{\Delta V}$):

$$\sigma_{\Delta M,rdn} = \sqrt{(\sigma_{\Delta V} \cdot f_{\Delta V})^2 + (\sigma_{f_{\Delta V}} \cdot \Delta V)^2}$$
(8)

Where ΔV is volume change, and the mass to volume conversion factor $(f_{\Delta V})$ and associated uncertainty $(\sigma_{f_{\Delta V}})$ are 850 kg m⁻³ and 60 kg m⁻³, respectively.

I calculated volumetric change error $(E_{\Delta V})$ following methods by Tennant and Menounos (2013):

$$E_{\Delta V} = \sqrt{\Delta H^2 E_A^2 + A^2 E_{\Delta H}^2} \tag{9}$$

Where ΔH is mean elevation change, $E_{\Delta H}$ is elevation change error, A is glacier area used for volume change calculations, and E_A is glacier area error.

2.4.3 Termini tracking

I used the Glacier Termini Change Tracking (GTT) toolbox, developed by Urbanski (2018) for ArcGIS, to calculate retreat rates of the Itsi Range glaciers using historic aerial photographs and satellite imagery. Following aforementioned methods, I constructed a high-resolution DEM (1 m) and orthomosaic from 9/9/2004 aerial photos which facilitated the precise delineation of moraines among the Itsi Range. I traced successive glacial termini for each glacier and date, storing them as 20 separate vector line shapefiles for estimating rates of glacial retreat. The Python-based GTT toolbox connected lines between each glacier terminus and moraine terminus proxy at 20 m intervals (Figure A.3). By using suitable nested moraines as proxies for

glacier termini and incorporating median moraine ages from the Itsi Glacier ¹⁰Be chronology, I extended the timeline for analyzing termini retreat rates to the LIA.

2.5 Results

2.5.1 Itsi Glacier ¹⁰Be chronology



Figure 2.4: Itsi Glacier and moraines. Aerial photo and Sentinel-2 derived glacier extents are overlayed for 1949, 2004, and 2022. DEM basemap was generated using stereoscopic aerial image pairs from 2004. Individually ¹⁰Be dated boulder ages from each moraine are included and reported with propagated error as LSDn exposure age BP.

A common geomorphologic feature crops up throughout the Mackenzie and Selwyn mountains: the preservation of nested moraines in close proximity to one another among alpine cirque glaciers (Figure 2.8) (Hawkins et al., 2023). The Itsi Glacier forefield contains two such nested moraines: an outer and inner moraine with median surface exposure ages of 960 years \pm 360 (n = 4) and 200 years \pm 50 (n = 3), respectively (Figure 2.5; Table 2.4). Both moraines are unvegetated, have minimal lichen cover, and are distinctly sharp-crested ridges suggesting a late Holocene age. Sampled inner moraine, outer moraine, and erratic boulders lie at average elevations of 1782 m, 1776 m, and 1757 m, respectively. In total, the Itsi Range contains nine pairs of morphologically similar nested moraines. The average distance to the headwall for all Itsi Range inner and outer moraines is 1268 and 1338 m, with corresponding average elevations of 1625 m and 1611 m.



Figure 2.5: ¹⁰Be ages from two neoglacial moraines emplaced by the Itsi Glacier reported in ka BP. Grey curves show ages of individual boulders rendered as Gaussian distributions. The coloured lines show the sum of each individual Gaussian distribution for the ¹⁰Be ages sorted by morphostratigraphic position. Box-and-whisker plots corresponding to each moraine denote the median \pm IQR, minimum, and maximum surface exposure ages.

Erratics beyond the outer moraine date to a median exposure age of 11.61 ka \pm 0.54 (Figure 2.6). Erratic 228B09, sampled utilizing standard methods, returned an exposure age of 10539 years \pm 706. Erratic 228B10, sampled following the experimental method described earlier, returned an exposure age of 12686 years \pm 917. This yielded results that were the inverse of what I expected as sampling 228B10 at a steep angle (59°) should foreshorten its exposure to the cosmic-ray flux, reduce ¹⁰Be production by spallation, and consequently return a younger age.



Figure 2.6: ¹⁰Be ages from two erratic boulders outboard the Itsi Glacier moraines reported in ka BP. The pink line shows the sum of each individual Gaussian distribution. Box-and-whisker plot denotes the median \pm IQR, minimum, and maximum surface exposure ages of the two erratics.

Sample	Latitude	Longitude	Elevation	Thickness	Shielding	Quartz	Carrier	Blank- corrected	Blank- corrected	Blank- corrected	Blank-corrected	Exposure	Exposure	AMS
ID	(d.d.)	(d.d.)	(m a.s.l.)	(cm)		(g)	added	¹⁰ Be/9Be ratio	¹⁰ Be/ ⁹ Be ratio	¹⁰ Be conc.	¹⁰ Be conc. 1σ	age (LSDn)	age uncertainty	facility
							(g) ^a		1σ uncertainty	(atoms g ⁻¹)	uncertainty (atoms g ⁻¹)	(yrs) ^{b, c}	(yrs)	
Outer mo	raine													
228B01	62.8904	-130.2817	1722	1.51	0.983	41.140	0.103	1.52E-13	2.91E-15	2.62E+04	5.75E+02	1156	82	LLNL-CAMS
228B03	62.8901	-130.2799	1722	1.88	0.991	40.141	0.102	1.04E-13	2.63E-15	1.82E+04	5.08E+02	761	59	LLNL-CAMS
228B04	62.8927	-130.2905	1750	1.71	0.986	23.153	0.102	1.53E-13	3.02E-15	4.66E+04	1.05E+03	2094	145	LLNL-CAMS
228B05	62.8928	-130.2903	1761	2.23	0.986	31.388	0.103	4.98E-14	1.55E-15	1.10E+04	3.82E+02	432	38	LLNL-CAMS
_												Median ± IQR	960 ± 360	
Inner mor	aine													
228B06	62.8938	-130.2913	1796	2.33	0.987	40.986	0.103	3.45E-14	1.48E-15	5.72E+03	2.73E+02	196	24	LLNL-CAMS
228B07	62.8938	-130.2910	1798	2.43	0.986	28.884	0.102	1.68E-14	1.35E-15	3.71E+03	3.46E+02	109	26	LLNL-CAMS
228B08	62.8936	-130.2916	1789	2.11	0.986	30.738	0.102	3.85E-14	1.48E-15	8.55E+03	3.65E+02	317	33	LLNL-CAMS
												Median ± IQR	200 ± 50	
Erratics														
228B09	62.8921	-130.2916	1757	1.39	0.982	13.974	0.102	4.42E-13	8.42E-15	2.24E+05	4.84E+03	10539	706	LLNL-CAMS
228B10	62.8921	-130.2916	1757	4.74	0.821	20.098	0.102	6.20E-13	1.67E-14	2.18E+05	6.28E+03	12686	917	LLNL-CAMS
												Median \pm	$11610 \pm$	
												IQR	540	

Table 2.4: Itsi Glacier ¹⁰Be sample information.

 ^a Be carrier concentration for all samples was 1045.9 ug g-1.
 ^b Ages are reported as years BP (1950 CE) and were calculated using the online exposure age calculator formerly known as CRONUS-Earth v3 found at https://hess.ess.washington.edu/ (last access: 13 November 2023) (wrapper 3.0.2, muons: 1A, constants as of 26 August 2020). Ages are calculated using the Lifton-Sato-Dunai (LSDn) scaling framework with the default Beryllium atom production rate.

^c Moraine ages and erratics are reported as the median \pm IQR rounded to the nearest decade.

2.5.2 Termini tracking

Results from the GTT tool record net retreat for all time periods (Table 2.5 and A.2). Yearly retreat rates ranged from 0.07 ± 0.14 to 8.31 ± 11.15 m yr⁻¹. Notably, the period from 2004 to 2022 records a marked increase in yearly retreat rate compared to the earlier periods with a more than two-fold increase from 1949 to 2022. The Itsi Glacier lost 50% of its glaciated area between 1949 and 2022 and contributed to a 43% percent reduction in total area for Itsi Range glaciers.

Table 2.5: Statistics derived from the GTT tool and area analysis among all glaciers of the Itsi Range.

Time period (CE) ^a	Elapsed years ^b	Mean yearly retreat rate (m yr ⁻¹)	Standard deviation (m yr ⁻¹)	Mean retreat distance (m)	Glacier count ^c	ΔArea %
$992 - 1754^{p}$	762	0.07	0.14	50	9	-
$1754 - 1949^{p}$	196	1.95	2.20	381	9	-
$1949 - 2004^{\circ}$	55	3.35	3.54	185	20	-26%
$2004-2022^{\rm o}$	18	8.31	11.15	149	20	-23%

^a Time periods marked by ^p and ^o denote proxy and observed, respectively.

^b Elapsed years corresponding to time period.

^c 9 of the 20 glaciers contained morphologically similar nested moraines appropriate for use as termini proxy. Two glaciers have distinguished tongues with independently developed moraines and are assessed as separate glaciers for retreat rate.

2.5.3 Geodetic mass balance

From 1949 to 2022, the total volumetric change among the 14 glaciers, which covered an area of 16.307 km² in 1949, amounted to a loss of -0.269 km³ \pm 0.207 (Table 2.6). The data reveal a significant increase in the rate of mass loss in the latter period. From 1949 to 2004, the glaciers record a negative geodetic mass balance of -0.190 m w.e. yr⁻¹ \pm 0.065. The period from 2004 to 2022, indicates the mass loss rate increased by 101%, with glaciers losing -0.382 m w.e. yr⁻¹ \pm 0.136, denoting accelerated mass loss over the past two decades. Area change among all glaciers of the Itsi Range follows similar trends (Table A.2).

Start year	End year	Elapsed	Analysis area	ΔV	Mass change
(CE)	(CE)	Years	(km ²)	(km ³)	$(m \text{ w.e. yr}^{-1})$
1949	2004	55	14.448	$\textbf{-0.178} \pm 0.173$	$\textbf{-0.190} \pm 0.065$
2004	2022	18	11.180	-0.091 ± 0.034	-0.382 ± 0.136

Table 2.6: Mass budget of Itsi Range glaciers (1949 – 2022).

2.6 Discussion

2.6.1 Geologic scatter

Constraining and correlating millennial-scale Holocene climate events from exposuredated alpine glacier chronologies requires knowledge of geologic scatter for effective mitigation (Balco, 2020a, 2011). Geologic scatter is a methodological issue stemming from geologic processes concerning the scatter in exposure ages derived from glacial landforms (Balco, 2011; Darvill, 2013; Heyman et al., 2011). Measurement uncertainty, post-depositional disturbance (unconformably young ages), and inheritance (unconformably old ages) can cause excess geologic scatter (Balco, 2020a, 2011; Heyman et al., 2011). Two primary issues are associated with the occurrence of geologic scatter: 1) it affects the ability to estimate central tendency, decreasing the accuracy of estimated sample ages and; 2) as geologic scatter reflects sample dispersion, i.e., variance for a normal distribution, a high dispersion of ages from an individual landform makes it difficult to attribute sample ages to a past climate event.

Averaging geological scatter produced by measurement uncertainty that is independent with normal distributions will yield a more accurate age; however, taking the mean will not improve accuracy of a calculated exposure age from scatter that results from geomorphic processes which affects exposure of a sample to the cosmic rays necessary for TCN production (Balco, 2011). Discerning between these two scatter types poses a challenge for TCND studies, yet accounting for uncertainty related to geologic scatter permits the confident attribution of derived ages as a direct result of climatic processes (Balco, 2020a, 2011; Goehring et al., 2022). The degree of geologic scatter has been estimated for each moraine using the reduced chi-squared (X^2) statistic as defined by Balco (2011):

$$X_R^2 = \frac{1}{n-1} \sum_{i=1}^n \left[t_i - \frac{\overline{t_i}}{\sigma t_i} \right]^2 \tag{10}$$

Where *n* is number of samples, t_i is the apparent exposure age, $\overline{t_i}$ is the mean exposure age, and σ is the measurement uncertainty (Crump et al., 2017). A calculated reduced X² statistic of less than 2 indicates minimal geologic scatter (Schaefer et al., 2009). Both moraines and erratics fail to pass the chi-squared statistical test (Table 2.7).

Utilizing the X² statistic assumes the sample distribution is normal (parametric), but this may not be the case for exposure ages from moraine boulders due to geologic scatter. The Kolmogorov-Smirnov (KS) test is a non-parametric test that can be used to determine if data are significantly similar to or different from a normal distribution (Massey Jr, 1951). To carry out the test, exposure ages are first standardized:

$$\frac{t_i - t}{\sigma_t} \tag{11}$$

Where \bar{t} is the mean and σ_t is the standard deviation. These standardized ages are then used for the KS test:

$$D_{stat} = max |F_{empirical}(t_i) - F_{normal}(t_i)|$$
(12)

Where $F_{empirical}(t_i)$ is the empirical cumulative distribution function (CDF) of the exposure ages, $F_{normal}(t_i)$ is the CDF of the normal distribution, and D_{stat} is the maximum difference between the two distributions (Massey Jr, 1951). The KS test is calculated in Python using the SciPy package. However, the KS test may become unreliable when dealing with the small sample sizes typical of moraine datasets (e.g., n = 2 to 6). Small sample sizes can return large p-values that misrepresent conclusions about the data's distribution. Because of this, the KS test is not used on individual moraine landform ages.

The non-parametric test Median Absolute Deviation (MAD) does not rely on the assumption of normality, which may be a better fit when considering the issue of geologic scatter skewing sampled moraine boulder distributions in this region. Unlike the mean and standard deviation, which can be disproportionately influenced by outliers, MAD is calculated as the median of the absolute deviations from the dataset's median, providing a better central tendency estimate when dealing with non-parametric data (Rousseeuw and Croux, 1993).

$$MAD = median(|t_i - \tilde{t}|) \tag{13}$$

Where t_i is the *i*th exposure age in the dataset and \tilde{t} is the median exposure age of the dataset, modified from Leys et al. (2013). A lower MAD value indicates that the data points are closely clustered around the median, suggesting minimal scatter, while a higher MAD value points to greater variability. In the context of exposure ages, a low MAD suggests that the ages are consistent, indicating minimal influence from geomorphic processes, while a high MAD may signal significant geologic scatter.

Table 2.7: Itsi Glacier moraine and erratic ages with accompanying reduced chi-squared (X2) and Mean Absolute Deviation (MAD) statistics.

Position	n	Median ± IQR	X^2	MAD
Outer moraine	4	$0.96 \text{ ka} \pm 0.36$	133.44	362
Inner moraine	3	$0.20 \text{ ka} \pm 0.05$	12.73	87
Erratics	2	$11.61 \text{ ka} \pm 0.54$	3.68	1073.5

Ice-cored moraines are moraines containing ice overlaid by sediment and disconnected from the active glacial ice-margin (Lukas, 2011). An ice-cored moraine must be considered a transitional landform as the establishment of a stable moraine requires the melt out of all ice incorporated into the till (Lukas, 2011). The delay of surface stabilization caused by the melting of dead-ice may not accurately reflect the age of the landform (Kirkbride and Winkler, 2012). The thaw of ice-cored moraines has potentially contributed to observations of scatter (Goehring et al., 2022; Hawkins et al., 2023) in this region as evidence for anthropogenic warming driving geomorphic transformation of glacial deposits in the Canadian Arctic has emerged in the last decade (Crump et al., 2017; Kokelj et al., 2017; Segal et al., 2016). As warming accelerates in Arctic regions, the melting of permafrost in ice-cored moraines threatens to mobilize sediment stores causing increased thaw slump activity (Kokelj et al., 2017; Obu et al., 2019; Segal et al., 2016). Melt of dead-ice in ice-cored moraines, causing a thaw slump beneath the layer of moraine debris, may contribute to post-depositional disturbance causing the dated exposure ages to appear younger than the true landform age – a form of self-censoring (e.g., Crump et al., 2017; Kirkbride and Winkler, 2012).



Figure 2.7: A regional overview of permafrost zone coverage in relation to the Itsi Range field site from Obu et al. (2019).

The surficial geology of the field site is described as Neoglacial till wherein 'end moraines may be tens of metres thick and contain masses of buried glacial ice,' lending credence to the plausibility of thaw slump induced post-depositional disturbance (Jackson et al., 1993). However, field observations of the moraines revealed that they are matrix-supported with little evidence of permafrost thaw, such as thermokarst features or morphological irregularities from differential ice core melting (Figure A.5). This suggests that thaw-related processes likely contribute minimally to geologic scatter in the Itsi Range. Subglacially entrained debris is of equal or greater importance than supraglacially entrained debris in low-relief glaciated valleys, significant because a fundamental assumption of TCND is that the rocks are subglacially, not supraglacially, i.e., rockfall, derived (Barr and Lovell, 2014). Rockfall derived supraglacially transported boulders bearing inherited nuclides and deposited as a moraine would represent maximum ages that also do not reflect the true depositional age (Hawkins, 2023). All lines of evidence considered, recent literature suggests post-depositional modification plays a less significant role for cirque glaciers than boulders derived from plucking of pre-exposed boulders from a glacier valley and boulders reincorporated from adjacent rockfall (Goehring et al., 2022; Tomkins et al., 2021).

2.6.2 Inner moraine

Boulders sampled from the inner moraine yield a median exposure age of 0.20 ka \pm 0.05 (Table 2.7). The dating uncertainty of glacier advances generally increases with age, so recent advances (e.g., LIA) are more confidently dated than older (e.g., LGM) ones (Kirkbride and Winkler, 2012). For the inner moraine, I assert that the large X² value of 12.73 does not indicate the presence of significant geological scatter, rather, it reflects the small sample size and narrow range of uncertainty associated with the extremely young ages. Exceptionally young surface exposure ages have narrow uncertainty windows. This small uncertainty, coupled with the heightened landform instability of young moraines (Tomkins et al., 2021), makes the ages less likely to overlap. The MAD value means that, on average, the exposure ages for the inner moraine deviate from the median by 87 years. This represents a close clustering of ages around the median; since the inner moraine IQR is also low, the MAD reflects a narrow time range of glacial activity.

Conversely, strict interpretation of the reduced chi-squared statistic of 12.73 implies that geomorphological factors beyond random measurement errors affect the surface exposure ages. For young moraine landforms (<1 ka), steep ice-proximal slopes reflect instability of the inchoate landform, which makes sampling moraine crests a poor strategy if making efforts to minimize boulder instability (Tomkins et al., 2021). An improved strategy of future sampling recommendations for presumably young, latest Holocene landforms should avoid solely sampling from moraine crests and instead sample from alternative intra-landform locations of ice-proximal and ice-distal slopes.

2.6.3 Outer moraine

Boulders sampled from the outer moraine yield a median exposure age of 0.96 ka \pm 0.36 (Table 2.7). This moraine displays obvious signs of geological scatter, returning a X² of 133.44, further supported by a MAD of 362 years. The anomalously old ages probably reflect accumulations of inherited nuclides. Melting permafrost in ice-cored moraines likely did not cause thaw slumps or contribute to the observed scatter, as boulder ages skew older than expected for Holocene maximum extents. Moraine boulders derived from plucking of pre-exposed boulders from a glacier valley and boulders reincorporated from adjacent rockfall may lead to inheritance in a cosmogenic nuclide dated sample. Regional observations of seismically induced large-scale rockslides further support mechanisms of inheritance rather than thaw slump induced post-depositional disturbance (Smith and Bennett, 2004).

Observations in the field (Figure 2.8) and analysis of the 2004 DEM (Figure 2.4) show an elongated ridge above the inner moraine, proximal to the ice-margin, signifying exposed rock repeatedly abraded during periods of less extensive glacier limits. This rock could have been subsequently overridden and redeposited at the existing terminal moraine (Balco, 2020a). If we

assume that the exposure ages are accurate which are a consequence of the actual climatic processes that led to deposition of the moraine, it must be considered that the moraine consists of an amalgamation of prior overridden advances; a representation of multiple neoglacial advances and retreats over the past two millennia.



Figure 2.8: An oblique angle of the Itsi Glacier forefield highlighting the inner and outer moraines as well as an elongated ridge of glacially abraded rock proximal to the ice margin and parallel to the moraine axes.

2.6.4 Erratics

Sampled erratic boulders yield a median exposure age of $11.61 \text{ ka} \pm 0.54$ (Table 2.7) denoting precise accordance of CIS melt with the end of the YD stadial recorded globally (Broecker et al., 2010; Menounos et al., 2017). This suggests that as the CIS deteriorated, it transitioned into a large valley glacier evidenced by the deposition of a large moraine far downvalley of the Itsi Glacier outer moraine. The valley glacier subsequently thinned and melted away, depositing the dated erratics immediately beyond the Itsi Glacier outer moraine. This indicates that the CIS was no more extensive than the erratics' position at this time. Both erratics fail the reduced chi-squared test albeit with a low chi-squared value of 3.68. The very high MAD value of 1073.5 years highlights extreme deviation from the median, but the small sample size (n = 2) limits ability for strict interpretation and indicates poor statistical robustness. Considering this, the boulders returned ages consistent with the regionally interpreted timeline of CIS decay, implying sample 228B10 was negligibly affected by the steep extraction angle (Menounos et al., 2017). This suggests that, at least at this site, the extraction angle may be less significant in influencing exposure ages than previously thought.

2.6.5 Geodetic mass balance and termini retreat

The large standard deviations associated with glacier terminus mean retreat rates reflect both the morphological diversity among glacier termini and the method used for their calculation (Urbanski, 2018). ERA5-Land data for the Itsi Range from 1950 to 2023 helps contextualize the significant increase in retreat rate and mass loss since deposition of the Itsi Glacier outer moraine (Tables 2.5 and 2.6). Glaciers in interior continental regions at high altitudes, such as the Itsi Glacier, primarily respond to fluctuations in summer temperatures during the melt season (Roe and O'Neal, 2009). Boreal summer and winter trends reveal a steady increase in mean temperature and precipitation along with a decrease in summer snow cover since 1950 (Figure 2.9). Although an increase in mean temperature is the most direct contributor to termini retreat and loss of glaciated area among Itsi Range glaciers, increasing summer precipitation on diminishing snow cover can further enhance glacier surface melting. MAT anomalies for 2023 provide additional evidence of this accelerated warming trend (Figure 2.10).



Figure 2.9: ERA5-Land values for the Itsi Range (62.9°, -130.2°) plotted to represent boreal summer melt (June, July, August (JJA)) and winter accumulation (December, January, February

(DJF)) seasonal means. Values are plotted as five-year moving averages with associated trendlines.



Figure 2.10: ERA5-Land 2023 MAT anomalies for the YT using reference period 1950-1980. Itsi Range is highlighted in green.

The Itsi Range straddles the Upper Pelly and Macmillan drainage basins within the greater Pelly catchment. Mean monthly discharge observations from the Ross River station, situated downstream of the Itsi Range in the Upper Pelly basin, corroborates with modelled ERA5 temperature and snow cover trends. Discharge trendlines illustrate a shift to an earlier melt season peaking in May, a departure from historically consistent observations of peak discharge values in June (Figure 2.11) (Smith and Bennett, 2004). This shift results in an earlier disappearance of snow cover on Itsi Range glaciers and consequently earlier exposure of bare ice. Exposed glacier ice accumulates light-absorbing impurities (e.g., algae, dust, supraglacial

meltwater, etc.) which reduces overall albedo, and, in combination with a prolongation of melt season as indicated by discharge trends, amplifies mass loss (Naegeli et al., 2019).



Figure 2.11: Observed mean discharge values at Ross River (61.99, -132.41; Station ID: 09BA001) within the Upper Pelly sub-sub-drainage area for May and June. Extracted from the Environment and Climate Change Canada Historical Hydrometric Data web site (https://wateroffice.ec.gc.ca/mainmenu/historical_data_index_e.html).

I compared the geodetic mass balance records obtained for the Itsi Range with those published by Hugonnet et al. (2021), utilizing the same glaciers to assess the consistency of volumetric and mass change values (Table 2.8). The results from this study show a volumetric change of -0.091 km³ \pm 0.034 over 18 years, with a corresponding mass change rate of -0.382 m w.e. yr⁻¹ \pm 0.136 across an analysis area of 11.180 km². In comparison, Hugonnet et al. (2021) reported a volumetric change of -0.180 km³ \pm 0.331 over 15 years, with a mass change rate of -0.692 m w.e. yr⁻¹ \pm 0.649 for a larger analysis area of 15.022 km².

Table 2.8: Mass budget of Itsi Range glaciers compared to Hugonnet et al. (2021).

Source	Start	End	Elapsed	Analysis area	ΔV	Mass change
	year	year	years	(km^2)	$(km^3)^a$	$(m \text{ w.e. yr}^{-1})^a$
This study	2004	2022	19	11.180	$\textbf{-0.091} \pm 0.034$	-0.382 ± 0.136
Hugonnet et al. (2021)	2004	2019	16	15.022	$\textbf{-0.180} \pm \textbf{0.331}$	$\textbf{-0.649} \pm 0.609$

^a Uncertainty estimations for $E_{\Delta V}$ and $\sigma_{\Delta M,rdn}$ match those described in section 2.4.2.

A portion of the differences in the volumetric and mass change values may be attributed to the sources used for analysis. Hugonnet et al. (2021) use the Randolph Glacier Inventory (RGI 6.0) in tandem with Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) DEMs. In contrast, I created dynamic glacier outlines for each year and processed higher resolution (8.11 m) DEMs sourced from aerial photographs and ArcticDEM. The RGI delineations are inherently less accurate due to the automated nature of its creation and static area over the assessed time period. Additionally, the 30 m resolution of ASTER DEMs heightens potential for error. Despite these methodological differences, both studies corroborate the overall trend of accelerated glacier mass loss in the Itsi Range and suggest that this study's geodetic mass balance estimations are conservative. These trends in mass loss suggest that, due to the lagged glacier response to climate fluctuations, retreat will continue even if the increase of temperature anomalies ceased, as glaciers in the Itsi Range remain in disequilibrium with the warming climate (Marshall et al., 2011).

2.7 Conclusion

I used TCND techniques on a pair of nested moraines to develop a latest Holocene chronology of the Itsi Glacier among the Itsi Range in the YT. Erratics from this site-specific chronology suggests that the decaying CIS exposed Itsi Range cirque glaciers correlative to the end of the YD, meaning that cirque glacier advances at this time were less extensive than their LIA extents. The most extensive cirque glacier advances occurred during the latest Holocene, recorded as two distinct pulses and further evidenced by the existence of several nested moraine pairs throughout the massif. Inner moraine exposure age of 0.20 ka \pm 0.05 is minimally affected by geologic scatter with uncertainty a reflection of moraine crest instability due to the young age of the landform, limited sample size, and narrow ranges of age uncertainty among individual

samples. Conversely, the outer moraine age of 0.96 ka \pm 0.36 exemplifies the issues of geologic scatter commonly reported in the northeastern Canadian Cordillera. The lack of further evidence allows for the following conclusions to be drawn about the outer moraine: 1) the oldest boulder age of 2.09 ka \pm 0.15 is correct and corresponds to the time of moraine construction, with younger boulder ages reflecting subsequent moraine stabilization, 2) the moraine reflects the thaw slump of an ice-cored moraine causing erroneously young ages, 3) the moraine reflects processes of inheritance due to plucking of pre-exposed boulders from the glacier forefield and/or from adjacent rockfall causing erroneously old ages, 4) all boulder ages are true and accurately reflect regional climatic processes, indicating an accretion of overridden glacier advances.

GTT derived rates of glacier retreat and geodetic mass balance results for Itsi Range glaciers point to an escalating trend of retreat in recent decades attributed to a consistent increase in MAT and decrease in summer snow cover since 1950. Discharge observations downstream from the Itsi Range since 1960 capture evidence of a shift to an earlier melt season which will propagate trends of glacier mass loss in this region. ERA5-Land temperature and snow cover trends further substantiate the effects of rapid warming that began in the 20th century.

3. Comparison with western North America moraine records

3.1 Introduction

The evidence of Holocene glacier fluctuations within the northeastern Canadian Cordillera remains sparse. Chapter 3 seeks to compare the Itsi Glacier ¹⁰Be chronology with available regional records of moraine and erratic ages. In addition, I provide complementary preexisting paleoclimate chronometer and proxy records from radiocarbon, dendrochronology, and lake sediment cores to further improve understanding of how these glacial landforms fluctuated throughout the latest Holocene.

Both the frequency and magnitude of climate drivers and the rate and duration of glacier advance contribute to the moraine record glaciers emplace over time (Kirkbride and Winkler, 2012). A theoretical experiment conducted by Kirkbride and Winkler (2012) found that glaciers with short response times to fluctuations of climate are more sensitive and record more advances in the moraine record than glaciers with long response times, wherein the process of obliterative overlap begins to censor the moraine record. If multiple glaciers have a similar age for a given moraine, it signifies an advance likely occurred followed by a long period of retreat, decreasing the likelihood of censoring by obliterative overlap (Kirkbride and Winkler, 2012). This reasoning provides the paradigm for comparing regional glacial records to draw paleoclimatic conclusions, as a lower likelihood of erosion censoring reduces opportunities for false correlations among geographically discrete glacier advances. Obliterative overlap, in the form of late Holocene glacier expansion, is the most likely reason for the lack of preservation of pre-Little Ice Age (LIA) moraines at the Itsi Glacier site.

3.2 Methodology

I primarily sourced terrestrial *in situ* cosmogenic nuclide (TCN) dated moraine boulder ages from the Informal Cosmogenic-Nuclide Exposure-age Database (ICE-D): Alpine project (Balco, 2020b). As few TCND studies have been carried out in the northeastern Canadian Cordillera, this chapter incorporates all available late Holocene ¹⁰Be dates from moraines in Alaska (AK), British Columbia (BC), Yukon Territory (YT), and Northwest Territories (NWT). I first recalculated ¹⁰Be samples to modern standards matching those described in chapter 2 to ensure consistency across regional comparisons. I assessed all relevant ¹⁰Be dated moraine positions in Western North America (WNA) and assigned them one of four positions as defined in Table 3.1. Rather than relying solely on ¹⁰Be-dated moraine ages, morphostratigraphic position was the primary criterion for assigning end groups, with moraine age the secondary criterion. This approach minimizes the impact of geologic scatter on moraine age estimates and allows for the identification of potential patterns in scatter associated with morphostratigraphic position. I normalized reported ages to ka BP and present them as median \pm interquartile range (IQR). To assess the presence of geologic scatter, I first applied the KS test to determine whether a parametric or non-parametric statistical approach was more appropriate. For parametric groups I report the reduced chi-squared statistic, whereas the MAD is provided for non-parametric groups.

Position	Age ^a	Sample locations ^b	Description
End Group 1	LIA (~0.90 – 0.06 ka)	AK, BC, NWT, YT	Singular innermost recessional or lateral moraine proximal to ice-margin.
End Group 2	LIA (~0.90 – 0.06 ka)	AK, NWT, YT	Any recessional or lateral moraines distal to End Group 1.
End Group 3	Pre-LIA (~4.40 – 1.20 ka)	AK, BC	Any recessional or lateral moraines distal to End Group 1 and/or 2.
Erratics	CIS Decay (~12.7 ka)	BC, NWT, YT	Erratic boulders.

Table 3.1: Description for each morphostratigraphic position used for WNA comparisons.

^aEnd Group 1 – 3 ages estimated using WNA Neoglacial advances discussed in chapter 3.3.1. ^bAK = Alaska, BC = British Columbia, NWT = Northwest Territories, and YT = Yukon Territory.

In addition to ¹⁰Be surface exposure ages, I grouped and assessed proxies derived by methods of radiocarbon dating, dendrochronology, and glaciolacustrine sedimentation records. These proxies, synthesized from numerous publications, correspond to intervals of glacial advance reported in the literature (Barclay et al., 2009; Menounos et al., 2009; Mood and Smith, 2015; Tomkins et al., 2008).

Radiocarbon dating techniques provide ages for when a carbon-based organism has died, mainly as subfossil wood (Braumann et al., 2020; Mood and Smith, 2015). Radiocarbon dating's relevance in reconstructing glacier chronologies relies on the assumption that advancing glaciers may kill trees growing within glacial valleys, preserving them as glaciogenic deposits. During subsequent glacier retreat, overridden wood becomes exposed. Once exposed, the sampled wood can provide a radiocarbon-derived kill date assigned to a glacier advance (Ryder and Thomson, 1986).

Dendrochronology, utilizes the growth of tree-rings, useful for resolving high temporal resolution glacier histories during the LIA (Jacoby and Cook, 1981; Osborn et al., 2001; Tomkins et al., 2008). The dendrochronology dating method works by analyzing the patterns of tree ring widths and ring densities, which vary annually due to a variety of environmental conditions. Tree rings record an aggregation of multiple environmental changes; therefore, depending on the application, e.g., correlating with summer temperature records, the desired signal must first be extracted from the noise of irrelevant contributing factors (Cook and Kairiukstis, 1990).

Glacially distal lacustrine sediments present a high resolution proxy record that records influxes of sediment that reflect regional environmental changes helpful in reconstructing regional paleoclimate and related glacial histories (Leonard, 1997; Nesje et al., 2001; Tomkins et al., 2008). Grain size variations in glaciolacustrine sediment records reflect the activity of wetbased, temperate glaciers (Nesje et al., 2001). Organic content is estimated using percent weight loss-on-ignition (LOI), which measures sample weight before and after high-temperature exposure. Increased erosion and input of clastic sediments, indicating heightened glacial activity, can be inferred from increases in magnetic susceptibility. Together, LOI and magnetic

50

susceptibility data enable continuous intercalation of Holocene glacier fluctuations (Maurer et al., 2012; Nesje et al., 2001). Although radiocarbon dating and dendrochronology primarily provide limiting ages, when combined with glaciolacustrine sedimentation records and/or ¹⁰Be dating, they enhance the precision of glacier fluctuation chronologies by better constraining the time series of glacier advances (Hawkins et al., 2021; Osborn et al., 2007).

For climatic context, I review the timing of alpine glacier advances among WNA from the early Neoglacial to the present, drawn from these published sources. I compiled these advances to represent minimum and maximum ages of advance (e.g., for a bracketed age of 1.9 – 1.4 ka, 1.9 ka would be the maximum-limiting age). Periods of glacial advance over the past millennia are presented visually using Python (v. 3.11.7).

3.3 Discussion of moraine records

3.3.1 Review of western North America Neoglacial advances

The Northern Hemisphere (NH) experienced generally cold, glacially extensive period following the Holocene Thermal Maximum (HTM); however, there still exists a significant degree of intrinsic climate variability (Clague et al., 2009; Renssen et al., 2009). Radiocarbon ages record two distinct early-middle Neoglacial advances between 4.40 - 3.97 ka and 3.70 -3.40 ka in the western Canadian Cordillera (Menounos et al., 2009; Mood and Smith, 2015). Generally, advances were slightly less extensive during the middle-to-late Neoglacial than during the LIA. Western Canadian Cordillera and Alaskan (AK) glaciers record three distinct advances during this period evidenced by dendrochronology and radiocarbon kill dates at 3.54 - 2.50 ka, 2.20 - 2.00 ka, and 1.71 - 1.20 ka. Literature frequently refer to the advance between 3.54 - 2.50ka as the contemporaneous Tiedemann-Peyto Advance of the BC Coast and Rocky Mountains (Luckman et al., 1993; Menounos et al., 2009; Ryder and Thomson, 1986). The Tiedemann-Peyto Advance is complex, more accurately characterized by multiple periods of glacier advance and retreat (Menounos et al., 2009). In AK, Barclay et al. (2009) reported radiocarbon evidence for an advance at 2.20 - 2.00 ka. Dendrochronology and radiocarbon evidence exists for a third advance interval of similar complexity to the Tiedemann-Peyto Advance, often referred to as the First Millennium Advance, throughout the western Canadian Cordillera and AK occurring during the interval 1.71 - 1.20 ka (Barclay et al., 2009; Menounos et al., 2009; Mood and Smith, 2015; Reyes et al., 2006). St. Elias Mountain valley glaciers advanced from 3.30 - 2.40 ka, coeval with the Tiedemann-Peyto Advance; this contrasts with younger ages reported within this region during the First Millennium Advance from 1.25 - 1.05 ka (Menounos et al., 2009). The large interval of time ascribed to the First Millennium Advance suggests complexity of glacial behavior akin to the Tiedemann-Peyto Advance (Menounos et al., 2009).

Neoglacial glacier advances occurred globally during the LIA, but there is great spatial variability in the timing of these advances at the centennial scale (Kaplan et al., 2016; Schaefer et al., 2009; Solomina et al., 2015). Many Canadian Cordillera glaciers attained their Holocene glacial limits during the mid-late 19^{th} century (Menounos et al., 2013a, 2009; Solomina et al., 2015). Dendrochronology, radiocarbon, and glaciolacustrine evidence identifies three phases of advance throughout WNA, all generally referred to as the LIA, from 0.80 - 0.40 ka, 0.50 - 0.24 ka, and 0.40 - 0.06 ka (Barclay et al., 2009; Hawkins et al., 2021; Luckman, 2000; Menounos et al., 2009; Mood and Smith, 2015; Solomina et al., 2015; Tomkins et al., 2008). The compilation of major WNA advances are presented in Table A.1

3.3.2 Radiocarbon, dendrochronology, and glaciolacustrine records

When a glacier enters equilibrium with the predominating climate, emplacement of terminal and lateral moraines occurs (Solomina et al., 2015). This means that TCND derived moraine ages do not directly indicate when the glacier started advancing; rather, it signifies the culmination of advance, i.e., onset of retreat. Itsi Glacier moraine ages correlate variably with glacial advances across WNA (Figure 3.1).



Figure 3.1: Regional compilation of glacial advance intervals from 4.0 ka to present based on regional reviews using evidence from radiocarbon dating, dendrochronology, and glaciolacustrine sedimentation records (Barclay et al., 2009; Hawkins et al., 2021; Menounos et al., 2009; Mood and Smith, 2015; Tomkins et al., 2008). Grey box widths are set by duration of advance interval with ages calibrated to ka BP 1950. Dashed lines indicate approximate age intervals when discrete ages were not provided. Orange and blue vertical bars are Itsi Glacier median moraine ages, signifying moraine stabilization and glacier retreat; bar widths are set by the IQR of ¹⁰Be dated moraine ages. Pink vertical bars denote ages of individual Itsi Glacier outer moraine boulders which comprise the median outer moraine age; bar widths are set by the propagated error of each boulder.

The outer moraine dates to approximately 250 years after the First Millennium Advance,

1.71 – 1.20 ka (Figure 3.1). The large IQR and scattered exposure ages suggest multiple

possibilities for moraine genesis as discussed in chapter 2. One explanation is that this moraine

includes boulders from smaller moraines that were destroyed when the glacier attained its

maximum Holocene extent. Extending the glacial advance record and plotting individual outer moraine boulder ages reveals the complexity of late Holocene advances in this region. The visual correlation in Figure 3.1 establishes a plausible relationship between glacial advances and outer moraine boulders. Notably, the onset of three advances precedes outer moraine boulder ages when accounting for error. In Alaska and western Canada, the record of middle-late Neoglacial regional advances at 2.20 - 1.80 ka and 1.71 - 1.20 ka align well with the outer moraine boulder ages of 2.09 ka \pm 0.15 and 1.16 ka \pm 0.08, respectively (Barclay et al., 2009; Hawkins et al., 2021; Menounos et al., 2009; Mood and Smith, 2015). Additionally, the latest Neoglacial regional advance from 0.90 - 0.40 ka coincides with the outer moraine boulders dated to 0.43 ka \pm 0.04 and 0.76 ka \pm 0.06 (Barclay et al., 2009; Hawkins et al., 2021; Menounos et al., 2009; Mood and Smith, 2015; Solomina et al., 2015). Glacial lake sediments assessed by Tomkins et al. (2008) indicate an advance among the southern Selwyn Mountains that ended at 0.50 ka, but the record does not extend prior to 0.56 ka, leaving timing for the onset of this advance uncertain. The advance likely correlates with other regional records beginning around 0.90 – 0.77 ka (Barclay et al., 2009; Hawkins et al., 2021; Mood and Smith, 2015).

The Holocene maximum extent of the Itsi Glacier occurred at 0.96 ka \pm 0.36. In contrast, Tomkins et al. (2008) cite Holocene maxima in the southern Selwyn Mountains at 0.17 – 0.06 ka. Menounos et al. (2009) and Mood and Smith (2015) similarly interpret later maximum Holocene extents during the 19th century. Hawkins et al. (2023) identify an earlier Holocene maximum at 0.39 ka. Both the Itsi Glacier and the sites sampled by Hawkins et al. (2023) in the Selwyn and Mackenzie Mountains display nested moraine formations beyond alpine cirques. Sample 228B05 from the Itsi Glacier end group 2 moraine provides an age of 0.43 ka \pm 0.04, plausibly making this the true minimum-limiting age for LIA maximum extents if accepting the interpretation that the moraine represents an amalgamation of multiple glacial advances; or, if interpreting the other end group 2 samples as containing inherited nuclides. Evidence for a circa 0.40 ka Holocene maximum is further supported by Barclay et al. (2009), who report southern coastal AK glaciers attained Holocene maximum extents during both middle and late LIA advance phases circa 0.41 - 0.24 ka and 0.14 - 0.07 ka. Given the limited glacier chronology records, this implies that in the northeastern Canadian Cordillera, LIA maximum extents were reached during the 0.90 - 0.40 ka advance, centuries earlier than in the western and southwestern Canadian Cordillera.

The Itsi Glacier inner moraine was emplaced at $0.20 \text{ ka} \pm 0.05$ and is interpreted to represent the culmination of a singular regional advance from 0.40 - 0.06 ka. Specifically, Itsi Glacier inner moraine ages correspond closely to an advance in the southern Selwyn Mountains prior to onset of glacial recession from 0.22 - 0.17 ka (Tomkins et al., 2008). Coeval intervals of advance are observed in the Coast Mountains of BC at 0.24 - 0.07 ka and the Sierra Nevada Mountains of the American Cordillera at 0.25 - 0.17 ka (Mood and Smith, 2015; Solomina et al., 2015). Varved lake sediments in the southern Selwyn Mountains record a later glacial recession starting at 0.06 ka, suggesting inner moraine surface exposure ages represent maximum limiting ages for the culmination of the glacial advance (Tomkins et al., 2008).

3.3.3 ¹⁰Be moraine ages

Hawkins et al. (2023) reported consistent LIA exposure ages circa 0.30 and 0.10 ka from outer and inner moraine pairs in the northeastern Canadian Cordillera, along with outliers ranging from 0.99 ka to 4.31 ka. Similar observations of scatter have been reported for late Pleistocene moraines along the northwestern margin of the Cordilleran Ice Sheet (CIS) and among mountain glaciers of the Grey Hunter massif in the YT, making clear interpretations of these ages difficult (Goehring et al., 2022; Stroeven et al., 2014). A critical question in the field of TCND is whether these outlier ages represent true landform ages or geological scatter. The MAD is used for discussion of geologic scatter as all end groups reject the null hypothesis of normality as posed by the KS test (Table 3.2).



Figure 3.2: Regional compilation of WNA late Holocene ¹⁰Be boulder ages, including those collected for this thesis, sampled from moraines and grouped by morphostratigraphic position (Dortch et al., 2010; Hawkins et al., 2023, 2021; Howley, 2008; Menounos et al., 2013a; Pendleton et al., 2017; Young et al., 2009). Thin grey curves show ages of individual boulders rendered as Gaussian distributions. Thick curves represent the sum of each Gaussian distribution for the entire population of ¹⁰Be moraine group ages, organised by colour. Itsi Glacier moraine groups are inclusive in the regional curves and also highlighted separately within the summed Gaussian distributions. Corresponding box-and-whisker plots denote the median \pm IQR, minimum, and maximum surface exposure ages of each moraine group.

Itsi Glacier end group 1 ages accord well with the regional ¹⁰Be dated moraine record (Figure 3.2 and A.1). Although we did not sample lateral moraines for this study, I included them in end group 1 to provide better context for the complex timing of glacial standstills/retreats in WNA during the last millennium. ¹⁰Be moraine ages throughout the Canadian Cordillera (n = 21) and AK (n = 4) provide median end group 1 ages of 0.13 ka ± 0.1 (n = 25). The KS test

yielded a p-value of less than 0.05, indicating End Group 1 is a non-parametric distribution of ages (Table 3.2). The MAD of 67 years reflects very low variability among the exposure ages in this group, suggesting that the ages in End Group 1 likely correspond to a specific climatic event.

End group 1 moraines sampled at high latitudes ($62.1^{\circ} - 63.6^{\circ}$) across AK, NWT, and YT date to 0.30 ka ± 0.10 (n = 11) (Dortch et al., 2010; Hawkins et al., 2023). At southern latitudes (50.9°), lateral moraines in the BC Coast Mountains record end group 1 emplacement at 0.09 ka ± 0.03 (n = 14) (Hawkins et al., 2021). End group 1 moraines show less geological scatter than end group 2, the latter returning an MAD of 201 years (Table 3.2).

Table 3.2: ¹⁰Be dated WNA moraines grouped by morphostratigraphic position with associated statistics (Dortch et al., 2010; Hawkins et al., 2023, 2021; Howley, 2008; Menounos et al., 2013a; Pendleton et al., 2017; Young et al., 2009).

Position	n	Median (ka)	IQR (ka)	X^2	MAD	D-stat ^a	p-value ^b
End group 1	25	0.13	0.10	10.82	67	0.283	0.029
End group 2	36	0.46	0.35	129.18	201	0.303	0.002
End group 3	42	2.05	0.69	321.50	787	0.436	0.000

^a KS test D-statistic: the maximum difference between the empirical CDF of the sample and the CDF of the normal distribution (Massey Jr, 1951).

^b KS test p-value: the probability that the sample comes from a normal distribution. A value less than 0.05 indicates a rejection of the null hypothesis, indicating data are not normally distributed (Massey Jr, 1951).

Itsi Glacier end group 2 ages fall within the 2σ uncertainty represented by the Gaussian distributions, skewing older than the regional record (Figure 3.2). Hawkins et al. (2023) dated eight outer moraines from paired LIA moraines among the Mackenzie and Selwyn mountain ranges. Out of 23 moraine group 2 samples, eight returned anomalously old ¹⁰Be ages, purportedly due to inheritance, ranging from 0.99 ka ± 0.10 to 4.31 ka ± 0.29. This pattern is consistent among end group 2 moraines in the Mackenzie and Selwyn Mountains, with all anomalously old ages in the region possessing a combined median of 2.29 ka ± 0.83 (*n* = 10). These ages closely resemble AK end group 3 ¹⁰Be ages at northern latitudes (63° – 68°), which have a combined median of 2.33 ka ± 0.63 (*n* = 20) (Pendleton et al., 2017; Young et al., 2009),

and an end group 3 moraine at the Tiedemann Glacier in BC dated to 2.60 ± 0.28 (n = 6) (Menounos et al., 2013a).

End group 3 gaussian distributions illustrate significant overlap with end group 2, largely due to instances of geologic scatter among end group 2 moraines. The MAD values of 201 and 787 years for end groups 2 and 3 respectively. The exposure ages in these groups do not represent a coherent climatic event and may reflect either a broader range of geological processes, such as inheritance or post-deposition disturbance, or multiple climatic events affecting the moraine formation. Alpine glaciers in northern and interior AK, as well as glaciers among the Wrangell and Saint Elias mountains, contain pre-LIA moraines rarely preserved within WNA due to obliterative overlap achieved during LIA maximum glacial extents (Barclay et al., 2009; Denton and Karlén, 1977; Pendleton et al., 2017). In the Brooks Range of AK, moraines overridden by pre-LIA advances record maximum Holocene extents between 3.5 - 2.6ka (Pendleton et al., 2017). Subsequent advances through to the LIA are evident, but cirque glaciers in the Brooks Range deposit topographically high moraines, making it difficult for the short, debris-rich glaciers to override them (Pendleton et al., 2017). The lack of LIA moraines may indicate morphological constraints inherent to the glaciers themselves rather than climatic trends, i.e., the lack of LIA maximum extents does not reflect a warming climate (Denton and Karlén, 1977; Pendleton et al., 2017). These pre-LIA Brooks Range moraines are coeval to the complex Tiedemann-Peyto Advance, suggesting a relation to multiple minor pulses of advance between 3.54 – 2.50 ka (Menounos et al., 2013a).

End group 2 moraines in the northeastern Canadian Cordillera represent the most extensive glacial limits throughout the Holocene, emplaced in response to the 0.90 - 0.40 ka regional advance. These moraines likely contain remnants of early Neoglacial advances reflected

by outlier ¹⁰Be ages that have been merged. Specifically, the Itsi Glacier and other end group 2 moraines potentially created a complex, amalgamated landform due to the inability of small alpine glaciers to override topographically high barriers formed by early Neoglacial terminal moraines. This amalgamation mechanism suggests that these glaciers, when advancing during the Holocene maximum extent circa 0.40 ka, encountered pre-existing moraines from earlier neoglacial phases of similar magnitude (Figure 3.4). Instead of completely overriding and resetting the ¹⁰Be boulder inventory, these glaciers added new material to the existing moraine. This process preserved older boulders with inherited ¹⁰Be ages, resulting in the consistent reports of geological scatter.



Figure 3.3: Regional compilation of WNA late Holocene ¹⁰Be moraine ages, including those collected for this thesis, grouped by morphostratigraphic position (Dortch et al., 2010; Hawkins et al., 2023, 2021; Howley, 2008; Menounos et al., 2013a; Pendleton et al., 2017; Young et al.,

2009). Left: map illustrating location of sampled moraines where size of marker is proportional to IQR. Right: scatter plot showing median moraine ages in relation to IQR.

In contrast, end group 1 moraines, culminating during the 0.40 - 0.06 ka regional advance, did so on relatively "clean" surfaces devoid of significant prior moraine deposits. Although the sample size of ¹⁰Be dated moraines in this region is small (n = 26), a discernible pattern emerges (Figure 3.3 and A.2). End group 2 moraines generally exhibit a consistently higher IQR compared to both end group 1 and, in most cases, end group 3. Moreover, the highest IQR values are centered along the YT/NWT border. The lack of inherited boulders from earlier advances resulted in more consistent ¹⁰Be ages and minimal scatter. The morphological constraints and subsequent amalgamation of moraines in the northeastern Canadian Cordillera provide a plausible explanation for the significant age scatter observed in end group 2 moraines.

Amalgamation scenario



Figure 3.4: Conceptual model of moraine amalgamation adapted from Bierman et al. (2021). An amalgamated moraine is an unstratified mixture of clasts, some of which have experienced periods of glacier transport and deposition one or several times before final transport and deposition on the contemporary moraine. At time step t_1 , (i) buried clasts are shielded from cosmic rays, (ii) glacially transported, and (iii) emplaced as a terminal moraine following deglaciation. At t_2 , (iv) buried boulders are, (v) glacially transported, and (vi) emplaced as a recessional moraine proximal to the ice margin. At t_3 , the glacier (vii) overrides the prior moraine deposit from the second glaciation and (viii) emplaces an unstratified mixture of previously deposited clasts as an amalgamated terminal moraine. The resulting exposure age distribution (ix) reflects multiple records of deglaciation.

3.3.4 ¹⁰Be erratic ages

I include additional ages from erratics and bedrock boulders to complement cirque and valley moraine ages for constraining the timeline of CIS decay (Menounos et al., 2017). Itsi Glacier erratics indicate CIS melt exposed alpine glaciers circa 11.61 ka \pm 0.54 (n = 2) among the Itsi Range. This accords with the larger dataset of Canadian Cordillera erratic samples 12.13 ka \pm 1.69 (n = 26) and resembles observations by Hawkins et al. (2023) in the northeastern Canadian Cordillera who recorded erratic deposits ranging from 10.94 ka \pm 0.31 to 12.27 ka \pm 0.22 (n = 4) (Figure 3.5). Together, these erratics date to 11.56 ka \pm 1.55 (Table 3.2). Alpine valley and cirque glaciers advanced coeval with and preceding the Younger Dryas (YD) interval throughout the Canadian Rockies (Hawkins et al., 2023; Luckman and Osborn, 1979; Menounos et al., 2009) which suggests cirque glacier advances correlative to the YD among the Itsi Range were less extensive than their LIA extents. Given that the suite of WNA erratic samples reflect the high spatiotemporal variability of CIS decay over millennia, it does not correspond to a single climatic event; thus, no attempt is made to address scatter in this context.


Figure 3.5: Regional compilation of WNA ¹⁰Be erratic boulder ages, including those collected for this thesis, from BC and YT (Hawkins et al., 2023; Menounos et al., 2017; Stroeven et al., 2010). Thin grey curves show ages of individual boulders rendered as Gaussian distributions. The thick pink curves represent the sum of each Gaussian distribution for the entire population of ¹⁰Be erratic ages. Itsi Glacier erratics are highlighted separately within the summed Gaussian distributions. A corresponding box-and-whisker plot denotes the median \pm IQR, minimum, and maximum surface exposure ages for all erratics.

3.4 Limitations

Different geochronological methods have their own formation processes and variable sensitivity to their environment (Tomkins et al., 2008). I compiled intervals of advance derived from radiocarbon, dendrochronology, and glaciolacustrine sedimentation methods (Barclay et al., 2009; Hawkins et al., 2021; Menounos et al., 2009; Mood and Smith, 2015; Tomkins et al., 2008). As they do not record identical climate signals, evaluating these different methods for timing of glacier advance and moraine emplacement requires consideration of their limitations.

Glaciolacustrine sedimentation records are a proxy for glacial meltwater production, which is dictated by regional climate and glacier dynamics (Tomkins et al., 2008). Glaciers have a lagged response to changes in climate. The magnitude of this lag can vary by tens to hundreds of years between mountain glaciers. Non-climatic factors affecting proglacial sedimentary record formation, like sediment supply, can obscure the glaciogenic clastic signal and consequent interpretation (Menounos et al., 2006; Tomkins et al., 2008). Although sedimentation rates generally increase with glacial ice extent, maximum ice extents do not always align with the highest rates of sedimentation (Leonard, 1997). The relationship between clastic sedimentation rates and glacier extent is more consistent on centennial-to-millennial timescales (Leonard, 1997). On shorter, decadal timescales, the highest sedimentation rates often occur during transitional periods, such as during glacier advances before a maximum or recessions afterward, possibly obscuring the signal of glacier advance (Leonard, 1997).

Dendrochronology records reflect immediate changes in climate. Tree-ring patterns can also record non-climatic factors like fire, disease, and tree physiology, making it difficult to disaggregate the desired signal (Cook and Kairiukstis, 1990; Tomkins et al., 2008).

Radiocarbon dating techniques have two main sources of uncertainty: 1) high uncertainty of radiocarbon ages relative to absolute age within the last millennium and 2) temporal and spatial uncertainties. Temporally, radiocarbon dates can only bracket a given glacier advance, and spatially, they rarely provide precise glacier positions unless dated materials are found *in situ* and still in growth position (Menounos et al., 2009; Mood and Smith, 2015). Climatic signals can still be ascertained among a suite of different chronometer and proxy measurements if they are all similarly influenced by the same climatic factors of sufficient magnitude (Braumann et al., 2020; Tomkins et al., 2008).

I synthesized morphologically relevant ¹⁰Be moraine ages because Balco (2020a) argues that the uncertainty inherent in dating individual glacial landforms is too high to make any correlations with millennial-scale climate drivers. This perspective is further supported by Roe and O'Neal (2009) who provide a case study proving that the stochastic variability of glacier

response to climate forcing can be misattributed to changes in climate. However, we can still attain a synoptic understanding of paleoclimate dynamics using the current global dataset of cosmogenic-nuclide exposure dated alpine glacier landforms (Balco, 2020a).

Correlation remains poorly quantified, i.e., lack of standardized statistical measures, in the field of glacier chronology reconstruction via moraine records, most often only qualitatively estimated through visual comparison of chronostratigraphic columns. Two situations can result in false correlations between glacier response and climate events (Kirkbride and Winkler, 2012):

- 1. Type I ("lumping" errors): Multiple climatic events are correlated as a single event because the glacier response time is less than resolution of the utilized dating methodology.
- Type II ("splitting" errors): A single climatic event is distinguished into multiple events because the timescales of glacier response do not match the chronological resolution, usually caused by high-resolution dating methods.

Inadequately resolved glacier chronologies (i.e., greater dating uncertainty/poor statistical confidence) have an increased likelihood of being "lumped" into regional-to-global correlations because greater dating uncertainty allows moraines, which may have been generated from multiple distinct climatic events, to appear as if they were generated from the same event (Kirkbride and Winkler, 2012). Conversely, highly resolved glacier chronologies are more likely to be "split" into many stages of glacier advance, making correlation difficult at broad spatial scales. A narrow range of dating uncertainty may cause a singular climatic event to be erroneously interpreted as numerous, distinguishable events (Kirkbride and Winkler, 2012).

Given the sparse availability of TCN dated surface exposure ages in the northeastern Canadian Cordillera, I sourced ¹⁰Be ages from across WNA to make assumptions about the

climatic events influencing the Itsi Range. Consequently, the regional ¹⁰Be chronology likely represents Type I "lumping" errors, as the gaussian distributions visualized in Figure 3.2 illustrate millennial-scale levels of 2σ uncertainty for each end group, far greater than the average response time for alpine glaciers. This approach risks over-generalizing multiple climatic events affecting glacial response, resulting in moraine formation, as representing a singular climatic event. The benefit of Type I error lumping is that it simplifies the interpretation of glacial records by reducing the number of discrete events to consider, potentially making it easier to identify broad climatic trends and patterns over time. However, this simplification comes at the cost of potentially overlooking the complexity and variability of past climate events, which could lead to a misleading portrayal of glacial responses and climatic history.

3.5 Conclusion

I compare the glacier chronology at Itsi Glacier with ¹⁰Be moraine ages across WNA, incorporating documented intervals of glacier advances compiled from multiple reviews using radiocarbon dating, dendrochronology, and glaciolacustrine sedimentation records. To facilitate comparisons, I categorize WNA ¹⁰Be moraine ages into four groups based on exposure age and morphostratigraphic position relative to the glacier terminus.

The inner moraine at Itsi Glacier, dated to 0.20 ka \pm 0.05 (n = 3), aligns well with the regional glacier advance period at 0.40 – 0.06 ka. The compiled regional ¹⁰Be moraine record yields an end group 1 age of 0.13 ka \pm 0.10 with a MAD value of 67 years (Table 3.2). When accounting for differences in latitude, northern latitudes show end group 1 moraines dating to 0.30 ka \pm 0.10 (n = 11), while southern latitudes (50.9°) record emplacement at 0.09 ka \pm 0.03 (n = 14). This suggests that end group 1 moraine emplacement occurred decades to centuries earlier at northern latitudes compared to southern latitudes.

The outer moraine at Itsi Glacier, dated to 0.96 ka \pm 0.36 (n = 4), does not align with any singular regional Neoglacial advance. Individually dated boulders accord to multiple advances: 2.09 ka \pm 0.15 (228B04) at 2.20 – 1.80 ka advance, 1.16 ka \pm 0.08 (228B01) at 1.71 – 1.20 ka advance, and 0.43 ka \pm 0.04 (228B05) at 0.90 – 0.40 ka advance. The regional ¹⁰Be moraine record yields an end group 2 age of 0.47 ka \pm 0.35 with a MAD value of 201 years (Table 3.2).

Although all end groups are non-parametric, end group 1 has a significantly lower MAD value, indicating less geological scatter than end group 2. If accepting the proposed mechanism of moraine amalgamation, Itsi Glacier sample 228B05 represents the true minimum-limiting age for maximum LIA extents at 0.43 ka \pm 0.04. According to the regional ¹⁰Be record and new ages from the Itsi Glacier, Holocene maximum extents were reached centuries earlier in the northeastern Canadian Cordillera when compared to alpine glaciers of the western and southwestern Canadian Cordillera. Additional ages from erratic boulders suggest CIS melt exposed alpine glaciers among the Itsi Range around 11.61 ka \pm 0.54 (*n* = 2), consistent with regional observations (Menounos et al., 2017).

Considering the effects of geologic scatter, particularly the roles of inheritance and postdepositional disturbance on Itsi Glacier moraines, the regional moraine record suggests that inheritance on the outer moraine is more likely than ice-cored moraine induced post-depositional disturbance. The most parsimonious explanation for consistent reports of ¹⁰Be geological scatter on outer moraines from LIA moraine pairs in the northeastern Canadian Cordillera is that they likely represent amalgamated moraines, heterogeneous mixtures of previously exposed and reincorporated moraine boulders. Notably, the ¹⁰Be ages of end group 1 moraines throughout the Selwyn and Mackenzie mountains exhibit significantly less scatter compared to the end group 2 moraines (Figure 3.2 and Figure 3.3). Moreover, the anomalous older boulder ages sampled on end group 2 moraines align with pre-LIA moraines dated in AK and BC, centering around the 3.54 – 2.50 ka regional advance (Barclay et al., 2009; Menounos et al., 2013a; Pendleton et al., 2017; Young et al., 2009).

Instead of obliterative overlap presiding, the outermost moraine might have posed a topographical barrier high enough to prevent further glacier advance, leading to multiple advances depositing boulders along the same ridge (Denton and Karlén, 1977; Pendleton et al., 2017). Evidence for this mechanism, beyond intervals of regional advance correlative with boulder ages, is lacking. Boulders exposed during less extensive glacial periods may simply be recycled by the advancing glacier or inherited debris transported supraglacially before deposited as a moraine (Balco, 2020a). More research quantifying the geomorphic uncertainties inherent to TCND, e.g., Tomkins et al. (2021), must be done before declaring whether the geologic scatter is recording "real" events or merely correlations between ages that happen to overlap with known periods of advance.

4. Comparison of moraine record with climate proxy records

4.1 Introduction

Mountain glaciers act as global climate indicators due to their sensitivity to climate forcings (Kaufman and Broadman, 2023). The presence of numerous periodicities (i.e., multiple climate drivers at different frequencies) results in the reinforcement and modulation of climate signals (Kirkbride and Winkler, 2012). The spatial variation of these climate signals may be attributed to the complex interplay of atmospheric and oceanic circulation and teleconnection patterns; therefore, globally synchronous climatic signals, such as a glacier advance, appear when high-frequency cycles reinforce low-frequency cycles (Kirkbride and Winkler, 2012). The Little Ice Age (LIA) represents an example of this, wherein minimal solar output (high-frequency; decadal-to-centennial scale), volcanic forcing (high-frequency; annual-to-decadal scale), declining orbital insolation (low-frequency; 10⁴-year timescale), and pre-industrial decline of atmospheric CO₂ combined to create a globally synchronous climatic signal (Kirkbride and Winkler, 2012; Koch et al., 2019; Owens et al., 2017).

Quasi-periodic teleconnections, such as the Arctic Oscillation (AO), Pacific Decadal Oscillation (PDO), and El Niño-Southern Oscillation (ENSO), demonstrably influence glacier mass balance (Hodge et al., 1998; Moore et al., 2009); however, they are rarely detected in moraine sequences due to the high-frequency (decadal) signal of these quasi-periodic oceanatmospheric circulation patterns (Kirkbride and Winkler, 2012). Because of this, attention is given to low-frequency climate signals at the centennial-to-millennial scale.

4.2 Methodology

Fluctuations of glacier length can only distill an indirect, delayed, and filtered signal of disturbances in mass balance due to the interplay of lag time components: reaction time and response time (Kirkbride and Winkler, 2012; Oerlemans, 1994; Solomina et al., 2015). Jóhannesson et al. (1989) define response time as "the time a glacier takes to adjust to all but a factor of 1/e (37%) net change in mass balance," making it a quantifiable constant for all glaciers (Solomina et al., 2015). The equation for response time is:

$$t_r = H_{MAX}/\dot{a}_0 \tag{14}$$

Where t_r is response time (years), H_{MAX} is mean thickness (m), and \dot{a}_0 is ablation rate near the glacier terminus (m yr⁻¹) (Cuffey and Paterson, 2010). Valley glacier response time generally ranges from 10 – 50 years (Oerlemans, 1994). Reaction time, defined as the time lag between a

change in mass balance and response at the glacier's terminus (Jóhannesson et al., 1989; Solomina et al., 2015), allows for direct observation but cannot be quantified as a constant among a suite of glaciers as it depends on a glacier's history of mass balance change (Solomina et al., 2015).

Response times vary significantly across alpine glaciers within the western North America (WNA) ¹⁰Be chronology, but likely exist on the order of $10^1 - 10^2$ years (Jóhannesson et al., 1989). Using an estimated 27.3 m mean thickness of the Itsi Glacier (Farinotti et al., 2019), and a mean yearly retreat rate of 1.93 m yr⁻¹ from 1754 to 1949 CE estimated using the GTT tool (Table Sx), t_r is estimated to be 14 years. Given that the uncertainties for median ± interquartile range (IQR) end group ages are on the centennial-to-millennial scale, the climate proxy discussions are conducted at a scale where adjustments for t_r is not required.

I qualitatively compare the developed glacier chronology to climate proxy records including volcanic forcing, North Pacific Index (NPI), Total Solar Irradiance (TSI), GHG emissions, orbital forcing, and temperature and precipitation anomalies. These datasets processed using Python (v. 3.11.9) help determine if the proxy records support or refute assertions made by terrestrial *in situ* cosmogenic nuclide (TCN) derived ages.

Volcanic forcing through Stratospheric Sulfur Injections (VSSI) significantly modulated Earth's climate over the Holocene (Kaufman and Broadman, 2023; Sigl et al., 2022; Toohey and Sigl, 2019). VSSI events of sufficient magnitude lead to a reduction of incoming solar insolation at the Earth's surface, contributing to climate cooling at the decadal scale (Porter, 1986; Sigl et al., 2022; Toohey and Sigl, 2019). A correlation coefficient of r = -0.6 between mean annual 2 m air temperature and volcanic forcing in the Northern Hemisphere (NH) further supports this relation (van Dijk et al., 2024). Increased clustering of volcanic eruptions can trigger ice-albedo

feedbacks and reduce ocean heat content, potentially inducing centennial scale cooling; the cooling duration depends on the clustering, pacing, and magnitude of eruptions (Kaufman and Broadman, 2023). Reconstructed volcanic eruption counts from Greenland and Antarctica ice-core sulfate and sulfur records provide a globally resolved Holocene chronology of every volcanic eruption with a mean uncertainty of $\pm 1 - 5$ years over the past 5 ka (Sigl et al., 2022). Eruption counts only include those with a minimum of 5 Tg S injected into the stratosphere (Sigl et al., 2021). The eVolv2k database provides supplementary VSSI values that record a 25% reduction in VSSI over the 20th century as compared to a pre-industrial 2.45 – 0.05 ka mean (Toohey and Sigl, 2019).

The NPI is associated with the Aleutian Low (AL) pressure centre which, a lowfrequency teleconnection pattern known to influence glaciation in British Columbia (BC) and Alaska (AK) (Mood and Smith, 2015; Osterberg et al., 2014). Construction of the NPI proxy involved measuring sodium concentrations by ion chromatography from an ice core collected at the summit of Mt. Logan, YT (60.6° N, -140.5° W) assembled at a 3 year resolution (Osterberg et al., 2014). Lower NPI values, indicating a stronger eastward AL shift, are associated with increased precipitation and cooler temperatures conducive to glacial growth. Higher NPI values, caused by westward AL shifts, bring dry and warm conditions, resulting in glacier retreat (Mood and Smith, 2015; Osterberg et al., 2014).

Solar radiation emitted by the sun, TSI, controls the amount of solar forcing on Earth and is reconstructed at a 5 year resolution by relating open solar magnetic field strength to ¹⁰Be measurements in polar ice cores (Steinhilber et al., 2009; Wanner et al., 2008). TSI has been demonstrated to influence mean surface temperature variations over the last millennium, with prolonged negative TSI periods influencing changes in high-latitude NH glaciers (Gray et al.,

2010; Solomina et al., 2015). The most established mechanism explains that variations in solar UV radiation directly affect the amount of ozone in the stratosphere, leading to temperature changes, which then alter wind patterns and subsequently impact broader climate dynamics (Gray et al., 2010). Amplification mechanisms, such as solar forcing triggering expansion of sea ice, may also explain how minor TSI anomalies lead to glacier advances (Solomina et al., 2015). However, there is no definite consensus concerning the relationship between solar variability and Earth's climate system (Gray et al., 2010).

Köhler et al. (2017) compile annually resolved GHG data from 13 globally distributed ice cores and atmospheric observations. While the CO₂ proxy reflects this global compilation, CH₄ data are collected solely from the SH to account for discrepancies in the timing of major climatic events between hemispheres (Köhler et al., 2017). This CH₄ data is estimated to represent the lower boundary of global CH₄ concentrations (Köhler et al., 2017). As GHG levels rise, they positively contribute to radiative forcing, directly leading to increased global temperatures (Köhler et al., 2017; Tierney et al., 2020). CO₂ is the primary contributor, accounting for twothirds of radiative forcing,

$$\Delta R_{[CO_2]} = 5.35 \cdot \ln\left(\frac{CO_2}{278 \, ppm}\right) \tag{15}$$

while CH₄ plays a smaller, but still significant, role (Köhler et al., 2017).

$$\Delta R_{[CH_4]} \sim 1.4 \cdot 0.036 \cdot \left(\sqrt{CH_4} - \sqrt{742 \ ppb}\right)$$
(16)

Both GHGs are converted to and plotted as radiative forcing (W m⁻²) using equation (15 for CO₂ and equation (16 for CH₄, where CO₂ is measured concentration in parts per million (ppm) and CH₄ is measured concentration in parts per billion (ppb) (Köhler et al., 2017; Myhre et al., 1998).

Orbital variations, or Milankovitch cycles, significantly influence Earth's glacialinterglacial cycles, earning the colloquialism "pacemaker of the ice ages" (Berger, 1988; Marshall, 2011). These cycles, consisting of eccentricity, obliquity, and precession, affect solar insolation distribution and align cyclically on $10^4 – 10^5$ -year scales, leading to cooler or warmer summers at high latitudes (Marshall, 2011; Wanner et al., 2008). NH orbital forcing primarily controls glacial-interglacial cycles by altering solar insolation distribution over the Earth (Berger, 1988; Marshall, 2011). A decrease in NH solar insolation at high latitudes can trigger glacial periods due to ice-albedo feedback loops that result from increased cryospheric albedo (Kaufman and Broadman, 2023). By 1950, summer insolation at 65° N decreased by 37.91 W m⁻² relative to the beginning of the Holocene at 12 ka and continues to steadily decline (Berger and Loutre, 1991).

I obtained temperature and precipitation anomalies from the TraCE-21ka transient climate simulation (Liu et al., 2009), centered on the Itsi Range (61.2° N, -131.2° W) with a temporal resolution of 10 years and a spatial resolution of 3.68° latitude x 3.75° longitude. Both variables use 5 - 3 ka as the reference period for anomaly calculations.

4.3 Results and discussion



Figure 4.1: Neoglacial climate forcings and indicators. (a) Global volcanic eruptions per decade (30-year moving average) and major VSSI event magnitudes (Tg S) (Sigl et al., 2021; Toohey and Sigl, 2019). (b) NPI at Mt. Logan (750-year spline) (Osterberg et al., 2014). (c) TSI

anomalies ($\Delta W m^{-2}$) plotted as 50-year moving averages (light line) and smoothed using a Gaussian filter (dark line) (Steinhilber et al., 2009). (d) A spline smoothed dataset of atmospheric GHGs CO₂ and CH₄ plotted as radiative forcing ($\Delta W m^{-2}$) (Köhler et al., 2017) Increases in GHG concentrations past 0.1 ka are not plotted. (e) July mean Solar Insolation at 65° N (W m⁻²) at a 1000 year resolution (Berger and Loutre, 1991). (f) Itsi Range temperature (ΔT) and precipitation anomalies (ΔP) (Liu et al., 2009) plotted as 50-year moving averages. (g) Regional compilation of late Holocene ¹⁰Be boulder ages, including those collected for this thesis (Dortch et al., 2010; Hawkins et al., 2023, 2021; Howley, 2008). Blue and orange lines represent the sum of each 2 σ Gaussian distribution for end groups 1 and 2, respectively. Itsi Glacier moraine end groups are highlighted separately within the distributions.

Analyzing the terrestrial record of ¹⁰Be ages to regionally relevant climate forcings can help deduce which climate forcings influence glacier expansion and retreat in WNA. Rather than evaluating each proxy individually, I focus on broader patterns at multicentennial-to-millennial scales by analyzing moraine end groups. This approach accounts for the high spatiotemporal variability among proxy types and acknowledges that multiple proxies may collectively influence moraine deposition.

Decreasing solar insolation primarily drove the steadily declining temperature anomalies observed in the Itsi Range, responsible for LIA glacial expansion to Holocene maximum extents in the NH (Clague et al., 2009; Kaufman and Broadman, 2023; Menounos et al., 2009; Wanner et al., 2011). While the onset of Neoglacial glacier growth varies widely across WNA, highfrequency climate forcings can obscure long-term low-frequency trends but still provide valuable insights into regional climate changes (Solomina et al., 2015). Understanding these forcings helps elucidate the mechanisms driving moraine deposition and glacier advances in WNA throughout the late Holocene.

4.3.1 Moraine end groups

Median end group 2 regional ages at 0.47 ka \pm 0.35 (n = 36) variably correspond with the climate proxies, while local Itsi Glacier ages at 0.96 ka \pm 0.36 (n = 4) have weaker correlations

(Figure 4.1). A VSSI spike of 5.33 Tg S at 2.4 ka coincides with a negative TSI anomaly, preceding a period of cooling that began at 2.3 ka and culminated in the 2.2 - 1.80 ka regional advance. Sample 228B04, dated at 2.10 ka \pm 0.15, may reflect this period of cooling if its deposition age is reliable.

While GHG levels remained relatively stable throughout the mid-Holocene, a sustained increase occurred from 1.0 - 0.4 ka, possibly linked to the agricultural practices of pre-industrial civilizations (Ruddiman et al., 2016). This rise led to an increase in radiative forcing by about 0.46 W m⁻² compared to the mean at 6 ka (Kaufman and Broadman, 2023), possibly counteracting natural cooling expected from orbital cycles. However, the impact would likely have been subtle compared to the more pronounced effects of natural climate variability (Roe and O'Neal, 2009).

A TSI minimum influencing glacial advance around 0.5 ka precedes the ages of many regional end group 2 moraines. Although the NPI is loosely reflected in temperature trends overall, an increase from 0.6 - 0.4 ka and the subsequent temperature rise from -0.3°C to -0.1°C appears to have facilitated end group 2 moraine deposition. However, the local Itsi Glacier end group 2 shows limited correlation with all climate proxies.

Median end group 1 regional ages at 0.13 ka \pm 0.10 (n = 25) and local Itsi Glacier ages at 0.20 ka \pm 0.05 (n = 3) show concordance with all assessed climate proxies (Figure 4.1). Volcanic aerosol injections have likely induced global cooling over the last two millennia (Kaufman and Broadman, 2023), with volcanic forcing contributing significantly to the high natural climate variability throughout the LIA (van Dijk et al., 2024). Increased clustering of eruptions starting around 1.3 ka likely amplified orbitally forced Neoglacial cooling, culminating in the LIA, as supported by Holocene transient simulations (Bader et al., 2020; van Dijk et al., 2024).

TSI and NPI trends correlate well with end group 1 moraine deposition. TSI anomalies below zero at 0.7 ka coincide with the onset of the LIA, persisting until 0.2 ka when positive anomalies emerged. This corresponds with the culmination of regional end group 1 ages. Decreasing NPI values from 0.4 ka align with increasing precipitation and decreasing temperature trends, facilitating the deposition of most end group 1 moraines.

From 0.4 - 0.3 ka, global CO₂ emissions gradually dropped by 0.5 ppm per decade following significant land-use changes in the Americas following contact between the New World and the Old World (King et al., 2024; Koch et al., 2019). Anthropogenic GHG emissions, starting around 0.2 ka, may have prematurely ended Neoglaciation as CO₂ and CH₄ levels began to rise (Wanner et al., 2011).

4.3.2 Anthropogenic warming

Although many NH glaciers were nearly as small or smaller than current extents from 9 – 5.5 ka, the observed global retreat of glaciers in the last century does not align with expected orbital forcing patterns expected to encourage glaciation and is primarily attributed to anthropogenic influences (Marzeion et al., 2014; Solomina et al., 2015). Agassiz Ice Cap core data reveal that current melt rates are the highest in 4200 years, comparable to the early Holocene thermal maximum (Fisher et al., 2012). Between 1851 and 2010 CE, approximately 25 \pm 35% of global glacier mass loss is attributed to anthropogenic warming, a figure that rises to 69 \pm 24% from 1991 to 2010 in western Canada (Marzeion et al., 2014).

Arctic amplification, driven by NH sea ice expansion, results in Arctic summer temperature variations 3 – 4 times larger than global changes, explaining the sensitivity of early Holocene and modern Arctic glaciers (Miller et al., 2013, 2010). Century-scale early Holocene advances thought to be caused by cooling from LIS meltwater, which weakened the Atlantic thermohaline circulation, mirror modern concerns with Greenland ice sheet melt (Menounos et al., 2017; Nesje et al., 2004).



Figure 4.2: Projected Itsi Range (62.9°N, -130.2°W,) SSP climate change scenarios (Cannon et al., 2015; ClimateData.ca, 2024; McKenney et al., 2011; Tam et al., 2019).

Shared Socioeconomic Pathways (SSPs) paired with different levels of radiative forcing (W m⁻²), measured as positive deviations from a pre-industrial baseline, assess GHG influence on the energy balance of the climate system (Riahi et al., 2017) (Figure 4.2). COP26 climate pledges indicate an estimated global mean temperature increase of $\pm 2.7^{\circ}$ C, corresponding to SSP2-4.5, which would result in near or total deglaciation of WNA glaciers (95 \pm 8%) by the 2100 CE (Rounce et al., 2023).

4.4 Limitations

While correlating glacier chronologies with climate proxies can validate paleoclimatic patterns, it does not imply causation. The absence of formal quantitative criteria for correlating

local glacier chronologies means that only qualitative conclusions can be drawn (Kirkbride and Winkler, 2012). Consequently, the interpretation of these proxy records in relation to moraine deposition is inherently subjective.

The lack of quantification for the response times of the glaciers represented by each end group introduces additional uncertainty when interpreting their relationship to climate proxies. The assumptions made about the moraines belonging to different end groups directly affect the interpretation of their Gaussian distributions in relation to the proxy records. Different approaches to grouping or analyzing the moraines could lead to alternative interpretations of the same data.

Proxies each have their own inherent limitations. Volcanic eruptions and associated VSSI aerosol spread have high spatiotemporal variability, meaning that it cannot be stated that all eruptions affected NH climate at high latitudes in the northeastern Canadian Cordillera. The NPI's 750-year low-resolution spline simplifies climatic variations at fine temporal scales, offering only a broad approximation helpful for relating low-frequency proxies (Osterberg et al., 2014). The correlation between 2 m air temperature and TSI is weak (r = 0.06), except during major LIA solar minima events at 0.56 – 0.39 ka and 0.33 – 0.23 ka. (van Dijk et al., 2024).

The Mt. Logan ice core record as reported by Osterberg et al. (2014) may be subject to revision. Preliminary results from a 2022 Mt. Logan ice core study (Holland et al., 2023) suggest significantly different age estimates (Holland et al., 2023). However, no formal publication of these findings has been released to date.

Other factors on glaciation in the northeastern Canadian Cordillera were not addressed due to limited proxy records or the complications associated with correlating high-frequency forcings at multicentennial-to-millennial scales. These include internal climate system variability (Roe and O'Neal, 2009) and the PDO known to influence western Canadian glacier behavior (Koch et al., 2009). When considering all these limitations, the qualitative correlations drawn between the glacier chronology and these proxies should be interpreted with caution.

4.5 Conclusion

A reduction of orbital insolation and increased volcanic forcing initiated LIA glaciation (Miller et al., 2012; van Dijk et al., 2024). The regional compilation of late Holocene ¹⁰Be ages shows two distinct peaks in glacial advances during this period, with median \pm IQR ages of WNA moraine deposition at 0.47 ka \pm 0.35 and 0.13 ka \pm 0.10. These peaks show some alignment with the climate proxies in Figure 4.1, indicating a synchronous regional response to certain multicentennial-scale climate forcings, many other peaks lack a coherent counterpart.

Both regional end group ages and Itsi Glacier end group 1 show stronger coherence with the climate proxies, in contrast to Itsi Glacier end group 2 ages which does not correlate well with any proxy beyond orbital forcing. This lack of synchronicity likely indicates significant scatter in the Itsi Glacier end group 2 ages, deviating from the regional ¹⁰Be age signature.

Apart from orbital forcing, proxies that most closely align with the ¹⁰Be dated end group 1 ages – and likely exert the most influence on glacial advance and retreat – include temperature and precipitation anomalies, GHG emissions, and TSI anomalies. Proxies with a less direct impact on end group 1 moraine deposition are NPI and volcanic forcing. For regional end group 2, TSI anomalies show the strongest correspondence, although temperature anomalies, GHG emissions, and NPI display plausible but less direct correlations when accounting for uncertainties.

The Mt. Logan NPI values suggest shifts in the AL significantly influenced glacial events. A westward shift of the AL likely drove glacial retreat after the deposition of end group 2. A subsequent eastward shift after end group 2's deposition contributed to another, though less extensive, glacial advance that culminated in end group 1 deposition. The rise in GHG emissions appears to have prematurely ended the advances of end group 1, despite NPI values, orbital trends, and temperature and precipitation data indicating conditions favorable for glacier growth. This suggests that anthropogenic influences have increasingly disrupted the natural drivers of glacier advance and retreat in this region.

The timing of end group 1 and 2 moraine deposition aligns with periods of negative TSI anomalies, stronger AL influence, and reduced summer insolation, supporting the possibility that these proxies collectively drove moraine deposition in the northeastern Canadian Cordillera.

5. Conclusions

5.1 Summary of advances

This thesis presents a latest Pleistocene and Holocene chronology for the Itsi Glacier in the Selwyn Mountains, YT using ¹⁰Be surface exposure dating of moraine boulders to determine onset of glacial retreat. This research marks the first application of cosmogenic nuclide dating on alpine cirque glacier moraines within the Itsi Range, YT. Erratics immediately beyond the Itsi Glacier terminal moraine date to 11.61 ka \pm 0.54 (n = 2) and indicate that the CIS was no more extensive than the erratics' position at this time. The Itsi Glacier reached its maximum Holocene extent at 0.96 ka \pm 0.36 (n = 4) as recorded by the outer moraine. Considering the proposed method of moraine amalgamation, I interpret Itsi Glacier sample 228B05 as the true minimumlimiting age at 0.43 ka \pm 0.04 as related to the regional advance between 0.90 – 0.40 ka and aligning with the timing of other maximum Holocene extents in the Selwyn and Mackenzie Mountains as concluded by Hawkins et al. (2023). These findings suggest that LIA maximum extents were reached earlier (~0.40 ka) in the northeastern Canadian Cordillera than the 19thcentury maxima recorded in the western and southwestern Cordillera (Menounos et al., 2009; Mood and Smith, 2015). Following the Holocene maximum, the Itsi Glacier experienced a minor re-advance at 0.20 ka \pm 0.05 (n = 3), as recorded by the inner moraine. This date aligns with the regional compilation of Neoglacial advance intervals in WNA from 0.40 – 0.06 ka discussed in Chapter 3.3.2.

Comparing end group 1 and end group 2 moraines in the Selwyn and Mackenzie Mountains reveals patterns of geologic scatter. All end group 1 moraines exhibit significantly less age scatter, in terms of NMAD and X^2 , and stronger correlations with regional climate proxies, suggesting a coherent glacial signal. Latest Holocene advances are primarily driven by orbital forcing, with other key influences including temperature and precipitation anomalies, GHG emissions, and TSI fluctuations.

In contrast, all end group 2 moraines show low concordance with climate proxies. This lack of synchronicity likely reflects significant geologic scatter, deviating from the expected ¹⁰Be age signature for the region. Anomalously old ¹⁰Be ages from end group 2 moraines across the Selwyn and Mackenzie Mountains (2.29 ka \pm 0.83, n = 10) appear correlative with end group 3 moraines at northern latitudes in Alaska (2.33 ka \pm 0.63, n = 20) (Pendleton et al., 2017; Young et al., 2009) and an end group 3 moraine at Tiedemann Glacier in British Columbia (2.60 \pm 0.28, n = 6) (Menounos et al., 2013a). These end group 3 moraines are coeval to the Tiedemann-Peyto Advance between 3.54 – 2.50 ka. The consistent geologic scatter in end group 2 likely reflects

amalgamated moraines, heterogeneous mixtures of previously exposed and reincorporated moraine boulders, deposited during the Tiedemann-Peyto Advance.

Photogrammetric analysis of the Itsi Range glaciers reveals an accelerating mass loss. Between 1949 and 2004, glaciers in the region experienced a geodetic mass balance loss of -0.190 ± 0.065 m w.e. yr⁻¹. This rate more than doubled to -0.382 ± 0.136 m w.e. yr⁻¹ during the period 2004 to 2022, representing a 101% increase in mass loss.

5.2 Research limitations

A key challenge when conducting TCND of alpine glaciers is managing the methodological issue of geologic scatter. While the Itsi Glacier's inner moraine and outboard erratics exhibit minimal scatter of ¹⁰Be ages, the outer moraine typifies this recurring problem. Of the four boulders sampled from the outer moraine, two yielded ages from the past millennium, while the other two indicated 1 - 2 ka of surface exposure. Presence of geologic scatter complicates the attribution of a moraine to a specific glacial advance, as the true age is obscured by these outliers.

The high cost of remote sampling and laboratory processing of boulders for ¹⁰Be concentration limits the ability to thoroughly assess late Holocene fluctuations at the Itsi Glacier. Collecting three to five samples per moraine is common given the expense, but this sample size often proves insufficient when outliers are present, as seen with the Itsi Glacier's outer moraine, reducing the reliability of the dataset for robust conclusions.

Using median \pm IQR is advantageous because it does not assume normality of data and is less sensitive to outliers, such as those found in moraines with scattered boulder ages. However, this approach can result in less certainty about dispersion/variance compared to measures like

mean \pm standard deviation, leading to broader error margins. These larger error bounds reduce the chronological resolution of moraine ages, increasing the risk of overinterpretation due to the increased apparent similarity of moraine ages; a challenge specifically when attributing the Itsi Glacier outer moraine to a regional advance as discussed in Chapter 3.

The geomorphic uncertainties inherent to TCND limit the ability to confidently attribute geologic scatter to moraine amalgamation. While the alignment of anomalously old boulder ages with pre-LIA advances suggests amalgamation as a possible mechanism, other factors, namely inheritance or post-depositional disturbance, cannot be ruled out. Without further research quantifying these uncertainties, it remains difficult to assert whether moraine amalgamation is the definitive cause of the scatter.

The Chapter 4 discussion of moraine ages in relation to climate forcings is also vulnerable to overinterpretation. This is because it relies on subjective visual correlations among climate proxies, many of which have large uncertainties. As a result, interpretation of climate forcings as related to moraine ages may only suggest broad, correlated trends, rather than definitive evidence for causation of specific glacial advances.

5.3 Research recommendations

1. Utilize tandem dating methods for geochronological research. Both ¹⁴C and ¹⁰Be can be extracted from quartz, enabling paired ¹⁴C-¹⁰Be analysis a way to build a more robust chronology (Lamp et al., 2019). This technique leverages differences in half-life decay and nuclide production to detect disequilibrium in exposure ages from the LGM to the late Holocene (Hippe, 2017; Schimmelpfennig et al., 2022). By identifying inheritance or exhumation processes that contribute to geologic scatter on moraines, paired dating

improves the reliability of exposure ages (Goehring et al., 2022; Stroeven et al., 2014). A more cost-effective option is utilizing organic material when available. Radiocarbon dating from glacier forefields can complement ¹⁰Be dating, improving the precision of glacial chronologies and constraining the duration of Holocene advances at individual sites (Hawkins et al., 2021). However, obtaining organic ¹⁴C in glacier forefields is often not possible because many sites at far northern latitudes, including the glaciers of the Itsi Range, lack the growing seasons necessary to support organic growth such as trees.

2. Strengthen understanding of geomorphic context in relation to moraine boulder

exposure ages. While sampling criteria were carefully considered in this study, uncertainties remain regarding best practices for sampling moraines as indicated by emerging research from Tomkins et al. (2021). For latest Holocene research that targets younger moraines (<1 ka), boulder sampling should avoid moraine crests and instead target more stable intra-landform locations on ice-proximal and ice-distal slopes (Tomkins et al., 2021). These preliminary conclusions about sampling strategies need further quantitative testing, with the framework of tandem TCND with Schmidt hammer sampling proposed by Tomkins et al. (2021, 2018) a viable path forward.

3. Further research mechanisms for moraine amalgamation. Although uncertainties for measuring concentrations ¹⁰Be have steadily decreased over the past thirty years, observations of geologic scatter remain reflecting geomorphic uncertainties in the field (Schaefer et al., 2022). Developing methods to distinguish amalgamated moraines from those representing single glacier advances is crucial for resolving persistent methodological issues of geologic scatter. One approach could compare X² or NMAD values across a large dataset of ¹⁰Be-dated moraines grouped by morphostratigraphic

position. Another strategy involves resampling previously dated moraines with signs of geologic scatter and correlating these to well-constrained local advances to qualitatively assess processes of moraine amalgamation in a more controlled manner.





Figure A.1: Regional compilation of WNA late Holocene ¹⁰Be boulder ages, including those collected for this thesis, sampled from moraines (Dortch et al., 2010; Hawkins et al., 2023, 2021; Howley, 2008; Menounos et al., 2013a; Pendleton et al., 2017; Young et al., 2009). Thin grey curves show ages of individual boulders rendered as Gaussian distributions. The thick curve represents the sum of each Gaussian distribution for the entire population of ¹⁰Be moraine end groups 1 - 3 ages. Itsi Glacier moraine end groups 1 and 2 are inclusive in the orange curve and also highlighted separately within the summed Gaussian distributions. Corresponding box-and-whisker plots denote the median \pm IQR, minimum, and maximum surface exposure ages of each moraine group.



Figure A.2: Regional compilation of WNA late Holocene scatter plots showing ¹⁰Be moraine ages, including those collected for this thesis, grouped by morphostratigraphic position (Dortch et al., 2010; Hawkins et al., 2023, 2021; Howley, 2008; Menounos et al., 2013a; Pendleton et al., 2017; Young et al., 2009). Y-axis displays IQR (ka) and x-axis displays n, latitude, and longitude.



Figure A.3: Output from the GTT toolbox showing line connections between traced Itsi Range glacier termini and corresponding moraines used in retreat rate calculations.



Figure A.4: Compilation of all sampled boulders with their associated sample number. Photos taken by Gerald Osborn.



Figure A.5: Compilation of Itsi Glacier moraine field photos. Red arrows indicate true north. Photos taken by Brian Menounos.





Table A.1: Intervals of major WNA advances discussed in chapter 3 and illustrated in figure 3.1.

Advance	Locations ^a	Evidence ^b	References					
(ka BP)								
Early-middle Neoglacial glacier advances (4.40 – 3.40 ka)								
4.40 - 3.97	CM, WCC	RC	(Menounos et al., 2009; Mood and Smith, 2015)					
3.70 - 3.40	СМ	RC	(Mood and Smith, 2015)					
Middle-late Neoglacial glacier advances (3.54 – 1.20 ka)								
3.54 - 2.50	AK, CM,	D, RC	(Barclay et al., 2009; Menounos et al., 2009;					
	WCC		Mood and Smith, 2015)					
2.20 - 1.80	AK, CM	D, RC, ¹⁰ Be	(Barclay et al., 2009; Hawkins et al., 2021)					
1.71 - 1.20	AK, CM,	D, RC, ¹⁰ Be	(Barclay et al., 2009; Menounos et al., 2009;					
	WCC		Mood and Smith, 2015)					
Latest Neoglacial glacier advances (0.90 – 0.06 ka)								
0.90 - 0.40	AK, CM, SM,	D, RC, GL,	(Barclay et al., 2009; Hawkins et al., 2021;					
	SN, WCC	10 Be	Menounos et al., 2009; Mood and Smith, 2015;					
			Solomina et al., 2015; Tomkins et al., 2008)					
0.50 - 0.24	AK, SM,	D, RC, GL	(Barclay et al., 2009; Menounos et al., 2009;					
	WCC		Tomkins et al., 2008)					
0.40 - 0.06	AK, CM, SM,	D, RC, GL,	(Barclay et al., 2009; Hawkins et al., 2021;					
	SN, WCC	10 Be	Menounos et al., 2009; Mood and Smith, 2015;					
			Solomina et al., 2015; Tomkins et al., 2008)					

^a Locations:

AK = Alaska, USA

CM = Coast Mountains, British Columbia, Canada

SM = Selwyn Mountains, Yukon Territory, Canada

WCC = Western Canadian Cordillera (encompassing Yukon Territory, Northwest Territories, British Columbia, and Alberta)

SN = Sierra Nevada Mountains, USA

^b Evidence:

D = Dendrochronology

RC = Radiocarbon dating

 ${}^{10}\text{Be} = {}^{10}\text{Be}$ terrestrial *in situ* cosmogenic nuclide dating

GL = Glaciolacustrine sedimentation records

Table A.2: Change in area among all Itsi Range glaciers observed from historical orthophotos (1949 and 2002) and satellite imagery (2022).

		Total area (km ²)								
Year	Glacier count	Max	Median	Min	Standard deviation	Average	Sum	ΔArea %		
1949	18	2.89	0.97	0.09	0.78	1.08	19.51	-		
2004	22	2.56	0.42	0.06	0.66	0.66	14.44	-26%		
2022	25	1.99	0.22	0.03	0.53	0.45	11.17	-23%		

References

- Bader, J., Jungclaus, J., Krivova, N., Lorenz, S., Maycock, A., Raddatz, T., Schmidt, H., Toohey, M., Wu, C.-J., Claussen, M., 2020. Global temperature modes shed light on the Holocene temperature conundrum. Nat. Commun. 11, 4726.
- Balco, G., 2020a. Glacier change and paleoclimate applications of cosmogenic-nuclide exposure dating. Annu. Rev. Earth Planet. Sci. 48, 21–48.
- Balco, G., 2020b. A prototype transparent-middle-layer data management and analysis infrastructure for cosmogenic-nuclide exposure dating. Geochronology 2, 169–175.
- Balco, G., 2011. Contributions and unrealized potential contributions of cosmogenic-nuclide exposure dating to glacier chronology, 1990–2010. Quat. Sci. Rev. 30, 3–27.
- Barclay, D.J., Wiles, G.C., Calkin, P.E., 2009. Holocene glacier fluctuations in Alaska. Quat. Sci. Rev. 28, 2034–2048.
- Barr, I.D., Lovell, H., 2014. A review of topographic controls on moraine distribution. Geomorphology 226, 44–64.
- Beedle, M.J., Menounos, B., Wheate, R., 2014. An evaluation of mass-balance methods applied to Castle Creek Glacier, British Columbia, Canada. J. Glaciol. 60, 262–276.
- Berger, A., 1988. Milankovitch theory and climate. Rev. Geophys. 26, 624–657.
- Berger, A., Loutre, M.-F., 1991. Insolation values for the climate of the last 10 million years. Quat. Sci. Rev. 10, 297–317.
- Berger, A.L., 1978. Long-Term Variations of Caloric Insolation Resulting from the Earth's Orbital Elements1. Quat. Res. 9, 139–167.
- Bierman, P.R., Bender, A.M., Christ, A.J., Corbett, L.B., Halsted, C.T., Portenga, E.W., Schmidt, A.H., 2021. Dating by cosmogenic nuclides, in: Encyclopedia of Geology. Academic Press Oxford, pp. 101–115.
- Bonsal, B., Shrestha, R.R., Dibike, Y., Peters, D.L., Spence, C., Mudryk, L., Yang, D., 2020. Western Canadian freshwater availability: Current and future vulnerabilities. Environ. Rev. 28, 528–545.
- Borchers, B., Marrero, S., Balco, G., Caffee, M., Goehring, B., Lifton, N., Nishiizumi, K., Phillips, F., Schaefer, J., Stone, J., 2016. Geological calibration of spallation production rates in the CRONUS-Earth project. Quat. Geochronol. 31, 188–198.
- Braumann, S.M., Schaefer, J.M., Neuhuber, S.M., Lüthgens, C., Hidy, A.J., Fiebig, M., 2021. Early Holocene cold snaps and their expression in the moraine record of the eastern European Alps. Clim. Past 17, 2451–2479.
- Braumann, S.M., Schaefer, J.M., Neuhuber, S.M., Reitner, J.M., Lüthgens, C., Fiebig, M., 2020. Holocene glacier change in the Silvretta Massif (Austrian Alps) constrained by a new 10Be chronology, historical records and modern observations. Quat. Sci. Rev. 245.
- Broecker, W.S., Denton, G.H., Edwards, R.L., Cheng, H., Alley, R.B., Putnam, A.E., 2010. Putting the Younger Dryas cold event into context. Quat. Sci. Rev. 29, 1078–1081.
- Brown, L., Klein, J., Middleton, R., Sacks, I.S., Tera, F., 1982. 10 Be in island-arc volcanoes and implications for subduction. Nature 299, 718–720.
- Cannon, A.J., Sobie, S.R., Murdock, T.Q., 2015. Bias correction of GCM precipitation by quantile mapping: how well do methods preserve changes in quantiles and extremes? J. Clim. 28, 6938–6959.

- Carlson, A.E., LeGrande, A.N., Oppo, D.W., Came, R.E., Schmidt, G.A., Anslow, F.S., Licciardi, J.M., Obbink, E.A., 2008. Rapid early Holocene deglaciation of the Laurentide ice sheet. Nat. Geosci. 1, 620–624.
- Clague, J.J., Menounos, B., Osborn, G., Luckman, B.H., Koch, J., 2009. Nomenclature and resolution in Holocene glacial chronologies. Quat. Sci. Rev. 28, 2231–2238.
- ClimateData.ca, 2024.
- Cogley, J., Hock, R., Rasmussen, L., Arendt, A., Bauder, A., Braithwaite, R., Jansson, P., Kaser, G., Möller, M., Nicholson, L., Zemp, M., 2011. Glossary of glacier mass balance and related terms. https://doi.org/10.5167/uzh-53475
- Colpron, M., 2022. Yukon Bedrock Geology Map.
- contributors, xdem, 2024. xdem. https://doi.org/10.5281/zenodo.11204531
- Cook, E.R., Kairiukstis, L.A., 1990. Methods of dendrochronology: applications in the environmental sciences. Kluwer Academic Publishers.
- Crump, S.E., Anderson, L.S., Miller, G.H., Anderson, R.S., 2017. Interpreting exposure ages from ice-cored moraines: a Neoglacial case study on Baffin Island, Arctic Canada. J. Quat. Sci. 32, 1049–1062.
- Cuffey, K.M., Paterson, W.S.B., 2010. The physics of glaciers. Academic Press.
- Dalton, A.S., Dulfer, H.E., Margold, M., Heyman, J., Clague, J.J., Froese, D.G., Gauthier, M.S., Hughes, A.L., Jennings, C.E., Norris, S.L., 2023. Deglaciation of the north American ice sheet complex in calendar years based on a comprehensive database of chronological data: NADI-1. Quat. Sci. Rev. 321, 108345.
- Darvill, C., Menounos, B., Goehring, B., Lesnek, A., 2022. Cordilleran Ice Sheet Stability During the Last Deglaciation. Geophys. Res. Lett. 49, e2021GL097191.
- Darvill, C., Menounos, B., Goehring, B., Lian, O.B., Caffee, M., 2018. Retreat of the western Cordilleran Ice Sheet margin during the last deglaciation. Geophys. Res. Lett. 45, 9710– 9720.
- Darvill, C.M., 2013. Cosmogenic nuclide analysis. Geomorphol. Tech. Br. Soc. Geomorphol. Lond. 1–25.
- Davis Jr, R., Schaeffer, O.A., 1955. Chlorine-36 in nature. Ann. N. Y. Acad. Sci. 62, 107–121.
- Denton, G.H., Karlén, W., 1977. Holocene glacial and tree-line variations in the White River Valley and Skolai Pass, Alaska and Yukon Territory. Quat. Res. 7, 63–111.
- Dommenget, D., 2009. The ocean's role in continental climate variability and change. J. Clim. 22, 4939–4952.
- Dortch, J.M., Owen, L.A., Caffee, M.W., Brease, P., 2010. Late Quaternary glaciation and equilibrium line altitude variations of the McKinley River region, central Alaska Range. Boreas 39, 233–246.
- Erb, M.P., McKay, N.P., Steiger, N., Dee, S., Hancock, C., Ivanovic, R.F., Gregoire, L.J., Valdes, P., 2022. Reconstructing Holocene temperatures in time and space using paleoclimate data assimilation. Clim Past 18, 2599–2629. https://doi.org/10.5194/cp-18-2599-2022
- Evans, D.J., 2009. Controlled moraines: origins, characteristics and palaeoglaciological implications. Quat. Sci. Rev. 28, 183–208.
- Farinotti, D., Huss, M., Fürst, J.J., Landmann, J., Machguth, H., Maussion, F., Pandit, A., 2019. A consensus estimate for the ice thickness distribution of all glaciers on Earth. Nat. Geosci. 12, 168–173.

- Fick, S.E., Hijmans, R.J., 2017. WorldClim 2: new 1-km spatial resolution climate surfaces for global land areas. Int. J. Climatol. 37, 4302–4315.
- Fisher, D., Zheng, J., Burgess, D., Zdanowicz, C., Kinnard, C., Sharp, M., Bourgeois, J., 2012. Recent melt rates of Canadian arctic ice caps are the highest in four millennia. Glob. Planet. Change 84, 3–7.
- Gibbons, A.B., Megeath, J.D., Pierce, K.L., 1984. Probability of moraine survival in a succession of glacial advances. Geology 12, 327–330.
- Girod, L., Nuth, C., Kääb, A., McNabb, R., Galland, O., 2017. MMASTER: improved ASTER DEMs for elevation change monitoring. Remote Sens. 9, 704.
- Goehring, B., Menounos, B., Osbron, G., Hawkins, A., Ward, B., 2022. Reconciling the apparent absence of a Last Glacial Maximum alpine glacial advance, Yukon Territory, Canada, through cosmogenic beryllium-10 and carbon-14 measurements. Geochronology 4, 311–322.
- Gordey, S.P., 2013. Evolution of the Selwyn Basin region, Sheldon Lake and Tay River map areas, central Yukon. Geol. Surv. Can. Bulletin 599, 176. https://doi.org/10.4095/293034
- Gosse, J.C., Phillips, F.M., 2001. Terrestrial in situ cosmogenic nuclides: theory and application. Quat. Sci. Rev. 20, 1475–1560.
- Gray, L.J., Beer, J., Geller, M., Haigh, J.D., Lockwood, M., Matthes, K., Cubasch, U., Fleitmann, D., Harrison, G., Hood, L., 2010. Solar influences on climate. Rev. Geophys. 48.
- Groisman, P.Y., Karl, T.R., Knight, R.W., 1994. Observed impact of snow cover on the heat balance and the rise of continental spring temperatures. Science 263, 198–200.
- Hawkins, A.C., 2023. Multiple geochronological approaches to constrain late Quaternary glacial histories in western Canada.
- Hawkins, A.C., Menounos, B., Goehring, B.M., Osborn, G., Pelto, B.M., Darvill, C.M., Schaefer, J.M., 2023. Late Holocene glacier and climate fluctuations in the Mackenzie and Selwyn mountain ranges, northwestern Canada. The Cryosphere 17, 4381–4397.
- Hawkins, A.C., Menounos, B., Goehring, B.M., Osborn, G.D., Clague, J.J., Jensen, B., 2021. Tandem dating methods constrain late Holocene glacier advances, southern Coast Mountains, British Columbia. Quat. Sci. Rev. 274, 107282.
- Heisinger, B., Lal, D., Jull, A., Kubik, P., Ivy-Ochs, S., Neumaier, S., Knie, K., Lazarev, V., Nolte, E., 2002. Production of selected cosmogenic radionuclides by muons: 1. Fast muons. Earth Planet. Sci. Lett. 200, 345–355.
- Heyman, J., Stroeven, A.P., Harbor, J.M., Caffee, M.W., 2011. Too young or too old: evaluating cosmogenic exposure dating based on an analysis of compiled boulder exposure ages. Earth Planet. Sci. Lett. 302, 71–80.
- Hippe, K., 2017. Constraining processes of landscape change with combined in situ cosmogenic 14C-10Be analysis. Quat. Sci. Rev. 173, 1–19.
- Hodge, S.M., Trabant, D.C., Krimmel, R.M., Heinrichs, T.A., March, R.S., Josberger, E.G., 1998. Climate variations and changes in mass of three glaciers in western North America. J. Clim. 11, 2161–2179.
- Holland, K.M., Criscitiello, A.S., McConnell, J., Winski, D., Markle, B., Campbell, S.W., Kreutz, K.J., Chellman, N., Wensman, S., McCrimmon, D., 2023. Age scale and preliminary findings of the 2022 Mt. Logan ice core, southwest Yukon. Presented at the AGU Fall Meeting Abstracts, pp. C43A-07.

- Howley, M.W., 2008. A late glacial and Holocene chronology of the Castner Glacier, Delta River valley, Alaska. University of New Hampshire.
- Hugonnet, R., Brun, F., Berthier, E., Dehecq, A., Mannerfelt, E.S., Eckert, N., Farinotti, D., 2022. Uncertainty analysis of digital elevation models by spatial inference from stable terrain. IEEE J. Sel. Top. Appl. Earth Obs. Remote Sens. 15, 6456–6472.
- Hugonnet, R., McNabb, R., Berthier, E., Menounos, B., Nuth, C., Girod, L., Farinotti, D., Huss, M., Dussaillant, I., Brun, F., 2021. Accelerated global glacier mass loss in the early twenty-first century. Nature 592, 726–731.
- Huss, M., 2013. Density assumptions for converting geodetic glacier volume change to mass change. The Cryosphere 7, 877–887.
- Hyndman, R., Cassidy, J., Adams, J., Rogers, G., Mazzotti, S., 2005. Earthquakes and seismic hazard in the Yukon-Beaufort-Mackenzie. CSEG Rec. 30, 32–66.
- IPCC, 2022. In: H.-O. Pörtner, D.C. Roberts, M. Tignor, E.S. Poloczanska, K. Mintenbeck, A. Alegría, M. Craig, S. Langsdorf, S. Löschke, V. Möller, A. Okem, B. Rama (eds.), Climate Change 2022: Impacts, Adaptation, and Vulnerability. Contribution of Working Group II to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change. Camb. Univ. Press Press.
- IPCC, 2021. In: Masson-Delmotte, V., P. Zhai, A. Pirani, S.L. Connors, C. Péan, S. Berger, N. Caud, Y. Chen, L. Goldfarb, M.I. Gomis, M. Huang, K. Leitzell, E. Lonnoy, J.B.R. Matthews, T.K. Maycock, T. Waterfield, O. Yelekçi, R. Yu, and B. Zhou (Eds.), Climate Change 2021: The Physical Science Basis. Contribution of Working Group I to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change. Camb. Univ. Press Press.
- IPCC, 2013. In: Stocker, T.F., D. Qin, G.-K. Plattner, M. Tignor, S.K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex and P.M. Midgley (Eds.), Climate Change 2013: The Physical Science Basis: Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change. Camb. Univ. Press Camb. U. K. N. Y. NY USA.
- Jackson, L., Isobe, J., 1990. Rock avalanches in the Pelly Mountains, Yukon Territory. Geol. Surv. Can. Pap. 263–269.
- Jackson, L.E., Jr., Morison, S.R., McKenna, K., 1993. Surficial geology, Prevost River, Yukon Territory. Map 1833A.
- Jacoby, G., Cook, E., 1981. Past temperature variations inferred from a 400-year tree-ring chronology from Yukon Territory, Canada. Arct. Alp. Res. 13, 409–418.
- Jóhannesson, T., Raymond, C., Waddington, E., 1989. Time–scale for adjustment of glaciers to changes in mass balance. J. Glaciol. 35, 355–369.
- Kaplan, M., Schaefer, J., Strelin, J.A., Denton, G., Anderson, R., Vandergoes, M., Finkel, R., Schwartz, R., Travis, S., Garcia, J., 2016. Patagonian and southern South Atlantic view of Holocene climate. Quat. Sci. Rev. 141, 112–125.
- Kaufman, D.S., Broadman, E., 2023. Revisiting the Holocene global temperature conundrum. Nature 614, 425–435.
- King, A.C., Bauska, T.K., Brook, E.J., Kalk, M., Nehrbass-Ahles, C., Wolff, E.W., Strawson, I., Rhodes, R.H., Osman, M.B., 2024. Reconciling ice core CO2 and land-use change following New World-Old World contact. Nat. Commun. 15, 1735.
- Kirkbride, M., Winkler, S., 2012. Correlation of Late Quaternary moraines: impact of climate variability, glacier response, and chronological resolution. Quat. Sci. Rev. 46, 1–29.

- Koch, A., Brierley, C., Maslin, M.M., Lewis, S.L., 2019. Earth system impacts of the European arrival and Great Dying in the Americas after 1492. Quat. Sci. Rev. 207, 13–36.
- Koch, J., Menounos, B., Clague, J.J., 2009. Glacier change in Garibaldi Provincial Park, southern Coast Mountains, British Columbia, since the Little Ice Age. Glob. Planet. Change 66, 161–178.
- Köhler, P., Nehrbass-Ahles, C., Schmitt, J., Stocker, T.F., Fischer, H., 2017. A 156 kyr smoothed history of the atmospheric greenhouse gases CO 2, CH 4, and N 2 O and their radiative forcing. Earth Syst. Sci. Data 9, 363–387.
- Kokelj, S.V., Lantz, T.C., Tunnicliffe, J., Segal, R., Lacelle, D., 2017. Climate-driven thaw of permafrost preserved glacial landscapes, northwestern Canada. Geology 45, 371–374.
- Lakeman, T.R., Clague, J.J., Menounos, B., 2008. Advance of alpine glaciers during final retreat of the Cordilleran ice sheet in the Finlay River area, northern British Columbia, Canada. Quat. Res. 69, 188–200.
- Lal, D., 1991. Cosmic ray labeling of erosion surfaces: in situ nuclide production rates and erosion models. Earth Planet. Sci. Lett. 104, 424–439.
- Lal, D., Arnold, J.R., Honda, M., 1960. Cosmic-ray production rates of Be 7 in oxygen, and P 32, P 33, S 35 in argon at mountain altitudes. Phys. Rev. 118, 1626.
- Lamp, J.L., Young, N.E., Koffman, T., Schimmelpfennig, I., Tuna, T., Bard, E., Schaefer, J.M., 2019. Update on the cosmogenic in situ 14C laboratory at the Lamont-Doherty Earth Observatory. Nucl. Instrum. Methods Phys. Res. Sect. B Beam Interact. Mater. At. 456, 157–162.
- LDEO, 2012a. Separation and Purification of Quartz from Whole Rock.
- LDEO, 2012b. Extraction of Beryllium from Quartz.
- Leonard, E.M., 1997. The relationship between glacial activity and sediment production: evidence from a 4450-year varve record of neoglacial sedimentation in Hector Lake, Alberta, Canada. J. Paleolimnol. 17, 319–330.
- Leys, C., Ley, C., Klein, O., Bernard, P., Licata, L., 2013. Detecting outliers: Do not use standard deviation around the mean, use absolute deviation around the median. J. Exp. Soc. Psychol. 49, 764–766.
- Lifton, N., 2016. Implications of two Holocene time-dependent geomagnetic models for cosmogenic nuclide production rate scaling. Earth Planet. Sci. Lett. 433, 257–268.
- Lifton, N., Sato, T., Dunai, T.J., 2014. Scaling in situ cosmogenic nuclide production rates using analytical approximations to atmospheric cosmic-ray fluxes. Earth Planet. Sci. Lett. 386, 149–160.
- Liu, Z., Otto-Bliesner, B., He, F., Brady, E., Tomas, R., Clark, P., Carlson, A., Lynch-Stieglitz, J., Curry, W., Brook, E., 2009. Transient simulation of last deglaciation with a new mechanism for Bølling-Allerød warming. science 325, 310–314.
- Luckman, B., Osborn, G., 1979. Holocene glacier fluctuations in the middle Canadian Rocky Mountains. Quat. Res. 11, 52–77.
- Luckman, B.H., 2000. The little ice age in the Canadian Rockies. Geomorphology 32, 357–384.
- Luckman, B.H., Holdsworth, G., Osborn, G.D., 1993. Neoglacial glacier fluctuations in the Canadian Rockies. Quat. Res. 39, 144–153.
- Lukas, S., 2011. Ice-cored moraines. Encycl. Snow Ice Glaciers 616-619.
- Marshall, S.J., 2011. The cryosphere. Princeton University Press.

- Marshall, S.J., White, E.C., Demuth, M.N., Bolch, T., Wheate, R., Menounos, B., Beedle, M.J., Shea, J.M., 2011. Glacier water resources on the eastern slopes of the Canadian Rocky Mountains. Can. Water Resour. J. 36, 109–134.
- Marzeion, B., Cogley, J.G., Richter, K., Parkes, D., 2014. Attribution of global glacier mass loss to anthropogenic and natural causes. Science 345, 919–921.
- Massey Jr, F.J., 1951. The Kolmogorov-Smirnov test for goodness of fit. J. Am. Stat. Assoc. 46, 68–78.
- Maurer, M.K., Menounos, B., Luckman, B.H., Osborn, G., Clague, J.J., Beedle, M.J., Smith, R., Atkinson, N., 2012. Late Holocene glacier expansion in the Cariboo and northern Rocky Mountains, British Columbia, Canada. Quat. Sci. Rev. 51, 71–80.
- McKenney, D.W., Hutchinson, M.F., Papadopol, P., Lawrence, K., Pedlar, J., Campbell, K., Milewska, E., Hopkinson, R.F., Price, D., Owen, T., 2011. Customized spatial climate models for North America. Bull. Am. Meteorol. Soc. 92, 1611–1622.
- Menounos, B., Clague, J.J., Clarke, G.K., Marcott, S.A., Osborn, G., Clark, P.U., Tennant, C., Novak, A.M., 2013a. Did rock avalanche deposits modulate the late Holocene advance of Tiedemann Glacier, southern Coast Mountains, British Columbia, Canada? Earth Planet. Sci. Lett. 384, 154–164.
- Menounos, B., Clague, J.J., Osborn, G., Davis, P.T., Ponce, F., Goehring, B., Maurer, M., Rabassa, J., Coronato, A., Marr, R., 2013b. Latest Pleistocene and Holocene glacier fluctuations in southernmost Tierra del Fuego, Argentina. Quat. Sci. Rev. 77, 70–79.
- Menounos, B., Goehring, B.M., Osborn, G., Margold, M., Ward, B., Bond, J., Clarke, G.K., Clague, J.J., Lakeman, T., Koch, J., 2017. Cordilleran Ice Sheet mass loss preceded climate reversals near the Pleistocene Termination. Science 358, 781–784.
- Menounos, B., Hugonnet, R., Shean, D., Gardner, A., Howat, I., Berthier, E., Pelto, B., Tennant, C., Shea, J., Noh, M., 2019. Heterogeneous changes in western North American glaciers linked to decadal variability in zonal wind strength. Geophys. Res. Lett. 46, 200–209.
- Menounos, B., Osborn, G., Clague, J.J., Luckman, B.H., 2009. Latest Pleistocene and Holocene glacier fluctuations in western Canada. Quat. Sci. Rev. 28, 2049–2074.
- Menounos, B., Pelto, B.M., Fleming, S.W., Moore, R.D., Weber, F., Hutchinson, D., Brahney, J., 2020. Glaciers in the Canadian Columbia Basin, Technical Report.
- Menounos, B., Schiefer, E., Slaymaker, O., 2006. Nested temporal suspended sediment yields, green lake basin, British Columbia, Canada. Geomorphology 79, 114–129.
- Miller, G.H., Brigham-Grette, J., Alley, R., Anderson, L., Bauch, H.A., Douglas, M., Edwards, M., Elias, S., Finney, B., Fitzpatrick, J.J., 2010. Temperature and precipitation history of the Arctic. Quat. Sci. Rev. 29, 1679–1715.
- Miller, G.H., Geirsdóttir, Á., Zhong, Y., Larsen, D.J., Otto-Bliesner, B.L., Holland, M.M., Bailey, D.A., Refsnider, K.A., Lehman, S.J., Southon, J.R., 2012. Abrupt onset of the Little Ice Age triggered by volcanism and sustained by sea-ice/ocean feedbacks. Geophys. Res. Lett. 39.
- Miller, G.H., Lehman, S.J., Refsnider, K.A., Southon, J.R., Zhong, Y., 2013. Unprecedented recent summer warmth in Arctic Canada. Geophys. Res. Lett. 40, 5745–5751.
- Milner, A.M., Khamis, K., Battin, T.J., Brittain, J.E., Barrand, N.E., Füreder, L., Cauvy-Fraunié, S., Gíslason, G.M., Jacobsen, D., Hannah, D.M., 2017. Glacier shrinkage driving global changes in downstream systems. Proc. Natl. Acad. Sci. 114, 9770–9778.
- Monger, J., Price, R., 2002. The Canadian Cordillera: geology and tectonic evolution. CSEG Rec. 27, 17–36.

- Mood, B.J., Smith, D.J., 2015. Holocene glacier activity in the British Columbia Coast Mountains, Canada. Quat. Sci. Rev. 128, 14–36.
- Moore, R., Fleming, S., Menounos, B., Wheate, R., Fountain, A., Stahl, K., Holm, K., Jakob, M., 2009. Glacier change in western North America: influences on hydrology, geomorphic hazards and water quality. Hydrol. Process. Int. J. 23, 42–61.
- Mukherjee, K., Menounos, B., Shea, J., Mortezapour, M., Ednie, M., Demuth, M.N., 2023. Evaluation of surface mass-balance records using geodetic data and physically-based modelling, Place and Peyto glaciers, western Canada. J. Glaciol. 69, 665–682.
- Myhre, G., Highwood, E.J., Shine, K.P., Stordal, F., 1998. New estimates of radiative forcing due to well mixed greenhouse gases. Geophys. Res. Lett. 25, 2715–2718.
- Naegeli, K., Huss, M., Hoelzle, M., 2019. Change detection of bare-ice albedo in the Swiss Alps. The Cryosphere 13, 397–412.
- Nesje, A., Dahl, S.O., Bakke, J., 2004. Were abrupt Lateglacial and early-Holocene climatic changes in northwest Europe linked to freshwater outbursts to the North Atlantic and Arctic Oceans? The Holocene 14, 299–310.
- Nesje, A., Matthews, J.A., Dahl, S.O., Berrisford, M.S., Andersson, C., 2001. Holocene glacier fluctuations of Flatebreen and winter-precipitation changes in the Jostedalsbreen region, western Norway, based on glaciolacustrine sediment records. The Holocene 11, 267–280.
- Nishiizumi, K., Imamura, M., Caffee, M.W., Southon, J.R., Finkel, R.C., McAninch, J., 2007. Absolute calibration of 10Be AMS standards. Nucl. Instrum. Methods Phys. Res. Sect. B Beam Interact. Mater. At. 258, 403–413.
- Nuth, C., Kääb, A., 2011. Co-registration and bias corrections of satellite elevation data sets for quantifying glacier thickness change. The Cryosphere 5, 271–290.
- Nye, J.F., 1960. The response of glaciers and ice-sheets to seasonal and climatic changes. Proc. R. Soc. Lond. Ser. Math. Phys. Sci. 256, 559–584.
- Obu, J., Westermann, S., Bartsch, A., Berdnikov, N., Christiansen, H.H., Dashtseren, A., Delaloye, R., Elberling, B., Etzelmüller, B., Kholodov, A., 2019. Northern Hemisphere permafrost map based on TTOP modelling for 2000–2016 at 1 km2 scale. Earth-Sci. Rev. 193, 299–316.
- Oerlemans, J., 1994. Quantifying global warming from the retreat of glaciers. Science 264, 243–245.
- Osborn, G., McCarthy, D., LaBrie, A., Burke, R., 2015. Lichenometric dating: science or pseudo-science? Quat. Res. 83, 1–12.
- Osborn, G., Menounos, B., Koch, J., Clague, J.J., Vallis, V., 2007. Multi-proxy record of Holocene glacial history of the Spearhead and Fitzsimmons ranges, southern Coast Mountains, British Columbia. Quat. Sci. Rev. 26, 479–493.
- Osborn, G., Robinson, B., Luckman, B., 2001. Holocene and latest Pleistocene fluctuations of Stutfield Glacier, Canadian Rockies. Can. J. Earth Sci. 38, 1141–1155.
- Osterberg, E.C., Mayewski, P.A., Fisher, D.A., Kreutz, K.J., Maasch, K.A., Sneed, S.B., Kelsey, E., 2014. Mount Logan ice core record of tropical and solar influences on Aleutian Low variability: 500–1998 AD. J. Geophys. Res. Atmospheres 119, 11–189.
- Owens, M.J., Lockwood, M., Hawkins, E., Usoskin, I., Jones, G.S., Barnard, L., Schurer, A., Fasullo, J., 2017. The Maunder minimum and the Little Ice Age: an update from recent reconstructions and climate simulations. J. Space Weather Space Clim. 7, A33.
- Pendleton, S.L., Briner, J.P., Kaufman, D.S., Zimmerman, S.R., 2017. Using cosmogenic 10Be exposure dating and lichenometry to constrain Holocene glaciation in the central Brooks Range, Alaska. Arct. Antarct. Alp. Res. 49, 115–132.
- Phillips, F.M., Argento, D.C., Balco, G., Caffee, M.W., Clem, J., Dunai, T.J., Finkel, R., Goehring, B., Gosse, J.C., Hudson, A.M., 2016. The CRONUS-Earth project: a synthesis. Quat. Geochronol. 31, 119–154.
- Porter, C., Howat, I., Noh, M.-J., Husby, E., Khuvis, S., Danish, E., Tomko, K., Gardiner, J., Negrete, A., Yadav, B., Klassen, J., Kelleher, C., Cloutier, M., Bakker, J., Enos, J., Arnold, G., Bauer, G., Morin, P., 2023. ArcticDEM - Mosaics, Version 4.1. https://doi.org/10.7910/DVN/3VDC4W
- Porter, C., Howat, I., Noh, M.-J., Husby, E., Khuvis, S., Danish, E., Tomko, K., Gardiner, J., Negrete, A., Yadav, B., Klassen, J., Kelleher, C., Cloutier, M., Bakker, J., Enos, J., Arnold, G., Bauer, G., Morin, P., 2022. ArcticDEM - Strips, Version 4.1. https://doi.org/10.7910/DVN/C98DVS
- Porter, S.C., 1986. Pattern and forcing of Northern Hemisphere glacier variations during the last millennium. Quat. Res. 26, 27–48.
- Porter, S.C., Denton, G.H., 1967. Chronology of neoglaciation in the North American Cordillera. Am. J. Sci. 265, 177–210.
- Rasmussen, S.O., Andersen, K.K., Svensson, A., Steffensen, J.P., Vinther, B.M., Clausen, H.B., Siggaard-Andersen, M., Johnsen, S.J., Larsen, L.B., Dahl-Jensen, D., 2006. A new Greenland ice core chronology for the last glacial termination. J. Geophys. Res. Atmospheres 111.
- Reichert, B.K., Bengtsson, L., Oerlemans, J., 2002. Recent glacier retreat exceeds internal variability. J. Clim. 15, 3069–3081.
- Renssen, H., Seppä, H., Heiri, O., Roche, D.M., Goosse, H., Fichefet, T., 2009. The spatial and temporal complexity of the Holocene thermal maximum. Nat. Geosci. 2, 411–414.
- Reyes, A.V., Wiles, G.C., Smith, D.J., Barclay, D.J., Allen, S., Jackson, S., Larocque, S., Laxton, S., Lewis, D., Calkin, P.E., 2006. Expansion of alpine glaciers in Pacific North America in the first millennium AD. Geology 34, 57–60.
- Reza, A.M., 2004. Realization of the contrast limited adaptive histogram equalization (CLAHE) for real-time image enhancement. J. VLSI Signal Process. Syst. Signal Image Video Technol. 38, 35–44.
- Riahi, K., Van Vuuren, D.P., Kriegler, E., Edmonds, J., O'neill, B.C., Fujimori, S., Bauer, N., Calvin, K., Dellink, R., Fricko, O., 2017. The Shared Socioeconomic Pathways and their energy, land use, and greenhouse gas emissions implications: An overview. Glob. Environ. Change 42, 153–168.
- Rinterknecht, V.R., 2003. Cosmogenic 10 Be chronology for the last deglaciation of the southern Scandinavian Ice Sheet.
- Roe, G.H., O'Neal, M.A., 2009. The response of glaciers to intrinsic climate variability: observations and models of late-Holocene variations in the Pacific Northwest. J. Glaciol. 55, 839–854.
- Rounce, D.R., Hock, R., Maussion, F., Hugonnet, R., Kochtitzky, W., Huss, M., Berthier, E., Brinkerhoff, D., Compagno, L., Copland, L., 2023. Global glacier change in the 21st century: Every increase in temperature matters. Science 379, 78–83.
- Rousseeuw, P.J., Croux, C., 1993. Alternatives to the median absolute deviation. J. Am. Stat. Assoc. 88, 1273–1283.

- Ruddiman, W.F., Fuller, D.Q., Kutzbach, J.E., Tzedakis, P.C., Kaplan, J.O., Ellis, E.C., Vavrus, S.J., Roberts, C.N., Fyfe, R., He, F., 2016. Late Holocene climate: Natural or anthropogenic? Rev. Geophys. 54, 93–118.
- Ryder, J., Thomson, B., 1986. Neoglaciation in the southern Coast Mountains of British Columbia: chronology prior to the late Neoglacial maximum. Can. J. Earth Sci. 23, 273– 287.
- Schaefer, J.M., Codilean, A.T., Willenbring, J.K., Lu, Z.-T., Keisling, B., Fülöp, R.-H., Val, P., 2022. Cosmogenic nuclide techniques. Nat. Rev. Methods Primer 2, 1–22.
- Schaefer, J.M., Denton, G.H., Kaplan, M., Putnam, A., Finkel, R.C., Barrell, D.J., Andersen, B.G., Schwartz, R., Mackintosh, A., Chinn, T., 2009. High-frequency Holocene glacier fluctuations in New Zealand differ from the northern signature. science 324, 622–625.
- Schimmelpfennig, I., Schaefer, J.M., Lamp, J., Godard, V., Schwartz, R., Bard, E., Tuna, T., Akçar, N., Schlüchter, C., Zimmerman, S., 2022. Glacier response to Holocene warmth inferred from in situ 10 Be and 14 C bedrock analyses in Steingletscher's forefield (central Swiss Alps). Clim. Past 18, 23–44.
- Segal, R.A., Lantz, T.C., Kokelj, S.V., 2016. Acceleration of thaw slump activity in glaciated landscapes of the Western Canadian Arctic. Environ. Res. Lett. 11, 034025.
- Sigl, M., Toohey, M., McConnell, J.R., Cole-Dai, J., Severi, M., 2022. Volcanic stratospheric sulfur injections and aerosol optical depth during the Holocene (past 11 500 years) from a bipolar ice-core array. Earth Syst. Sci. Data 14, 3167–3196.
- Sigl, M., Toohey, M., McConnell, J.R., Cole-Dai, J., Severi, M., 2021. HolVol: Reconstructed volcanic stratospheric sulfur injections and aerosol optical depth for the Holocene (9500 BCE to 1900 CE).
- Smith, C.A., Bennett, B.A. (Eds.), 2004. Ecoregions of the Yukon Territory: Biophysical properties of Yukon landscapes. Summerland, BC: Agriculture and Agri-food Canada.
- Solomina, O.N., Bradley, R.S., Hodgson, D.A., Ivy-Ochs, S., Jomelli, V., Mackintosh, A.N., Nesje, A., Owen, L.A., Wanner, H., Wiles, G.C., 2015. Holocene glacier fluctuations. Quat. Sci. Rev. 111, 9–34.
- Steinhilber, F., Beer, J., Fröhlich, C., 2009. Total solar irradiance during the Holocene. Geophys. Res. Lett. 36.
- Stroeven, A.P., Fabel, D., Codilean, A.T., Kleman, J., Clague, J.J., Miguens-Rodriguez, M., Xu, S., 2010. Investigating the glacial history of the northern sector of the Cordilleran Ice Sheet with cosmogenic 10Be concentrations in quartz. Quat. Sci. Rev. 29, 3630–3643.
- Stroeven, A.P., Fabel, D., Margold, M., Clague, J.J., Xu, S., 2014. Investigating absolute chronologies of glacial advances in the NW sector of the Cordilleran Ice Sheet with terrestrial in situ cosmogenic nuclides. Quat. Sci. Rev. 92, 429–443.
- Tam, B.Y., Szeto, K., Bonsal, B., Flato, G., Cannon, A.J., Rong, R., 2019. CMIP5 drought projections in Canada based on the Standardized Precipitation Evapotranspiration Index. Can. Water Resour. JournalRevue Can. Ressour. Hydr. 44, 90–107.
- Tennant, C., Menounos, B., 2013. Glacier change of the Columbia Icefield, Canadian Rocky Mountains, 1919–2009. J. Glaciol. 59, 671–686.
- Thiagarajan, N., Subhas, A.V., Southon, J.R., Eiler, J.M., Adkins, J.F., 2014. Abrupt pre-Bølling–Allerød warming and circulation changes in the deep ocean. Nature 511, 75–78.
- Tierney, J.E., Poulsen, C.J., Montañez, I.P., Bhattacharya, T., Feng, R., Ford, H.L., Hönisch, B., Inglis, G.N., Petersen, S.V., Sagoo, N., 2020. Past climates inform our future. science 370, eaay3701.

- Tomkins, J.D., Lamoureux, S.F., Sauchyn, D.J., 2008. Reconstruction of climate and glacial history based on a comparison of varve and tree-ring records from Mirror Lake, Northwest Territories, Canada. Quat. Sci. Rev. 27, 1426–1441.
- Tomkins, M.D., Dortch, J.M., Hughes, P.D., Huck, J.J., Pallàs, R., Rodés, Á., Allard, J.L., Stimson, A.G., Bourlès, D., Rinterknecht, V., 2021. Moraine crest or slope: An analysis of the effects of boulder position on cosmogenic exposure age. Earth Planet. Sci. Lett. 570, 117092.
- Tomkins, M.D., Huck, J.J., Dortch, J.M., Hughes, P.D., Kirbride, M.P., Barr, I.D., 2018. Schmidt Hammer exposure dating (SHED): Calibration procedures, new exposure age data and an online calculator. Quat. Geochronol. 44, 55–62.
- Toohey, M., Sigl, M., 2019. Reconstructed volcanic stratospheric sulfur injections and aerosol optical depth, 500 BCE to 1900 CE, version 3. https://doi.org/10.26050/WDCC/eVolv2k_v3
- Urbanski, J.A., 2018. A GIS tool for two-dimensional glacier-terminus change tracking. Comput. Geosci. 111, 97–104.
- van Dijk, E.J., Jungclaus, J., Sigl, M., Timmreck, C., Krüger, K., 2024. High-frequency climate forcing causes prolonged cold periods in the Holocene. Commun. Earth Environ. 5, 242.
- Walker, A., Hidy, A.J., Zimmerman, S., Tumey, S.J., Brown, T.A., Braumann, S.M., Schwartz, R., Schaefer, J.M., 2022. Improved precision for 10Be-based chronology by optimizing carrier addition.
- Wanner, H., Beer, J., Bütikofer, J., Crowley, T.J., Cubasch, U., Flückiger, J., Goosse, H., Grosjean, M., Joos, F., Kaplan, J.O., 2008. Mid-to Late Holocene climate change: an overview. Quat. Sci. Rev. 27, 1791–1828.
- Wanner, H., Solomina, O., Grosjean, M., Ritz, S.P., Jetel, M., 2011. Structure and origin of Holocene cold events. Quat. Sci. Rev. 30, 3109–3123.
- Wiles, G.C., D'Arrigo, R.D., Villalba, R., Calkin, P.E., Barclay, D.J., 2004. Century-scale solar variability and Alaskan temperature change over the past millennium. Geophys. Res. Lett. 31.
- Young, N.E., Briner, J.P., Kaufman, D.S., 2009. Late Pleistocene and Holocene glaciation of the Fish Lake valley, northeastern Alaska Range, Alaska. J. Quat. Sci. Publ. Quat. Res. Assoc. 24, 677–689.
- Zemp, M., Hoelzle, M., Haeberli, W., 2009. Six decades of glacier mass-balance observations: a review of the worldwide monitoring network. Ann. Glaciol. 50, 101–111.