Abstract

This thesis is a five-chapter investigation of glacier change in the Cariboo Mountains of British Columbia. In chapter one I discuss the importance of glaciers, introduce the glaciers of the Cariboo Mountains, and outline the objectives and structure of this thesis.

In chapter two I compare three methods to estimate annual glacier mass balance of a 9.5 km² mountain glacier for years 2009, 2010, and 2011. I find two geodetic methods, real-time kinematic GPS (global positioning system) and photogrammetry, to provide a valuable measure of glacier-wide annual mass balance that is complementary to the glaciological method.

In chapter three I reconstruct the terminus position of the same mountain glacier for the period 1959-2007 from a series of annual push moraines. Annual recession of this glacier, the longest record for a North American glacier, is controlled by air temperature during the ablation season and accumulation season precipitation during the previous decade. I demonstrate an immediate glacier terminus reaction to summer and annual mass balance and a delayed reaction to winter and annual balance.

In chapter four I calculate dimensional change for 33 representative glaciers in the Cariboo Mountains for the latter half of the twentieth century. I show the period 1952-1985, when nine glaciers advanced, to be one of little net change for Cariboo Mountains glaciers. After 1985, however, rates of recession and thinning increased substantially. Comparison with climatological records reveals this marked change is due to both increased ablation season temperature and decreased accumulation season precipitation.
I show glacier response to climate over this period to be highly variable and that relations between response and glacier morphometry are not consistent temporally.

In chapter five I conclude this thesis with the progress gained through my research, study limitations, and the knowledge gaps that remain. Finally, I make 10 recommendations that will address knowledge gaps, and improve understanding of glacier change.
# Table of Contents

Abstract ...................................................................................................................................... ii

Table of Contents ..................................................................................................................... iv

List of Tables ............................................................................................................................ vi

1. Introduction ........................................................................................................................1
   1.1. Importance of glaciers .............................................................................................1
   1.2. Glaciers and ice caps ...............................................................................................3
   1.3. Glaciers of the Cariboo Mountains .........................................................................5
   1.4. Thesis objectives and outline ..................................................................................7

2. An evaluation of mass balance methods applied to Castle Creek Glacier, British Columbia, Canada ...................................................................................................................10
   2.1. Abstract ...................................................................................................................10
   2.2. Introduction ............................................................................................................11
   2.3. Theory: Conservation of mass and vertical velocity .........................................13
   2.4. Methods ..................................................................................................................16
   2.5. Error analysis .........................................................................................................27
   2.6. Results ....................................................................................................................33
   2.7. Discussion and recommendations for future mass-balance monitoring ...........43

3. Annual Push Moraines as Climate Proxy ...................................................................... 48
   3.1. Abstract ...................................................................................................................48
   3.2. Introduction ............................................................................................................49
   3.3. Study area and methods .........................................................................................50
   3.4. Results ....................................................................................................................54
   3.5. Discussion and conclusions ..................................................................................58

4. Glacier change in the Cariboo Mountains, British Columbia, Canada (1952-2005) .62
   4.1. Abstract ...................................................................................................................62
   4.2. Introduction ............................................................................................................63
   4.3. Methods ..................................................................................................................66
List of Tables

Table 2.1  Aerial photography and stereo model details. GCP, as used in column eight, is an acronym for ground control point.................................................................19

Table 2.2  Estimates of glacier-wide mass balance ($B_a$) by the three different methods. .34

Table 2.3  Varying photogrammetric $B_a$ results when using different melt factors in my temperature index model correcting for dates of photography. Percentages in parentheses indicate differences from results using Place Glacier melt factors........40

Table 3.1  Descriptive data for the 10 aerial photos used in this study. 'A' denotes federal photography, accessed from the National Air Photo Library, Natural Resources Canada, Ottawa, Ontario. 'BC' denotes provincial photography, Crown Registry and Geographic Base, Victoria, British Columbia..............................................................51

Table 3.2  Correlation table of western Canada and Pacific Northwest U.S.A. glacier frontal variation (FV) and mass balance ($b_s$ and $b_a$) time series. Bold values indicate significance ($p < 0.05$). Number of paired observations ($n$) displayed on lower half of table. See Table 3.3 for varying years of observation of each time series........53

Table 3.3  Years of observation for the records of glacier frontal variation and mass balance used in this study. The record 'Castle FV' has been derived from push moraines (this paper). The abbreviation 'FV' denotes frontal variation and $b_s$ and $b_a$ represented annual balance and summer balance respectively..............................54

Table 4.1  Aerial photography used to derive glacier extent and thickness change data. .67

Table 4.2  Morphometric properties of the 33-glacier subset...............................................................74

Table 4.3  Estimated error in extent for each year and extent change for three successive periods and two cumulative periods. ..............................................................................75

Table 4.4  Glacier extent and area change data for glaciers and summed by subregion and size class .................................................................81

Table 4.5  Median glacier morphometry for glaciers that underwent some advance, receded continuously, or exhibited no discernable change (within margin of error) during the period 1952-1985.................................................................84

Table 4.6  Correlation table (Pearson product-moment) of relations of glacier area change with respect to glacier morphometry for three periods. ................................................85
Table 4.7  Average annual thickness change in meters of water equivalent of seven glaciers for three periods. ..................................................................................................................86
List of Figures

Figure 1.1 Study area. The three red rectangles indicate subregions of focus in this thesis (Castle, Quanstrom, and Premier). All western Canada glaciers, below 60°N (Bolch et al., 2010) are shown in the inset in gray, along with the location of the Cariboo Mountains (red rectangle)..........................................................6

Figure 2.1 Location of Castle Creek Glacier and nearby glaciers with reference mass-balance series (inset). The inset also includes nearby towns, major cities and two main stems of the Fraser River, into which Castle Creek Glacier meltwater flows. Check patches are labeled by their respective mean elevation.........................17

Figure 2.2 Elevation residuals of check patches collected from three different stereo models. Each boxplot displays the distribution of surface-elevation residuals of 25 check points, defined as the difference in surface elevation of the same horizontal coordinates from the first stereo model to the second in a given epoch. The three boxplots for each check patch show residuals for the epochs 2008 - 2009, 2009 - 2011, and 2008 - 2011 respectively from left to right. See figure 2.1 for check-patch locations.........................................................20

Figure 2.3 (A) Grids of photogrammetric mass points for 2011 (black dots) and 2008 and 2009 (black dots and open circles), (B) subsets of mass points for “long profile” (black dots) and “grid” points (open triangles), and (C) “walkable routes” (open circles) and “arrays” (black triangles). ...........................................................................21

Figure 2.4 Thickness change for balance year 2009 from manual aerial photogrammetry. The two sets of mass points are from the first (v1) and sixth (v6) measurements of mass points, illustrating improved performance of the photogrammetrist..........22

Figure 2.5 Measurements of at-a-point mass balance and thickness change, and associated profiles with elevation, for balance years 2009, 2010 and 2011. Spatial extrapolation uses the hypsometry displayed at left. Error bars in the top panel indicate 1σ of the photogrammetric measurements within each 50 m elevation bin. ..........................................................................................................................25

Figure 2.6 Four subsets (triangles) of the 2009 photogrammetric mass points (circles): a) Points on a longitudinal profile along the center of the glacier, b) points along safely walkable longitudinal profiles, c) 37 array-point locations, and d) points from an evenly-spaced grid. Values in the upper right corner of each panel indicate B, for the associated subset. Spatial locations of each subset are displayed in figure 2.3.35

Figure 2.7 Same as Figure 2.5, but for two multi-year periods: 2009 – 2011 and 2010 - 2011.................................................................37
Figure 2.8 Frequency distributions of repeat-measurement residuals of 275 checkpoints for stereo models from 2008, 2009, and 2011 aerial photography used to constrain potential error and bias of the analyst in manual photogrammetry. Each distribution displays surface-elevation residuals, defined as the difference in surface elevation of the same horizontal coordinates from the same stereo model and by the same operator, but from initial measurements and a repeat measurement at a later date...41

Figure 2.9 Spatial distribution of mass balance, thickness change, and vertical velocity from the 20-point ablation-zone array for the month of August in 2008, 2009, and 2010. All measurements in ice equivalent units. I derived mass balance from stake measurements, thickness change from RTK GPS, and vertical velocity as the difference of the two. Locations of stakes and GPS measurements are indicated by black dots...44

Figure 3.1 A) Location of glaciers and climate stations referenced in this paper; B) Oblique photo of a portion of the Castle Creek Glacier forefield taken 11 September 2008. Dates refer to the year of moraine formation. The white circle encloses a 6 m high weather station. Some moraines used in my study are not visible in this photo. .................................................................50

Figure 3.2 Annual and cumulative frontal variation (1959 to 2007) of Castle Creek Glacier derived from annual push moraines (grey). Annual and cumulative length changes of South Cascade Glacier, Washington (red) are displayed for comparison. ............................................................................................................56

Figure 3.3 Cross correlation analysis of Castle Creek Glacier frontal variation versus Place Glacier seasonal and annual mass balance. Solid circles denote significance ($p < 0.05$). Solid lines are second-order polynomial curves fitted to the series of lagged correlation coefficients. The common period of analysis is 1965 to 2007 for $b_8$ ($n = 43$), and 1965 to 1989; 1994 to 1995 for $b_w$ and $b_5$ ($n = 27$). ................. 59

Figure 4.1 Study area: the three red rectangles indicate subregions in this study (Castle, Quanstrom, and Premier) and the location of the three maps in Figure 4.2 ...............65

Figure 4.2 Cariboo Mountains subregions showing subset of 33 glaciers, including extents for 1952, 1970, 1985, and 2005. Panels show from left to right the Castle, Quanstrom, and Premier subregions. Numerical glacier identification, numbered by 2005 surface area from smallest to largest, corresponds to that of Tables 4.2 and 4.4. ............................................................................................................70

Figure 4.3 Box-and-whisker plots display the maximum, interquartile range, median, and minimum of surface elevation residuals of 275 checkpoints. Checkpoints are used to determine relative accuracy of stereo models and bias correction for measurement
of surface elevation change of glaciers in the Castle and Quanstrom regions in three periods. ..............................................................................................................................71

Figure 4.4 Fraction of total glacierized extent by size class for all Cariboo Mountains glaciers (gray) and my 33-glacier subset (black)........................................................................78

Figure 4.5 Hypsometries for my subset of 33 individual glaciers (gray), the 33-glacier subset (black), all Cariboo Mountains glaciers (red), and the Castle, Quanstrom, and Premier subregions (darker to lighter blue respectively)..............................79

Figure 4.6 Fraction of total surface area by aspect for all Cariboo Mountains glaciers (black) and my 33-glacier subset (gray)...........................................................................80

Figure 4.7 Box-and-whisker plots displaying the maximum, interquartile range, median, and minimum of average annual rates of relative area change for (A) 32 glaciers for three periods, and (B) for 26 individual glaciers for four periods. Average annual rates are calculated using actual duration between imaging for each glacier, however this duration differs by subregion (Table 4.1)...............................83

Figure 4.8 Box-and-whisker plots display the maximum, interquartile range, median, and minimum of average annual thickness change in meters of water equivalent of seven glaciers for three periods. .................................................................86

Figure 4.9 ClimateWNA temperature records of deviation from the long-term (1952-2005) mean of (A) average accumulation season temperature, (B) average ablation season temperature, and (C) average annual temperature. Gray bars and values indicate average deviation from the long-term mean for the periods 1952-1970, 1970-1985, and 1985-2005.............................................................................................88

Figure 4.10 ClimateWNA precipitation records of deviation from the long-term (1952-2005) mean of (A) total accumulation season precipitation, (B) total ablation season precipitation, and (C) total annual precipitation. Gray bars and values indicate average deviation from the long-term mean for the periods 1952-1970, 1970-1985, and 1985-2005.................................................................................................89

Figure 4.11 Geopotential height (700 hPa) anomalies (m) for the accumulation and ablation seasons for three epochs. Color scale shows magnitude of anomaly. Data for the period 1952-2005 defines the mean fields. Dashed and solid contours respectively denote negative and positive anomalies that are significantly different ($p = 0.05$) from the mean at a given grid point............................................................................90

Figure 4.12 Box-and-whisker plots display the maximum, interquartile range, median, and minimum percent area change for 28 glaciers from 1985 to 2005 for this study and that of Bolch et al. (2010). .................................................................94
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1. Introduction

The primary focus of this thesis is to refine methods and use them to address shortfalls in our understanding of glacier change. I employ field-based and remote sensing methods to measure glacier change of the Cariboo Mountains of British Columbia (BC), Canada. A secondary focus of this thesis is the changes in extent and volume of glaciers of the Cariboo Mountains since the 1950s, and on the climatic drivers of this change.

1.1. Importance of glaciers

Glaciers are fundamental components of Earth’s climate and hydrologic systems. Accumulation season snowfall and ablation season ice melt represent the primary factors that drive annual changes in glacier mass. Long-term changes in glacier mass balance ultimately control glacier fluctuations at time scales of a decade or more. Climate variability affects accumulation and ablation and results in glacier change with impacts on numerous natural and human systems.

A majority of Earth’s freshwater is held in the ice and snow of glaciers. Glacier change thus serves as a governor of freshwater flow volume, flow timing, temperature, and quality in glacierized watersheds (e.g., Moore et al., 2009). Glacier recession can have adverse effects on municipal, agricultural and industrial water supply (e.g., Barnett et al., 2005; Marshall et al., 2011). Loss of glacier meltwater in a watershed may also impact riverine ecosystems and salmon productivity (Dorava and Milner, 2000), and coastal ecosystems through alteration of dissolved organic matter in rivers (Hood et al., 2009). Melting glaciers may also be a prominent source of persistent organic pollutants,
whereby pollutants which accumulate in glaciers through snowfall, are subsequently released during ice melt, often in concentrated amounts (e.g., Bogdal et al., 2009).

As glaciers change they can alter global sea level, which falls as glaciers expand and rises when glaciers shrink. Much uncertainty remains regarding the quantity and rate of ongoing and projected 21st century sea level rise (e.g., Meier et al., 2007; Radić and Hock, 2011; Rignot et al., 2011; and Jacob et al., 2012). It is plausible that glacier wastage will raise the global ocean by about 0.4 m by the end of this century (Bahr et al., 2009). Higher sea level rise from changes in global ice volume cannot be ruled out, due to a lack of understanding of the dynamic response of the Greenland and Antarctic ice sheets to current and projected warming (e.g., Rignot et al., 2011).

Glacier loss may increase geomorphic hazards such as rock fall, rock avalanches, debris flows, debris avalanches, and debris slides (e.g., Moore et al., 2009). Glacier change can also lead to flooding events through glacier outburst floods (ice or moraine dammed lakes). While most events occur in remote locations, notable exceptions exist where these floods caused significant damage to infrastructure and killed thousands of people (e.g., Huggel et al., 2008).

Glaciers also represent an important natural resource for mountain communities given their tourist appeal. Visitors contribute to local economies through glacier sightseeing, dogsledding, and climbing adventures. Additionally, some ski resorts rely heavily on glaciers for operations, including British Columbia resorts such as Whistler Blackcomb in the southern Coast Mountains and the proposed Jumbo Glacier Resort in the Purcell Mountains. In the Alps expensive efforts have been made in attempts to artificially preserve glaciers on which the ski industry depends (Smiraglia et al., 2008).
Glaciers receive their nourishment and are depleted by meteorological phenomena, but glaciers also control their local climate (e.g., Barry, 2002). Their highly reflective surfaces reflect shortwave radiation resulting in cool temperatures locally, regionally, and, in the case of major ice sheets, even continentally. This cooling also influences local winds and general atmospheric circulation. Conversely loss of high-albedo glacier surfaces results in warming and possible changes in atmospheric circulation.

Glacier change constitutes an independent, natural response of local-to-global climate change, especially for remote high-elevation locations where few meteorological measurements are made. Recent work has reconstructed global and hemispheric temperature records for the past 400 years from records of glacier recession (Leclercq and Oerlemans, 2011). Additionally, imagery of receding glaciers is among the most poignant depictions of recent climate change, and arguably plays a central role in the motivation to act on global warming:

“If the world does address the great challenge of global warming, it will be in part because of the way that glaciers serve as icons to make this challenge visible.”

(Orlove et al., 2008)

1.2. **Glaciers and ice caps**

This thesis focuses on changes of glaciers and ice caps (GIC), a term used here to refer to all glaciers excluding the Greenland and Antarctic ice sheets. GIC surface area comprises approximately 683,000 km² (Arendt et al., 2012), and total volume is estimated to represent a sea level rise equivalent (SLE) of 43 cm (Huss and Farinotti, 2012). Most GIC exist in western North America, Central Asia, and the Arctic islands of
Canada, Norway, and Russia (Zemp and van Woerden, 2008). The majority of the GIC in continental North America are found in Alaska and western Canada. At present the GIC of western Canada (south of 60°N) represents an area of 26,000 km$^2$, with most of the glacierized area located in the Saint Elias and Coast mountains of BC (Bolch et al., 2010).

Glaciers and how they change are measured via remote sensing methods or through direct measurements in the field (e.g., Cogley, 2005; Barry, 2006; Bamber and Rivera, 2007). Remote sensing of glaciers can be applied to measure extent, surface elevation, and surface velocity across large areas. Field measurements of extent, surface elevation, density, thickness, and velocity are made at limited points for a given glacier. Many of these field measurements are made with much greater precision than with remote sensing techniques. Remote sensing methods yield a comprehensive picture of global GIC change since the 1980s; however, detailed field measurements of seasonal and annual GIC change are scarce, with much understanding of glacier change coming from a subset of some 30 global ‘benchmark’ glaciers (Zemp et al., 2009).

Remote sensing and field-based measurements have yielded a general understanding of global GIC mass loss. In many regions this loss has accelerated in recent decades, primarily driven by increased surface air temperatures (e.g., Kaser et al., 2006). There have been and continue to be some exceptions to this general pattern, however. Regionally, some GICs of New Zealand and Norway advanced during the 1990s (e.g., Zemp and van Woerden, 2008), and GICs of the western Karakoram were noted to be growing in the 1990s (Hewitt, 2005); more recent work indicates that these latter glaciers are gaining mass (Gardelle et al., 2012). In each of these regions, gains in
mass and glacier advance have been attributed to increased accumulation-season snowfall, but also to complex high-elevation topography and the role of debris cover in the case of Karakoram GIC. Other individual glaciers that are an exception to the general pattern of recession have unique, marine-terminating ice dynamics (e.g., Ritchie et al., 2008; Truffer et al., 2009). However, the dominant global pattern is of GIC mass loss and recession, and is projected to continue throughout the 21st century and contribute significantly to global sea level (e.g., Radić and Hock, 2011).

1.3. **Glaciers of the Cariboo Mountains**

This thesis examines glaciers and glacier change of the Cariboo Mountains in the central interior of BC (Fig. 1.1). These glaciers, at the northern extent of the Columbia Mountains, are at the headwaters of the Fraser, North Thompson (a tributary of the Fraser) and Columbia rivers. The 200-km long Cariboo Mountains are bounded by the Rocky Mountain Trench and Rocky Mountains to the north and east respectively, the North Thompson River valley and Selkirk Mountains to the south, and the Interior Plateau to the west (Figure 1.1). Bowron Lake, Cariboo Mountains, and Wells Gray Provincial Parks lie at the western edge of the range.

The climate of the Cariboo Mountains is influenced by maritime and continental air masses, but the annual precipitation of the western margins of the range is high, with mean annual precipitation of 788-1,240 mm. This interior temperate rainforest, or wetbelt, is characterized by much higher precipitation than in the surrounding regions (Stevenson et al., 2011).
Figure 1.1 Study area. The three red rectangles indicate subregions of focus in this thesis (Castle, Quanstrom, and Premier). All western Canada glaciers, below 60°N (Bolch et al., 2010) are shown in the inset in gray, along with the location of the Cariboo Mountains (red rectangle).
Bolch et al. (2010) completed the first comprehensive inventory of the glaciers of western Canada. This data set includes the 526 glaciers of the Cariboo Mountains, which shrank from 845 km\(^2\) in 1985 to 731 km\(^2\) in 2005, a loss of 114 km\(^2\) or 13%. Schiefer et al. (2007) derived volume change of all BC glaciers for the period 1985-1999. This data set indicates Cariboo Glaciers lost -0.47 km\(^3\) a\(^{-1}\) for this period, equivalent to a thinning rate of -0.58 m a\(^{-1}\) (both values expressed in units of water equivalent). In the earliest previous study to focus explicitly on glaciers of the Cariboo Mountains, Luckman et al. (1986) studied 31 termini of glaciers in the Premier Range of the southern Cariboo Mountains. They found most of these glaciers advanced during the 1970s, and they attributed this advance to: (1) increased winter precipitation during the period 1951-1976; and (2) cooler than average summer temperatures from the period 1954-1968. Brewis (2012) examined glacier change in the Canoe Basin (a subset of Premier Range glaciers) and found glaciers shrank and lost volume from 1948-2005, but that their recession slowed and some advanced during the period 1955-1985.

1.4. Thesis objectives and outline

Previous work has synthesized the current understanding of global glacier change and made recommendations for future efforts and areas of focus (e.g., Fountain et al., 1999; Haeberli et al, 2007; Paul et al., 2007; Zemp et al., 2009). These studies recommend: (1) validating and calibrating field measurements with repeat mapping of glacier surface elevation; (2) extending annually measured glaciers to remote regions that are underrepresented; and (3) detailing strategies to study regional-to-global glacier change through a tiered approach. The two primary objectives of this thesis are built
upon these recommendations. The first objective of this thesis is to develop and test new methods that address glacier change. The second objective of this thesis is to determine glacier change of the Cariboo Mountains since the 1950s and the climatic drivers of this change.

This thesis is paper-based, and chapters 2, 3, and 4 are written as stand alone papers. I have made slight modifications to chapter content in this thesis to reduce repetition; these modifications are restricted to the introductory content of each chapter. At the beginning of each chapter I note the status of each paper within the peer-review process, which varies by chapter from ‘published’ to ‘in preparation’.

In chapter two, I compare glaciological, photogrammetric and real-time kinematic global positioning system (RTK GPS) methods of measuring annual mass balance of Castle Creek Glacier within the Castle region of the Cariboo Mountains. This chapter illustrates the facility of geodetic methods to improve our understanding of glacier mass balance, and recommends best practices for future mass-balance studies. This chapter is ‘in press’ in Journal of Glaciology (Beedle et al., in press).

Chapter three documents a series of annual push moraines in the forefield of Castle Creek Glacier. From these geomorphic features I reconstruct the longest annually-resolved record of glacier length change for a North American glacier (1959-2007), and relate these changes to regional climate variability. This chapter has been peer reviewed and is published in Geophysical Research Letters (Beedle et al., 2009).

In chapter four, I use aerial photogrammetry to investigate the extent and volume change of Cariboo Mountains glaciers from the 1950s to 2005. This chapter focuses on a subset of glaciers from three subregions of the Cariboo Mountains. From northwest to
southeast these are the Castle, Quanstrom, and Premier regions, which include most of the glaciers in the Cariboo Mountains (Fig. 1.1). This chapter has not been peer reviewed and is in preparation for submission to The Cryosphere (Beedle et al., *in prep.*).

I conclude this thesis with a fifth chapter that synthesizes the previous chapters. This synthesis focuses on the broad implications of, and applications for, the research completed herein, including suggested applications of refined methodologies, and approaches for future glacier monitoring on the scale of individual glaciers to mountain ranges.
2. An evaluation of mass balance methods applied to Castle Creek Glacier, British Columbia, Canada

Publication details:

This chapter is 'in press' for publication in Journal of Glaciology. Please see Appendix A: Authorship Statements for details of the contributions of each author.


2.1. Abstract

I estimate glacier mass balance for years 2009, 2010, and 2011 for a 9.5 km² mountain glacier using three approaches. The photogrammetric, global positioning system (GPS), and glaciological methods yielded sampling densities of 100, 5 and 2 points km⁻² with measurement precisions of ±0.40, ±0.10, and ±0.10 m water equivalent (w.e.) respectively. My glaciological measurements likely include a positive bias, due to omission of internal and basal mass change, and error in determining the interface between snow and firm with a probe (±0.10 m w.e.). Measurements from my photogrammetric method include a negative bias introduced by the manual operator and my temperature index model used to correct for different dates of imaging (0.15 m w.e.), whereas GPS measurements avoid these biases. The photogrammetric and GPS methods are suitable to estimate glacier-wide annual mass balance, and thus they provide a
valuable measure that is complementary to the glaciological method. These approaches, however, cannot be used to estimate mass balance at a point or mass-balance profiles without a detailed understanding of the vertical component of ice velocity.

2.2. Introduction

Geodetic estimates of glacier mass balance quantify changes in surface elevation over a given time period and employ density assumptions for snow and ice. Most geodetic studies assess mass change over periods of a decade or more (e.g., Arendt et al., 2002; Schiefer et al., 2007). Some studies, however, have used geodetic methods to determine mass balance on shorter temporal scales. Meier and Tangborn (1965), for example, used aerial photography taken three years apart to estimate annual mass balance and to investigate short-term ice dynamics of South Cascade Glacier, Washington. Tangborn et al. (1975) made repeat field survey measurements of a 112-point grid on South Cascade Glacier during one ablation season. They found that their geodetic measurements were similar to the glaciological method. Rasmussen and Krimmel (1999) used aerial photogrammetry to derive annual specific mass balance over a portion of South Cascade Glacier for balance years 1993 and 1994. They demonstrated the utility of geodetic measurement, but also identified potential systematic biases in both geodetic and glaciological mass-balance measurements. Krimmel (1999) presented a comparison of South Cascade Glacier annual mass balance derived from both glaciological and photogrammetric methods for 12 balance years (1986-1997). The South Cascade Glacier cumulative geodetic balance was more negative than glaciological balance by about 0.25 m water equivalent (w.e.) a⁻¹, indicative of bias in one or both methods, and possibly due
to basal melt, density estimates, and stakes melting into the glacier surface. The study concluded that photogrammetry can be used to determine annual mass balance if geodetic control is consistent for stereo models.

In the late-1990s, global positioning systems (GPS) were tested as a method to estimate mass balance (Eiken et al., 1997; Gandolfi et al., 1997; Jacobsen and Theakstone, 1997). These studies demonstrated the potential of using GPS in kinematic mode, whereby point measurements are made continuously at a preset time interval, to record glacier surface elevation. Without using real-time kinematic (RTK) GPS, where a real-time differential correction is received from the base station through radio transmission allowing centimeter accuracy, however, these early studies were only able to re-occupy points on a glacier to within tens of meters. Hagen et al. (1999) used GPS to estimate mass balance of Kongsvegen, Svalbard for the period 1991-1995 and found that the results fit well with field measurements of a glacier with negligible vertical velocity. Previous work noted the rapidity and accuracy of kinematic GPS measurements, which might lead to an increase in the number of monitored glaciers (Hagen et al., 1999; Theakstone et al., 1999). Hagen et al. (2005) presented multi-year comparisons of GPS profiles along central flowlines of three Svalbard glaciers. They concluded that changes in glacier geometry cannot be used to assess mass balance without independent knowledge of vertical velocity. In this study, RTK GPS was not used and re-occupation of previously measured points was made only to within 30-90 meters. Other studies have used RTK GPS to accurately re-occupy and re-measure survey points on a glacier to assess multi-year mass balance (e.g., Nolan et al., 2005).
In this chapter I compare glaciological, photogrammetric, and RTK GPS methods to estimate annual glacier mass balance. My objectives are to: 1) test the RTK GPS method; 2) investigate potential error in the glaciological method by comparison with two geodetic methods; and 3) to make recommendations for future mass balance monitoring.

2.3. Theory: Conservation of mass and vertical velocity

Geodetic measurements of glacier thickness change incorporate two dominant terms, surface mass balance and the vertical component of ice flow, necessitating consideration of flux divergence. Conservation of mass at a point on the surface of a glacier (in ice equivalent units, ice eq.) can be stated as:

\[
h = \frac{b - \nabla \cdot \vec{Q}}{\rho}
\]  
(2.1)

where \( \dot{h} \) is the rate of thickness change, \( \dot{b} \) is the specific surface mass balance rate, \( \rho \) is density of the surface material relative to water, and \( \nabla \cdot \vec{Q} \) is a flux-divergence term (e.g., Rasmussen and Krimmel, 1999; Cuffey and Paterson, 2010). Implicit in Eqn (2.1) are the assumptions that densification, internal and basal mass changes, isostatic displacement, and erosion of the bed surface are all negligible. When integrated across the entire glacier surface, and assuming a negligible influence from changing surface area and no flux across the glacier boundary (e.g., from avalanching or calving), flux divergence is zero, yielding:
\[ \dot{H} = \frac{\dot{B}}{\rho} \]  

(2.2)

where \( \dot{H} \) and \( \dot{B} \) are glacier-wide integrations of thickness change and surface mass balance respectively.

Flux divergence, however, is not zero for a given point, and the vertical component of ice velocity at the surface (vertical velocity, \( w_s \)) plays a confounding role in deriving \( \dot{b} \) from measurements of \( \dot{h} \). Previous studies use Eqn (2.1) to estimate \( \dot{b} \) from geodetic measurements and include a complete treatment of the flux divergence term in Eqn (2.1), with consideration of the vertical profile of velocity and the component of surface flow due to sliding (e.g., Gudmundsson and Bauder, 1999; Rasmussen and Krimmel, 1999).

In studies to quantify \( w_s \), some efforts include consideration of sub-surface glacier flow or assume steady-state conditions (e.g., Reeh et al, 1999; Reeh et al., 2003), while others neglect the vertical profile of velocity (e.g., Meier and Tangborn, 1965; Holmlund, 1988; Pettersson et al., 2007), and rely on the kinematic boundary condition at the glacier surface:

\[ \dot{h} = \frac{\dot{b}}{\rho} + w_s - u_s \frac{\partial S}{\partial x} - v_s \frac{\partial S}{\partial y} \]  

(2.3)

where \( u_s \) and \( v_s \) are the horizontal components of ice velocity at the glacier surface \( S \) (Cuffey and Paterson, 2010). Often, Eqn (2.3) is used to estimate \( w_s \) in only the ablation zone and is further reduced to a simple geometric expression (e.g., Meier and Tangborn, 1965; Holmlund, 1988; Pettersson et al., 2007):

\[ w_s = \dot{h} + u_s \tan \alpha \]  

(2.4)
where $\dot{h}$ is thickness change measured at a marker (usually a stake) moving with glacier flow, $u_s$ is oriented along the flow, $v_s$ is assumed to be zero, and $\alpha$ is surface slope (Cuffey and Paterson, 2010).

Previous work that employs either Eqn (2.1) to estimate $\dot{b}$ from geodetic measurements, or Eqns (2.3) and (2.4) to estimate $w_s$ relies on a Lagrangian frame of reference, whereby horizontal flow and occasionally $\dot{h}$ are measured at a marker, such as a stake, as it moves with the ice. In an Eulerian frame of reference, where measurements are made at fixed coordinates, $\dot{h}$ changes as the sum of $\dot{b}$ and $w_s$ (Cuffey and Paterson, 2010). However, horizontal ice flux at the surface ($u_s$ and $v_s$ in Eqn (2.3)) advects glacier surface topography through locations where geodetic repeat measurements are made. I discuss this as a source of error below, but otherwise neglect advection of topography in my geodetic measurement of mass balance and estimation of $w_s$.

Omission of these horizontal velocity terms yields:

$$\dot{h} = \frac{\dot{b}}{\rho} + w_s$$ (2.5)

which may be rearranged to solve for specific mass balance:

$$\dot{b} = (\dot{h} - w_s)\rho$$ (2.6)

or for vertical velocity:

$$w_s = \dot{h} - \frac{\dot{b}}{\rho}$$ (2.7)

In addition to advection of topography, Eqns (2.5-2.7) also neglect densification, internal and basal mass balance, isostatic displacement, and erosion of the bed surface.
Vertical velocity at the surface \( w_z \) is typically negative or downward (submergence) in the accumulation zone and positive or upward (emergence) in the ablation zone. To avoid ambiguity in my discussion of these, I use the term emergence to refer to positive (upward) flow, submergence to refer to negative (downward) flow, and vertical velocity as a general term without a specified sign.

2.4. Methods

I measured mass balance of Castle Creek Glacier, in the Cariboo Mountains of British Columbia, Canada (Beedle et al., 2009). This 9.5 km\(^2\) mountain glacier flows north for 5.9 km, has an elevation range of 2,827 to 1,810 m above sea level (asl), and contributes meltwater to Castle Creek, a tributary of the Fraser River (Fig. 2.2).

My terminology and notation follow the recommendations of Cogley et al. (2011). The glaciological method measures surface mass balance, whereas the two geodetic estimate mass change from measurements of elevation change, which includes surface, internal and basal ablation and accumulation. For simplicity, however, I refer only to annual mass balance regardless of method. A mass-balance profile, \( b(z) \), is defined as the variation of mass balance with elevation. I use \( dh \) as notation for thickness change, and \( dh(z) \) to refer to the profile of thickness change with elevation. All measurements of glacier-wide mass balance presented are conventional balances whereby values are averaged over a changing glacier surface (Elsberg et al., 2001), allowing direct comparison of glaciological and geodetic methods. The changing ice geometry is based on digital elevation models (DEMs) from 2008, 2009, and 2011 aerial
Figure 2.1 Location of Castle Creek Glacier and nearby glaciers with reference mass-balance series (inset). The inset also includes nearby towns, major cities and two main stems of the Fraser River, into which Castle Creek Glacier meltwater flows. Check patches are labeled by their respective mean elevation.
photos. I estimate the 2010 glacier geometry by linear interpolation from the 2009 and 2011 DEMs. Castle Creek Glacier surface areas used to convert from volume change to specific (per unit area) mass balance are 9.56 (2009), 9.52 (2010), and 9.49 km² (2011).

2.4.1. Glaciological method

Snow pits and probing are used to directly measure accumulation, whereas stakes measure surface ablation (e.g., Østrem and Brugman, 1991; Kaser et al., 2003). I use the stratigraphic system to define the annual mass balance, whereby measurements are made between successive annual minima, typically in early-September at Castle Creek Glacier. Conversion to w.e. is made by assuming ice density to be 900 kg m⁻³, and from snow density measured in snow pits. Measured snow density at the end of the ablation season is less spatially variable than snow depth, with relative standard deviations of <1% and 33% respectively. Point measurements of mass balance ($b_a$) were made at 21, 12, and 18 sites during the balance years of 2009, 2010 and 2011, equating to sampling densities of 2.2, 1.3, and 1.9 points km⁻² respectively. Glaciological measurements are absent for a portion of the middle of the glacier where an icefall impedes safe travel.

I apply the balance-gradient (or regression) method to extrapolate from $b_a$ measurements to glacier-wide annual mass balance ($B_a$), using surface area defined by 50 m elevation intervals (e.g., Fountain and Vecchia, 1999). For the glaciological method, I use a three-part linear spline to represent the variation of $b_a$ with elevation. This spline is derived from the $b_a$ measurements, with the intercept set to the observed elevation of the annual snowline ($b_a = 0$). My measurements do not reach the highest elevation bins of Castle Creek Glacier; I apply the measurements from my highest measurements to these
uppermost elevation zones of the glacier (e.g., Cogley et al., 1996). Use of a linear spline to interpolate from mass-balance observations has been found to be similar to a quadratic interpolation and superior to the contour method (Fountain and Vecchia, 1999).

2.4.2. Photogrammetric method

I use aerial photographs taken in 2008, 2009 and 2011 to derive $B_a$ for 2009, and cumulative balances for the periods 2008 – 2011, and 2009 – 2011 (Table 2.1). Ground sampling distance of the 2008 and 2009 images is 0.25 m, and 0.53 m for the 2011 images. Ground control points (GCPs) were obtained from stereo models of 2005 aerial triangulation scans made available by the Province of British Columbia.

I created stereo models from photography using the Vr Mapping photogrammetry software suite (Cardinal Systems LLC). A common set of 18 GCPs, consisting of bedrock features or stable boulders distributed around the glacier at various elevations and 50-70 tie points were used for exterior orientation and generation of stereo models (e.g., Schiefer and Gilbert, 2007; Barrand et al., 2009). The use of the same GCPs for all years ensured that positional errors were randomly distributed (Kääb & Vollmer, 2000; Schiefer and Gilbert, 2007; Schiefer et al., 2007). Analysis of 11 check patches allows me to measure the relative accuracy and assess systematic bias between stereo models. These check patches consist of 25 individual check points in a 5 m grid, located on stable
Each boxplot displays the distribution of surface-elevation residuals of 25 check points, defined as the difference in surface elevation of the same horizontal coordinates from the first stereo model to the second in a given epoch. The three boxplots for each check patch show residuals for the epochs 2008 - 2009, 2009 - 2011, and 2008 - 2011 respectively from left to right. See figure 2.1 for check-patch locations.

bedrock near the glacier (Figs. 2.1 and 2.2). I estimated systematic bias among stereo models and derived trend surfaces based on the mean residuals of the 11 check patches; these trend surfaces were then used to apply a correction for elevation points on the glacier. I also compared the average slope angle of each check patch with the mean residual of the 25 individual points to detect any horizontal bias among models.

To map glacier surface elevation, I manually digitized mass points (series of $x,y,z$ data points collected on a predetermined grid) on a 100 m grid within the glacier extent as manual measurements yield results superior to automated extraction methods (McGlone et al., 2004). Good contrast in the 2008 and 2009 photography enabled me to
measure every grid point \( (n = 937) \), but poor photographic contrast due to fresh snow cover in the 2011 photography reduced measurements by 26\% (Fig. 2.3).

The largest source of error in my photogrammetric methodology is my ability to perceive and measure the glacier surface, a methodological shortcoming noted by others (e.g., Rasmussen and Krimmel, 1999). This bias decreased with repeated measurements (Fig. 2.4). I define blunders as points that fall outside 68\% (± 1\( \sigma \)) of measured elevation change for a given 50 m elevation interval, and I re-measured blunders five times for the 2008 and 2009 models, and three times for the 2011 models. To correct persistent blunders (<5\%), I performed ordinary kriging interpolation from the non-blunder mass points, and extracted elevations from these trend surfaces.

As the photograph dates did not match the date of field measurements, I made corrections using a temperature-index model (Hock, 2003). I derived glacier surface temperature for the model from observations at an automated weather station near the east margin of the glacier (2105 m asl, Fig. 2.1), which is outside the influence of
katabatic winds (Déry et al., 2010). I assume a lapse rate of 0.006 °C m⁻¹, apply melt factors derived from mass-balance data (Shea et al., 2009) for Place Glacier ($k_s = 2.76$, $k_i = 4.67$), and compare my findings to results using melt factors from Peyto Glacier ($k_s = 2.34$, $k_i = 5.64$) and my more limited observations at Castle Creek Glacier ($k_s = 3.45$, $k_i = 4.33$).

Geodetic mass-balance methods necessitate assumptions of the density of ice and snow lost or gained from the surface of a glacier (e.g., Huss, 2013). I tested three scenarios to assess my density assumptions: the first scenario (A) assumes that the density profile of the glacier remains unchanged with time (Sorge’s law; Bader, 1954) and 900 kg m⁻³ is used to convert from ice eq. to w.e. My second scenario (B) uses 900
kg m$^{-3}$ for all points below the equilibrium line altitude (ELA), and 750 kg m$^{-3}$ for all points above the ELA – assuming that the loss or gain of material in the accumulation zone is not entirely composed of ice, but at least partially of firn and snow (e.g., Zemp et al., 2010). The third scenario (C) uses 900 kg m$^{-3}$ for all points below the ELA, and 600 kg m$^{-3}$ for all points above the ELA – assuming that the loss or gain of material in the accumulation zone has a density equivalent to end-of-season densities measured in my snow pits. These three scenarios use maps of accumulation and ablation zones based on glacier extent and observations of the ELA in the latter year of each period. Average glacier-wide densities for Castle Creek Glacier elevation change vary from 800-850 kg m$^{-3}$ for B, and from 699-726 kg m$^{-3}$ for C.

To calculate $B_a$ via the photogrammetric method, I use two methods of spatial extrapolation. For balance year 2009, when the 100 m grid is measured in its entirety, I apply the arithmetic mean of all 937 points. This assumes that all vertical velocities sum to zero (Eqn 2.2). The second method sums the product of the average $dh(z)$ from each 50 m elevation interval and its surface area. In this second method, the integral of the emergence velocities from the subset of points might be zero (e.g., along a flowline; Cogley, 2005), or non-zero, introducing an error or bias in the estimate of $B_a$.

### 2.4.3. RTK GPS method

To perform RTK GPS surveys (hereafter denoted as GPS) I used Topcon GB-1000 dual-frequency receivers (measurement precision of ±0.03 m). Re-occupation of previously measured points indicated a horizontal accuracy of ±0.03 m, and repeat measurements of three check points on stable bedrock respectively indicated horizontal
and vertical accuracy of 0.02 ± 0.01 m and 0.03 ± 0.02 m. I attached the rover antenna to a short antenna pole that was in turn attached to my backpack. The distance from the rover antenna to the glacier surface was measured on a flat surface and assumed to be constant (e.g., Nolan et al., 2005).

I made measurements of $dh$ via GPS along longitudinal profiles and at grid points (Figs. 2.1 and 2.3). For initial measurement of the longitudinal profiles, I collected points in kinematic mode at 5 s intervals and differentially corrected these points. From these kinematic profiles I selected points every 10 m in elevation for GPS measurement in successive years. I established grids of points on cross-glacier profiles at common elevations, yielding 20 points in the ablation zone and 16 points in the accumulation zone (Fig. 2.1).

Calculated $dh$ was converted to w.e. using the three density scenarios discussed above. GPS measurements were made at the same time as the glaciological measurements and thus no melt correction was needed between them. Unfortunately, poor line of sight with the base station and insufficient base radio power resulted in few measurements in the accumulation zone and I thus adopt the balance-gradient method to achieve $B_a$ using a linear spline and the period-specific hypsometry. For GPS measurements, I use a two-part linear spline fit to my 2011 observations (Fig. 2.5). This spline is then shifted to match 2009 and 2010 ablation-zone $b_a$ measurements. With this shift I assume the profile shape does not change from year to year, and that few lower-elevation observations can be used to adequately define $b(z)$ (e.g., Rasmussen and Krimmel, 1999).
Figure 2.5 Measurements of at-a-point mass balance and thickness change, and associated profiles with elevation, for balance years 2009, 2010 and 2011. Spatial extrapolation uses the hypsometry displayed at left. Error bars in the top panel indicate 1σ of the photogrammetric measurements within each 50 m elevation bin.
Additionally, I tested the sensitivity of using fewer, GPS measurements to estimate $B_a$ by using four different subsets of the 937 photogrammetric measurements for balance year 2009 (Fig. 2.3). These subsets include: 1) a “long profile” of 41 points along the glacier center-line, 2) “walkable routes” consisting of 56 points, which are the safely navigable routes of my GPS longitudinal-profiles, 3) “arrays” consisting of the 36 ablation and accumulation zone points measured in situ, and 4) a “grid” consisting of 61 points, regardless of safe travel.

2.4.4. Estimation of vertical velocity

I employed an Eulerian frame of reference to estimate vertical velocity at fixed coordinates using Eqn (2.7) with GPS measurements of $\dot{h}$ and glaciological measurements of $\dot{b}$. From August to September (2008-2010), I estimated vertical velocity for point arrays in the ablation and accumulation zones of Castle Creek Glacier (Fig. 2.1). I placed twenty ablation stakes (ablation array) in four across-glacier profiles. At each stake, which typically travelled 5-20 m a\textsuperscript{1} down glacier, I measured surface ablation; $dh$ was measured with GPS at fixed coordinates where stakes were initially placed. I assume that ablation measured at a transient stake is representative of ablation at the site where it was initially placed.

To estimate vertical velocity in the accumulation zone I made four probing observations within 3 m of the location where surface elevation was measured with GPS. The average of these multiple observations was used to minimize errors stemming from a non-vertical probe, the observer’s ability to accurately probe the previous summer surface, and the effects of meter-scale variability in the summer surface (e.g., from sun
cups, meltwater channels, and differential ablation). I assume that the difference between probing observations at successive times is representative of $b$, even though horizontal flow in the period between the two observations (2 - 10 m) results in a different snowpack and surface being probed. Complications with GPS radio transmission reduced my initial 16-point accumulation array to seven observations in one August to September period (2009).

I compare these estimates of vertical velocity made at fixed coordinates (Eulerian frame of reference) with those from the often-used geometric relation at the surface (Eqn (2.4)) made at a stake (Lagrangian frame of reference). I employ Eqn (2.4) with GPS measurements of $\dot{h}$ made at stakes, and $\alpha$ derived from a common DEM.

### 2.5. Error analysis

My error analysis assumes all compounded error terms are uncorrelated. Error estimates in my measurement of cumulative mass balance from glaciological and GPS methods likewise assume annual measurements are uncorrelated.

#### 2.5.1. Glaciological method

Many studies have reported a random error of $\pm 0.20$ m w.e. a$^{-1}$ for $b_a$, and this estimate is often taken to be a reliable estimate of $B_a$ uncertainty given the spatial autocorrelation of mass balance measurements (Cogley and Adams, 1998). Recent re-analysis of glaciological measurements finds that this error is $\pm 0.34$ m w.e. a$^{-1}$ (Zemp et al., 2013). A major source of error in glaciological mass balance measurements is the spatial variability of $b_a$ (e.g., Kaser et al., 2006). I quantify random error in my
Glaciological measurements from uncertainties in the measurements and their extrapolation. Random errors from stake measurements arise from the determination of the surface due to surface roughness and ablation caused by the stake and average ±0.10 m w.e. a⁻¹ (e.g., Huss et al., 2009); I adopt this error term as it accords with my observations.

Accumulation measurements rely on depth and density measurements in pits and depth measurements by probing. Measurement errors of snow depth include misidentification of the previous year's surface and determination of the undulating present-year surface; pit measurements are less problematic than soundings with a probe. Penetration by a probe into underlying firm would overestimate mass balance (Thibert et al., 2008). I found that misidentification of an overlying ice lens as the previous year's surface was as common as probing into the underlying firm, and treated it as a random error term instead of a bias. Deviation of a probe from vertical, however, overestimates snow depth and introduces a positive bias (Østrem and Haakensen, 1999). Over three balance years I probed snow depth at 98 sites; I probed snow depth in the four cardinal directions around each site. The average standard deviation of these 392 probing observations was 0.07 m ice eq. Many studies estimate the compounded random error of accumulation measurements (typically ±0.30 m w.e. a⁻¹) to be greater than those from the ablation zone, (e.g., Huss et al., 2009). However, my average standard deviation of 0.07 m ice eq. from four probing measurements at 98 locations suggests a reduced error. I use an error estimate of ±0.10 m ice eq. a⁻¹ for all stake, pit, and probing measurements of ablation or accumulation depth.
Errors in measurements of $b_a$ also arise from an assumption of the density of ice (900 kg m$^{-3}$) and the measurement of snow density in pits. I used a large 500 cm$^3$ tube core for the purpose of sampling snow within pits. Snow cutters have a typical measurement error of 11%, and the larger tube cutters are of higher precision (Conger and McClung, 2009). My field scale has a measurement error of ±3.3%, or ±10 grams for an average sample weight of 300 grams. I thus conservatively assume an error in my accumulation zone density measurements of 10%, or ±60 kg m$^{-3}$ for average conditions in my three years of study.

I therefore estimate error in $b_a$ as:

$$\sigma_{b_a} = \sqrt{\sigma_i^2 + \sigma_l^2 + \sigma_{\rho}^2}$$ (2.8)

where $\sigma_i$ is the estimated error in my measurements of the length of ablation and accumulation (±0.10 m ice eq. a$^{-1}$), $l$ is an area-weighted average ablation and accumulation measurement (2.0 m ice eq. a$^{-1}$), $\sigma_{\rho}$ is an area-weighted average of error in density assumptions and measurements expressed as a conversion factor (±0.04), and $\rho$ is an area-weighted density expressed as a conversion factor (0.72).

I calculate sampling error - extrapolation from $b(z)$ to $B_a$ via the hypsometry - by the standard deviation of the residuals between my observations and the linear spline. These residuals are 0.36, 0.37, and 0.25 m w.e. for balance years 2009, 2010, and 2011 respectively.

Error in planimetric area (Granshaw and Fountain, 2006; Bolch et al., 2010), defined as the sum of squared horizontal error in stereo model registration and digitizing
error (±5 pixels), yields ±0.3% for balance years 2009 and 2010, and ±0.6% for balance year 2011.

Measurement and extrapolation errors for each balance year are thus:

\[ \sigma_{Glac} = \sqrt{\sigma_{b_e}^2 + \sigma_{Ext}^2} \]  
(2.9)

where \( \sigma_{Glac} \) is the estimated error for glaciological \( B_a \), \( \sigma_{b_e} \) is measurement error, and \( \sigma_{EXT} \) extrapolation error.

2.5.2. Photogrammetric method

I quantify the uncertainty in \( dh \) using the standard deviation of elevation residuals of 275 points in 11 check patches (Figs. 2.1 and 2.3). These residuals reveal the combined error in my stereo models \((x,y,z)\) and my precision and accuracy in manually digitizing points \((z)\). I follow Rolstad et al. (2009) to assess uncertainty in sequential DEM analysis when the correlation range - the extent of spatial autocorrelation - is less than the averaging area:

\[ \sigma^2_A = \sigma^2_{\Delta z} \frac{1}{5} \frac{A_{cor}}{A} \]  
(2.10)

where \( \sigma^2_A \) is the variance of the spatially averaged elevation difference, \( \sigma^2_{\Delta z} \) is the variance of the elevation difference, \( A_{cor} \) is the correlation area, and \( A \) is the glacier surface area.

I calculate \( A_{cor} \) (Rolstad et al., 2009) as:

\[ A_{cor} = \pi a_1^2 \]  
(2.11)

where \( a_1 \) is the correlation range determined using the Incremental Spatial
Autocorrelation tool in ArcGIS 10.1, which gives the Global Moran’s I statistic over a series of increasing distances (e.g., Getis and Ord, 1992).

For all geodetic measurements of $B_a$ presented below, I convert to w.e. using my B scenario that assumes 900 kg m$^{-3}$ for all points below the ELA, and 750 kg m$^{-3}$ for all points above the ELA. To quantify error in this density assumption I use a range of possible density values, with the maximum density error defined by the A scenario, which assumes a density of 900 kg m$^{-3}$ for all points. The minimum density is defined by the C scenario, which assumes 900 kg m$^{-3}$ below the ELA and 600 kg m$^{-3}$ above the ELA. I consider these extrema as the plausible range of densities ($\pm 3\%$) and use these as a conservative estimate of density-assumption error.

I propagate the DEM error and density-assumption error as:

$$\sigma_{b_p} = \sqrt{\sigma_d^2 \rho^2 + \sigma_{\rho}^2 A^2}$$

where $\sigma_{b_p}$ is the error in my photogrammetric $B_a$ measurements (m w.e. a$^{-1}$), $\sigma_d$ is the DEM error (0.30 m ice eq.), $A$ is area-weighted surface thickness change (1.00 m ice eq. a$^{-1}$), $\sigma_{\rho}$ is my estimate of density-assumption error from a range of possible values expressed as a conversion factor ($\pm 0.09$), and $\rho$ is an area-weighted density expressed as a conversion factor (0.81).

To estimate error in my correction for differing observation dates, I use the standard deviation of the residuals of measured and modeled ablation from my mid-summer array measurements. The total number of array points is 71 from 2008, 2009, and 2010, but most of these points (64) are in the ablation zone. Residuals from all points have a mean of -0.01 m ice eq. and a standard deviation of 0.20 m ice eq.
However, the seven points from the accumulation zone have a residual mean of -0.24 m ice eq., which indicates there is a potential systematic bias in my date correction.

I estimate error in my date-corrected photogrammetric $B_a$ measurements ($\sigma_{\text{Phigrm}}$) as:

$$\sigma_{\text{Phigrm}} = \sqrt{\sigma_{B_p}^2 + \sigma_{\text{Corr}}^2}$$  \hfill (2.13)

where $\sigma_{B_p}$ is the error in my photogrammetric $B_a$ measurements (Eqn (2.12)), and $\sigma_{\text{Corr}}$ is the estimated error in my correction for differing observation dates.

2.5.2. GPS method

The random error in $dh$ arises from movement of the antenna attached to my pack, my stance on an uneven surface, and foot or ski penetration into a firn or snow surface. I thus adopt a conservative measurement error of ±0.10 m (e.g., Nolan et al., 2005), which is three times greater than the measurement error (±0.03 m) I observed by resurveying benchmarks.

For GPS measurements, I employ the same density assumptions and error estimates as my photogrammetric approach. As with my glaciological measurements, I estimate error in specific GPS measurements ($b_{\text{GPS}}$) as:

$$\sigma_{b_{\text{GPS}}} = \sqrt{\sigma_h^2 \rho^2 + \sigma_\rho^2 h^2}$$  \hfill (2.14)

where $\sigma_h$ is the estimated error in my measurements of thickness change (±0.10 m ice eq. a$^{-1}$), $h$ is an area-weighted average measurement of thickness change (-0.70 m ice eq. a$^{-1}$), $\sigma_\rho$ is an area-weighted average of error in density assumptions and measurements.
expressed as a conversion factor (±0.09), and $\rho$ is an area-weighted density expressed as a conversion factor (0.81).

GPS interpolation error is approximated by the standard deviation of the residuals between my observations and the linear spline. I use $±0.57$ m to estimate sampling error for all three balance years; this value arises from measurements made in 2011 which included the accumulation area.

The estimated error for the GPS method is:

$$
\sigma_{GPS} = \sqrt{\sigma_{GPS}^2 + \sigma_{Ext}^2}
$$

where $\sigma_{GPS}$ is the estimated error for glacier-wide GPS mass balance ($B_{GPS}$), $\sigma_{GPS}^2$ is measurement error, and $\sigma_{Ext}$ is extrapolation error.

2.6. Results

The glaciological method yields two years of negative mass balance, followed by a third year of mass gain, resulting in a cumulative balance for the period 2009-2011 of $0.10 \pm 0.63$ m w.e. (Table 2.2). The photogrammetric result for balance year 2009 (-0.15 ± 0.36 m w.e.) overlaps with the glaciological method (-0.12 ± 0.39 m w.e.). Photogrammetrically-based mass change for the periods 2009 - 2011 and 2010 - 2011, however, are more negative than those derived by the glaciological method (Table 2.2). My GPS-derived mass balance estimates for the year 2010 (-0.34 ± 0.57 m w.e.) and the period 2010-2011 (-0.44 ± 0.81 m w.e.) overlap with the glaciological method (respectively -0.31 ± 0.40 and -0.44 ± 0.26 m w.e.), but the GPS results for balance years 2009, 2011, and cumulative period 2009-2011 are more negative than either the
Table 2.2 Estimates of glacier-wide mass balance ($B_a$) by the three different methods.

<table>
<thead>
<tr>
<th>Balance Year</th>
<th>Glaciological (m w.e.)</th>
<th>Photogrammetric (m w.e.)</th>
<th>RTK GPS (m w.e.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2009</td>
<td>-0.12 ± 0.39</td>
<td>-0.15 ± 0.36</td>
<td>-0.82 ± 0.58</td>
</tr>
<tr>
<td>2010</td>
<td>-0.31 ± 0.40</td>
<td>no data</td>
<td>-0.34 ± 0.57</td>
</tr>
<tr>
<td>2011</td>
<td>0.53 ± 0.29</td>
<td>no data</td>
<td>-0.10 ± 0.57</td>
</tr>
<tr>
<td>2009 - 2011</td>
<td>0.10 ± 0.63</td>
<td>-0.59 ± 0.33</td>
<td>-1.26 ± 0.99</td>
</tr>
<tr>
<td>2010 - 2011</td>
<td>0.22 ± 0.49</td>
<td>-0.44 ± 0.26</td>
<td>-0.44 ± 0.81</td>
</tr>
</tbody>
</table>

glaciological or photogrammetric results (Table 2.2). For the period 4 August to 14 September 2009 – an additional period for which I made both glaciological and GPS measurements, with points distributed in both the accumulation and ablation zones – I find similar results of -1.02 ± 0.28 and -0.98 ± 0.28 m w.e.

I tested the reliability of using the reduced subset of photogrammetric points measured in 2011 ($n = 698, 74\%$) to represent glacier-wide elevation change by comparing $B_a$ for the year 2009 derived from 100\% of the points with a synthetically thinned subset of points matching the point locations of those measured in 2011. This comparison yielded $B_a$ of -0.15 m for both 100\% and 74\% coverage, suggesting that incomplete coverage of the glacier in 2011 is likely not the source of differences in the methods.

Using a subset of geodetic point measurements (e.g., via GPS) yielded results comparable to those from measurements across the entire glacier surface (e.g., photogrammetry) if measurements were made in all elevation bins. I measured $B_a$ as -0.08 ($n = 41$), -0.15 ($n = 56$), 0.02 ($n = 37$), and -0.10 m w.e. ($n = 61$) in four subsets, compared with -0.15 ± 0.36 m w.e. ($n = 937$) for the 100-m grid covering the entire glacier (Fig. 2.6). The subset with a slightly positive $B_a$ (0.02, $n = 37$) is the only subset
Figure 2.6 Four subsets (triangles) of the 2009 photogrammetric mass points (circles): a) Points on a longitudinal profile along the center of the glacier, b) points along safely walkable longitudinal profiles, c) 37 array-point locations, and d) points from an evenly-spaced grid. Values in the upper right corner of each panel indicate $B_3$ for the associated subset. Spatial locations of each subset are displayed in figure 2.3.

that does not include measurements in all elevation bins, with a data gap over the mid-glacier icefall (Fig. 2.3).

2.6.1. Error

The error ($\pm 0.35$ m w.e.) in the glaciological measurements is dominated by the spatial variability of $b_a$. The precision of my accumulation-zone measurements is $\pm 0.10$ m, three times more precise than Huss et al. (2009). The coefficient of variation (ratio of standard deviation to mean) of accumulation-zone $b_a$, however, is twice that of the
ablation zone. The large error term in my glaciological measurements thus arises from the high spatial variability of $b_a$ in the accumulation zone.

Errors in my photogrammetric methodology are dominated by the precision and accuracy of my manual measurements of photogrammetric mass points (DEM uncertainty), and number of independent samples in my sampling grid, which depends on the spatial autocorrelation of the surface elevation data. Photogrammetric measurements of $dh$ are five times less precise than either the glaciological or GPS data. Rolstad et al. (2009) found spatial correlation within DEMs generated by automated photogrammetry at three spatial scales: hundreds of meters, a few kilometers, and at tens of kilometers. They hypothesize that correlation at hundreds of meters was the result of matching errors in their automated methodology, whereas correlation at tens of kilometers was due to inaccurate georeferencing. Others (e.g., Nuth et al., 2007; Barrand et al., 2010) have relied on this study to justify an assumption of a correlation area of 1 km$^2$. These studies use automated techniques to investigate decadal change in contrast to my manual methods to assess annual change. Based on my assessment of correlation range, I estimated values of 1600, 700, and 1100 m in balance year 2009, and periods 2010-2011 and 2009-2011 respectively, resulting in correlation areas of 8.0, 1.5, and 3.8 km$^2$, which yield more conservative estimates of DEM error than if I had assumed a correlation area of 1 km$^2$.

Our errors in $B_a$ using a GPS arise from few upper-elevation measurements, and the lack of glacier-wide elevation measurements may bias $dh(z)$ (Figs. 2.5 and 2.7). I estimate interpolation error of ±0.57 m, based on the residuals between a two-piece linear
Figure 2.7  Same as Figure 2.5, but for two multi-year periods: 2009 – 2011 and 2010 - 2011.
spline and observations for 2011; a shorter monitoring period (August-September 2009) where I could make measurements in both the ablation and accumulation zones yielded a smaller interpolation error (±0.17 m). However, I cannot reliably determine $dh(z)$ for the GPS method given the lack of suitable accumulation zone measurements. Further investigation should be made using the GPS method with measurements well distributed across the entirety of a glacier's surface, and with the rover antenna on a stadia rod to maximize measurement precision.

To convert my geodetic measurements of $dh$ to $B_a$ m.w.e. I used a bulk density of $810 \pm 90$ kg m$^{-3}$, which is similar to the density of $850 \pm 60$ kg m$^{-3}$ recommended by Huss (2013). My estimated error due to assumed density (±0.05 to ±0.15 m w.e.) is lower than other error sources used in my geodetic estimates of mass change. Density errors annually vary due to the changes in extent of ablation and accumulation zones, and the magnitude of $dh$. However, I find error from density assumptions to be minimal on an annual basis.

I do not attempt to quantify errors due to advection of topography, but recognize that this process may inflate geodetic errors of $B_a$, especially in cases when the photogrammetric method does not sample the entire glacier surface. I observed topographic advection on a number of occasions as I navigated back to a point for re-measurement via GPS and found a crevasse where a relatively level surface existed previously. This same advection of surface features also likely plays a role in the apparent 'blunders' in my photogrammetric methodology, particularly over the minor, mid-glacier icefall.
2.6.2. Bias

All three methods used in this study differ in their sources of systematic bias. My glaciological measurements may include a positive bias from the omission of internal and basal mass balance. The photogrammetric measurements include a negative bias introduced by the manual operator and my temperature index model. In contrast, the GPS data do not include these biases.

The glaciological method suffers from potential biases associated with probing the previous summer surface, and no measurement of internal and basal mass-balance. My profiles of cumulative mass balance are more positive than cumulative values for my upper-most pit alone (Fig. 2.7), indicating a potential positive bias of the probing measurements. I use the average difference (0.07 m w.e.) between $B_d$ derived from all points and only from pits as an approximation of this potential bias.

Sinking (self drilling) of ablation stakes may produce a negative bias in the glaciological method (Riedel et al., 2010). I tested this possibility by measuring ablation at adjacent stakes, one with an insulated cap at the base of the stake and one without. No differences were noted, and I thus assume that self-drilling did not occur.

Glaciological measurements do not capture internal and basal mass balance. These mass changes are glacier dependent, with internal and basal accumulation playing a more significant role in cold, continental climates (e.g., Storglacieren, Sweden; Zemp et al., 2010), whereas internal and basal ablation dominates in warm, maritime climates (e.g., Franz Josef Glacier, New Zealand; Alexander et al., 2011). Castle Creek Glacier is a temperate, continental glacier and does not experience a high geothermal flux. Estimates of internal ablation for temperate glaciers vary dramatically from 1 cm w.e. a$^{-1}$.
Table 2.3 Varying photogrammetric $B_a$ results when using different melt factors in my temperature index model correcting for dates of photography. Percentages in parentheses indicate differences from results using Place Glacier melt factors.

<table>
<thead>
<tr>
<th>Balance Year</th>
<th>Place* (m w.e.)</th>
<th>Castle Creek' (m w.e.)</th>
<th>Combined† (m w.e.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2009</td>
<td>-0.15 ± 0.36</td>
<td>-0.19 ± 0.36 (-31%)</td>
<td>-0.11 ± 0.36 (+18%)</td>
</tr>
<tr>
<td>2010</td>
<td>no data</td>
<td>no data</td>
<td>no data</td>
</tr>
<tr>
<td>2011</td>
<td>no data</td>
<td>no data</td>
<td>no data</td>
</tr>
<tr>
<td>2009 - 2011</td>
<td>-0.59 ± 0.33</td>
<td>-0.58 ± 0.33 (+2%)</td>
<td>-0.60 ± 0.33 (-1%)</td>
</tr>
<tr>
<td>2010 - 2011</td>
<td>-0.44 ± 0.26</td>
<td>-0.36 ± 0.26 (+17%)</td>
<td>-0.47 ± 0.26 (-10%)</td>
</tr>
</tbody>
</table>

*Place Glacier melt factors: $k_s = 2.76$, $k_t = 4.67$ (mm w.e. °C^{-1} d^{-1})

†Castle Creek Glacier melt factors: $k_s = 3.45$, $k_i = 4.33$

‡Place Glacier snow, Castle Creek Glacier ice melt factors: $k_s = 2.76$, $k_i = 4.33$

(e.g., Thibert et al., 2008) to 2.5 m w.e. a^{-1} at lower elevations (Alexander et al., 2011). I conservatively estimate internal ablation and accumulation to respectively be 10% and 4% of $B_a$ (Zemp et al., 2010; Alexander et al., 2011). Combining these values with my probing bias, I estimate positive biases of 0.08, 0.09, and 0.10 m w.e. for the years 2009, 2010 and 2011 respectively, and 0.27 and 0.19 m w.e. for periods 2009-2011 and 2010-2011 respectively.

I corrected for stereo model bias in my photogrammetric measurements of $B_a$, but neglect densification and sub-glacial erosion, processes that I assume to be negligible in my study. To assess bias in the temperature index model I differenced modeled from observed ablation at array points for August 2008, 2009, and 2010. The mean of the residuals for 64 observations in the ablation zone is 0.02 m w.e., whereas it is -0.24 m w.e. for seven observations in the accumulation zone, suggesting a possible negative bias. However, this potential bias is perhaps related to error in my probing measurements at the seven accumulation-zone locations, and the limited number of observations. When comparing observed and modeled ablation measurements, I found melt factors derived
Figure 2.8 Frequency distributions of repeat-measurement residuals of 275 check points for stereo models from 2008, 2009, and 2011 aerial photography used to constrain potential error and bias of the analyst in manual photogrammetry. Each distribution displays surface-elevation residuals, defined as the difference in surface elevation of the same horizontal coordinates from the same stereo model and by the same operator, but from initial measurements and a repeat measurement at a later date.
from mass-balance measurements (Shea et al., 2009) yielded better agreement when using Place Glacier melt factors (-0.01 ± 0.20 m w.e.) than those from Peyto Glacier (+0.23 ± 0.28 m w.e.). I am hesitant to rely on Castle Creek Glacier melt factors given my short period of observation and few measurements for snow surfaces (n = 7). The use of melt factors derived from either Place of Peyto Glacier (Shea et al., 2009) does not substantially affect my estimated photogrammetric mass change (Table 2.3).

After correcting for stereo model bias, manual digitization can still introduce significant bias in photogrammetric measurements of $dh$ (McGlone et al., 2004). My repeated measurements from the same stereo models of 275 check points assess this bias (Fig. 2.8), and yield a potential bias of -0.26 ± 0.46, -0.34 ± 0.58, and 0.09 ± 0.74 m for stereo models from 2008, 2009, and 2011 respectively. The mean of all 825 points gives a bias of -0.15 m. These results indicate a potential manual-operator bias, but one that is inconclusive in terms of both sign and magnitude. Additionally, measurement uncertainty varies according to glacier surface characteristics. Repeat measurements of surface elevation from 2009 models yielded residuals of 0.01 ± 0.44, 0.16 ± 0.66, and -0.15 ± 1.44 m ice eq. for ice, firn and snow respectively. This error, however, did decrease with repeated measurements, revealing the importance of operator experience (Fig. 2.4). I found that GPS and photogrammetric point measurements of $dh$ accord in the ablation zone (Figs. 2.5A and 2.7).

I am unaware of other studies assessing the potential biases of GPS $B_a$ measurements, and I do not attempt to quantify GPS bias, which would require a complete understanding of a glacier's spatially varying vertical velocity.
The two geodetic methods give results that are comparable to, or significantly more negative than those of the glaciological method. These results indicate negative biases inherent in the geodetic method or positive biases in the glaciological method, similar to findings at South Cascade Glacier (Krimmel, 1999). Other studies detect no systematic difference between the two methodologies (Cogley, 2009; Fischer, 2011).

2.6.3. Estimation of vertical velocity

My estimates of ablation, \( dh \), and estimated vertical velocity for the ablation zone vary for the years 2008, 2009, and 2010 (Fig. 2.9). My method of estimating vertical velocity (Eqn (2.7)) yields values that average 0.30 m ice eq. more emergence than estimates from Eqn (2.4). GPS measurements at seven points in the accumulation zone in 2009 produce a submergence estimate of \(-0.38 \pm 0.15\) m ice eq.

I do not quantify errors in my estimate of vertical velocity. Using my approach, this error term will be dominated by advection of surface topography, but I lack a detailed estimate of the glacier’s surface roughness. Advection of an irregular surface (0.5-2 m roughness) will greatly exceed my estimated precision in measuring height change from ablation stakes or with a GPS (\( \pm 0.10 \) m ice eq.). I suspect that some of the variability in vertical velocity (Fig. 2.9) may arise from advection of topography.

2.7. Discussion and recommendations for future mass-balance monitoring

Limitations and advantages are inherent in each method to assess glacier mass change. Surface mass balance estimated by the glaciological method should be continued for index glaciers and potentially expanded for under-represented mountainous regions.
Figure 2.9 Spatial distribution of mass balance, thickness change, and vertical velocity from the 20-point ablation-zone array for the month of August in 2008, 2009, and 2010. All measurements in ice equivalent units. I derived mass balance from stake measurements, thickness change from RTK GPS, and vertical velocity as the difference of the two. Locations of stakes and GPS measurements are indicated by black dots.
A key benefit of the glaciological method is its ability to record \( b(z) \), a prerequisite for modeling the response of a glacier to changes in climate (e.g., Radić and Hock, 2011). Unfortunately, the glaciological method also suffers from a number of logistical shortcomings, which include the time and energy intensive nature of the method, and its limitation to glaciers that are relatively small and safe for travel.

I recommend increased use of the photogrammetric method to monitor \( B_a \). When GCPs already exist, fieldwork is not necessary, which can significantly reduce measurement time and cost. However, a lack of density measurements remains a source of error. Remote sensing geodetic methods afford the best possibility to monitor representative glaciers, including those that are large, complex and difficult to visit. Additionally, such methods enable the monitoring of more glaciers in a region, avoiding expensive and time intensive field studies.

Poor contrast, particularly for glaciers covered by fresh snow cover, limits the use of manual or automated feature extraction from aerial photography (e.g., Krimmel, 1999; Bamber and Rivera, 2007). Multispectral aerial photography or high-resolution, multispectral satellite imagery improves contrast in the accumulation zone of glaciers, however. The inclusion of the near infrared band in my 2011 stereo models, for example, enabled me to measure elevation for 74% of the glacier's surface that was covered by fresh snow.

I advocate the adoption of my GPS technique to measure \( B_a \). Previous studies concluded that geodetic measurements alone cannot be used to measure \( B_a \) in the absence of a knowledge of dynamics (e.g., Hagen et al., 2005), but I find that a well-distributed subset of point measurements can adequately mitigate the confounding role of vertical
velocity to yield reliable estimates of $B_a$. In the case of Castle Creek Glacier, I find a sample density of 4 km$^2$ can be used to derive $B_a$ using geodetic-grade GPS receivers (Fig. 2.6). If completed for a well-distributed subset of points across the glacier surface, the GPS method may circumvent some of the limitations inherent in the other two approaches, but more studies are required to quantify errors and bias inherent in the GPS approach.

Use of the *in situ* GPS method is restricted to those glaciers that are accessible, and safe for travel. Additionally, measurements yield $dh$, which is modulated by vertical velocity, making the GPS method ill-suited for determination of $b_a$ and $b(z)$ (c.f. Figs. 2.5 and 2.7). GPS is advantageous on larger glaciers where the glaciological method is impractical due to necessary commitments of time and energy. Use of the GPS method at the end of the accumulation season, when glacier surfaces are covered with snow, may enable travel and measurement on surfaces inaccessible at the end of the balance year. I recommend future efforts to refine the GPS method be undertaken at an established field station with a source of power adequate for the high-powered base station radio (35W).

Combining at-a-point glaciological and GPS measurements provide one method to assess the spatial and temporal changes of vertical velocity for a glacier. My methodology to estimate vertical velocity (Eqn (2.7)) compares well with that of the geometric relation at the surface (Eqn (2.4)). However, a major shortcoming is that it is field-intensive, whereas it is possible to employ the kinematic boundary condition remotely (e.g., Gudmundsson and Bauder, 1999). Furthermore, the use of an Eulerian frame of reference in this method imparts a potential error due to advection of topography, an error that is likely to be on the order of 0.5 m, but may be in excess of 5
m in areas of complex surface topography. Changes in ice flux have been found to be significant in determining recent thinning (Berthier and Vincent, 2012), and combining GPS and glaciological measurements allows insight into the fine-scale structure of seasonal ice dynamics. Additionally, my in situ methodology enables the estimation of submergence in the accumulation zone. Remote sensing studies, which derive submergence from measurements of horizontal motion and elevation change (Eqn (2.4)), often fail in the accumulation zone due to a lack of surface features to track to determine surface velocity. However, a study of the potential error imparted by advection of topography is required.

My future efforts at Castle Creek Glacier will include continued monitoring of annual length change (Beedle et al., 2009), and glaciological measurements of annual mass balance. Additionally, I plan to acquire aerial photographs of Castle Creek Glacier annually for geodetic measurements of annual and cumulative mass balance.

There is a need to continue long-term mass-balance measurements, resume interrupted series, expand to important regions and more representative glaciers, and improve error analysis (Fountain et al., 1999; Zemp et al., 2009). Geodetic methods provide a valuable measure of $B_a$ that is complementary to those of the glaciological method. Photogrammetric and GPS methods provide means to improve understanding of glacier change, and to help understand and predict the fate of mountain glaciers.
3. Annual Push Moraines as Climate Proxy

Publication details:

This chapter has been peer reviewed and is published in Geophysical Research Letters. Please see Appendix A: Authorship Statements for details of the contributions of each author.


3.1. Abstract

I reconstruct the terminus position of a mountain glacier in British Columbia, Canada from annual push moraines formed between 1959 and 2007. My reconstruction represents the longest, annually-resolved record of length change for a North American glacier. Comparison of annual recession with climate records indicates that glacier recession is controlled by air temperatures during the ablation season and accumulation season precipitation during the previous decade. Analysis among records of glacier frontal variation and mass balance in western North America similarly reveals an immediate terminus reaction to summer and net mass balance and a delayed reaction to winter and net balance. Other mountain ranges may contain long series of push moraines
that could be exploited as climate proxies, and to improve understanding of glacier response to climate.

### 3.2. Introduction

Mountain glaciers provide fresh water to millions of people, and contribute to global sea level rise (Barnett et al., 2005; Meier et al., 2007). Records of glacier terminus fluctuations and mass balance provide insight into how climate affects this important freshwater source. Meier et al. (2007) estimate there are 300,000 to 400,000 mountain glaciers and small ice caps on Earth, but length change and mass balance records exist for only 1,800 and 230 glaciers, respectively. Only 39 of these glaciers have records that exceed 30 years in length (Zemp and van Woerden, 2008), and there is a strong European bias in these records.

I use push moraines to reconstruct the longest, annually-resolved record of terminus position for a North American glacier. Glaciers form annual push moraines during the accumulation season when forward movement of the glacier snout exceeds ablation, resulting in a seasonal advance (Bennett, 2001). Formation of a push moraine at the glacier margin requires a deformable till sheet. Preservation of annual moraines requires ablation season recession to be greater than advance of the terminus during the following accumulation season. To my knowledge, all studies of annual push moraines are for maritime glaciers with high mass-balance gradients in either Iceland (Sharp, 1984; Boulton, 1986, Krüger, 1995; and Bradwell, 2004) or Norway (Andersen and Sollid, 1971; and Worsley, 1974).
3.3. Study area and methods

I reconstructed the frontal position of Castle Creek Glacier (53° 2' N., 120° 24' W., unofficial name), British Columbia (BC), Canada from a continuous series of push moraines that front the glacier (Fig. 3.1). The glacier has an area of 9.4 km², a length of 5.85 km, and an elevation range of 2,827 to 1,810 m.

To determine the age of the moraines, I mapped glacier terminus position for 10 dates between 1946 and 2005 from orthorectified aerial photographs (Table 3.1). I photogrammetrically scanned aerial photograph negatives from the Canadian and BC governments with a ground sampling resolution of ≤ 1.0 m. To orthorectify these images, I used common ground control points and the 25 m BC Terrain Resource...
Table 3.1 Descriptive data for the 10 aerial photos used in this study. 'A' denotes federal photography, accessed from the National Air Photo Library, Natural Resources Canada, Ottawa, Ontario. 'BC' denotes provincial photography, Crown Registry and Geographic Base, Victoria, British Columbia.

<table>
<thead>
<tr>
<th>Date</th>
<th>Roll ID</th>
<th>Nominal scale</th>
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<td>2005-08-25</td>
<td>BCC05111</td>
<td>1:20,000</td>
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</tbody>
</table>

Information Management Program digital elevation model (DEM). Root mean squared error of these horizontal control points was ≤1.6 m and typically below 1.0 m. I surveyed push moraines that formed after my latest aerial photographs (2005), and terminus position at the end of the ablation season in 2007 and 2008 with a geodetic-grade global positioning system.

Annual push moraine position indicates a prior maximum glacier extent achieved at the end of the accumulation season, while glacier terminus position is mapped from imagery acquired at the end of the ablation season. Thus, I dated the push moraines immediately down-valley of the known terminus position as being formed at the end of the previous accumulation season. The number of intervening moraines coincides with the years between consecutive images, confirming the annual nature of the moraines. I represent changes in glacier length as the total area between moraines or mapped terminus position divided by the curvilinear width. This method integrates recession across the entire glacier terminus and accounts for retreat along an irregular glacier margin.
To examine the climatic controls on mass changes and related frontal variation, I compared glacier length change records derived from annual push moraines with homogenized climate station records of air temperature and precipitation (Vincent, 1998; Mekis and Hogg, 1999) from the Prince George, BC climate station, ~180 km to the northwest (Fig. 3.1). For my study, I define ablation season as April - September and accumulation season as October - March. I compared un-lagged and lagged total accumulation season precipitation to assess the role of total accumulation during the following ablation season and a delayed terminus reaction to a precipitation signal integrated over the glacier surface.

To evaluate the representativeness of length change records derived from annual push moraines, I compared the record from Castle Creek Glacier to annual and seasonal mass balance series and glacier length change records from western North America (Tables 3.2 and 3.3). I used cross-correlation analysis to investigate the lagged relation between seasonal ($b_w$ and $b_a$) and annual ($b_a$) mass balance and frontal variation. McClung and Armstrong (1993), Laumann and Nesje (2009), and Winkler et al. (2009) compared $b_a$ with frontal variation, but I am unaware of previous analysis that considers the relation between frontal behavior and seasonal mass balance.

The bed slope near the glacier terminus can affect frontal reaction of a glacier as the position of the terminus depends on ice velocity and the product of bed slope and ablation near the snout (Nye, 1965; Boulton, 1986). To investigate the potential role that bed gradient plays as a control of annual glacier length change, I compared the frontal recession for a given year to the average bed slope obtained between two consecutive push moraines. I used two DEMs to assess slope: the 25 m DEM used for image
Table 3.2 Correlation table of western Canada and Pacific Northwest U.S.A. glacier frontal variation (FV) and mass balance ($b_s$ and $b_d$) time series. Bold values indicate significance ($p < 0.05$). Number of paired observations ($n$) displayed on lower half of table. See Table 3.3 for varying years of observation of each time series.

<table>
<thead>
<tr>
<th></th>
<th>Castle FV</th>
<th>Blue FV</th>
<th>S. Casc. FV</th>
<th>Blue $b_s^2$</th>
<th>Helm $b_s^1$</th>
<th>Helm $b_s^1$</th>
<th>Peyto $b_s^3$</th>
<th>Peyto $b_s^3$</th>
<th>Place $b_s^1$</th>
<th>Place $b_s^1$</th>
<th>S. Casc. $b_s^4$</th>
<th>S. Casc. $b_s^4$</th>
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<td>29</td>
<td>13</td>
<td>38</td>
<td>36</td>
<td>41</td>
<td>27</td>
<td>47</td>
<td></td>
</tr>
</tbody>
</table>

Data sources:

* Record of frontal variation derived from annual push moraines, this paper

1 World Glacier Monitoring Service (http://www.wgms.ch)

2 University of Washington (http://www.geophys.washington.edu/Surface/Glaciology/projects/blue_glac/blue.html)

3 [Demuth et al., 2009]

4 United States Geological Survey (http://ak.water.usgs.gov/glaciology)
Table 3.3 Years of observation for the records of glacier frontal variation and mass balance used in this study. The record 'Castle FV' has been derived from push moraines (this paper). The abbreviation 'FV' denotes frontal variation and 'ba' and 'bs' represented annual balance and summer balance respectively.

<table>
<thead>
<tr>
<th>Glacier Record</th>
<th>Years of Observation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Castle FV</td>
<td>1959-2007</td>
</tr>
<tr>
<td>Blue ba</td>
<td>1956-1999</td>
</tr>
<tr>
<td>Place ba</td>
<td>1965-2007</td>
</tr>
<tr>
<td>Place bs</td>
<td>1965-1989, 1994-1995</td>
</tr>
<tr>
<td>S. Casc. ba</td>
<td>1959-2005</td>
</tr>
<tr>
<td>S. Casc. bs</td>
<td>1959-2005</td>
</tr>
</tbody>
</table>

orthorectification, and one derived from about 15,000 points collected with a differential GPS in kinematic mode.

3.4. Results

3.4.1. Annual push moraines

The forefield contains a continuous series of push moraines that represent the annual terminus position from 1959 to 2007. These deposits sit on low relief till sheets with little intervening bedrock and have been largely protected from fluvial erosion (Fig. 3.1). The height of the push moraines varies from 0.07 – 1.10 m and averages 0.33 m, while moraine width ranges from 0.61 – 4.39 m and averages 1.90 m. These dimensions are similar to push moraines described in Iceland (Sharp, 1984; Boulton, 1986; Krüger, 1995).

3.4.2. Glacier length change

Castle Creek Glacier receded 701 m between 1959 and 2007, averaging 14.3 m a⁻¹ with a minimum of 3 to 4 m a⁻¹ in the mid- to late-1970s. Recession rates accelerated to
over 40 m a\textsuperscript{-1} in the early 1990s (Fig. 3.2). The earliest aerial photographs indicate 185 m of recession between 1946 and 1959, an average rate of 14.2 m a\textsuperscript{-1} that accords with the period 1959-2007. Whereas some glaciers in western North America advanced in the 1960s and 1970s (Luckman et al., 1987), Castle Creek Glacier did not advance in any year since 1959. The slowed recession during the mid-to-late 1970s, however, indicates that this glacier responded to climatic conditions that caused many North American glaciers to advance.

3.4.3. Relation to climate

Average ablation season temperature at Prince George correlates with the annual recession from 1959 to 2007 ($r = 0.55, p < 0.001, n = 49$) indicating that annual glacier length change is partly controlled by summer temperature. There is an inverse relation between annual recession and March precipitation of the current year ($r = -0.42, p < 0.01, n = 49$). A physical mechanism to account for this correlation is the degree to which late-lying snow can alter the energy balance at the glacier terminus; years of light snow cover during spring lengthen the ablation season and vice versa. Annual recession is correlated to accumulation season precipitation, and this correlation is strongest when precipitation lags terminus response by a decade ($r = -0.38, p < 0.01, n = 49$). I interpret this delayed reaction of the terminus as the time required for the glacier to begin adjusting its length to changes in upper-elevation mass, which is partly controlled by accumulation season precipitation.

Multiple linear regression reveals that 40% of glacier length change variance is explained by ablation season temperature, March precipitation, and lagged accumulation...
Figure 3.2 Annual and cumulative frontal variation (1959 to 2007) of Castle Creek Glacier derived from annual push moraines (grey). Annual and cumulative length changes of South Cascade Glacier, Washington (red) are displayed for comparison.

season precipitation. Ablation season temperature is the most important model term ($p < 0.001$), followed by March precipitation ($p < 0.01$) and lagged accumulation season precipitation ($p < 0.1$).

Years of extreme recession in the early-1990s (Fig. 3.2) significantly influence the relation between annual glacier fluctuation and climate variables. Omission of the year 1992 in the multiple regression analysis, for example, results in 46% of variance explained, with lagged accumulation season precipitation the most important model term ($p < 0.001$), followed by ablation season temperature ($p < 0.01$), and March precipitation ($p < 0.05$). Removing additional years with anomalously high rates of recession (1991 and 1993) does not further alter the relations.
Residuals from both multiple regression models, however, are autocorrelated, and cumulative departures of these residuals reveal prominent shifts in the late-1970s and late-1980s. Hypothesizing that large-scale atmospheric circulation plays a prominent role in driving annual recession, I used stepwise linear regression with the climate station variables in the previous models, as well as climatic indices that have been shown previously to be important drivers of glacier mass balance in northwest North America (Rasmussen and Conway, 2004). I find annual indices of the Southern Oscillation Index (SOI) (Trenberth, 1984) and Pacific Decadal Oscillation (PDO) (Mantua et al., 1997) to correlate with Castle Creek Glacier recession, during the current year and when lagged by a decade. When data from 1992 are included in regression analysis, 45% of annual recession is explained by Prince George March precipitation ($p < 0.01$), annual average PDO lagged 10 years ($p < 0.01$), Prince George summer temperature ($p < 0.05$), annual average PDO ($p < 0.1$), and annual average SOI ($p = 0.1$). When data from 1992 are excluded in regression analysis, 55% of annual recession variance is explained by Prince George accumulation season precipitation lagged 10 years ($p < 0.01$), Prince George March precipitation ($p < 0.01$), annual average PDO ($p < 0.05$), annual average PDO lagged 10 years ($p < 0.05$), Prince George summer temperature ($p < 0.1$), and annual SOI ($p < 0.1$).

### 3.4.4. Relation to Other Mass Balance and Frontal Variation Records

The 49-year record accords with frontal variation, $b_s$, and $b_n$ of glaciers in western Canada and Washington, U.S.A. (Fig. 3.1; Table 3.2). My record correlates with frontal variation of South Cascade and Blue glaciers; both glaciers are over 500 km to the south
of Castle Creek Glacier. Interestingly, the recession rates for all three glaciers tripled in the early 1990s (Fig. 3.2; Blue Glacier not shown). Castle Creek Glacier frontal variation correlates with $b_s$ of Peyto, Place, and South Cascade glaciers and is also correlated to $b_a$ of Blue, Helm, Peyto, Place, and South Cascade glaciers (Table 3.2).

Cross correlation analysis of the Castle Creek Glacier frontal variation record and the $b_s$ record of Place Glacier reveals that both records are correlated at zero lag, while the relation between frontal behavior and $b_w$ is highest when Place Glacier $b_w$ leads the record from Castle Creek Glacier by 13 years (Fig. 3.3). A similar pattern is apparent when the frontal variation is compared to Place Glacier $b_a$, as net balance is the sum of winter and summer balance, revealing both an immediate (zero lag) as well as a lagged relation. This immediate relation between frontal variation and $b_s$ and $b_a$, and lagged reaction to $b_w$ and $b_a$ is also found for South Cascade Glacier frontal variation and mass balance records, but only an immediate relation exists between Blue Glacier frontal variation and $b_a$. I find no relation between the length change record and average bed slope angle between moraines from either DEM.

3.5. Discussion and conclusions

I show that high-resolution imagery can be used to identify and map annual push moraines in glacier forefields. Series of moraines can be dated, and records of annual glacier length change can be reconstructed when repeat imagery is available. To my knowledge, this study is the first to use push moraines as climate proxies outside of the maritime environments of Iceland or Norway. A preliminary search of glacier forefields
Figure 3.3 Cross correlation analysis of Castle Creek Glacier frontal variation versus Place Glacier seasonal and annual mass balance. Solid circles denote significance ($p < 0.05$). Solid lines are second-order polynomial curves fitted to the series of lagged correlation coefficients. The common period of analysis is 1965 to 2007 for $b_a (n = 43)$, and 1965 to 1989; 1994 to 1995 for $b_w$ and $b_s (n = 27)$.

in the mountains of western Canada using aerial photographs reveals that annual push moraines are common, but series of moraines that exceed a decade in length are rare. Other mountain ranges, however, may contain long series of push moraines that should be exploited as climate proxies, and to examine glacier response to climate.

Nye (1965) concluded that modeling the mass balance history of South Cascade Glacier and Storglaciaren, Sweden, from length change records was unreliable, primarily due to the errors in measuring frontal variation. Records of glacier frontal variation derived from push moraines avoid some of the problems that confront measurements obtained through other methods. Glacier length change measurements from well-preserved series of push moraines allow averaging of recession across much of a
terminus and, as push moraines faithfully preserve accumulation season maximum extent, they avoid the problem of measurements that do not coincide with the end of the ablation season.

Summer temperature is an important control of annual length change for Castle Creek Glacier; this finding agrees with previous work on push moraines and glacier length change (Sharp, 1984; Krüger, 1995; Bradwell, 2004; and Sigurdsson et al., 2007). Accumulation season precipitation and $b_a$ also affect annual recession but with a delay that ranges between two to five years (Salinger et al., 1983; Sigurdsson et al., 2007; Laumann and Nesje, 2009; Winkler et al., 2009). For Castle Creek Glacier, I observe that this delay is on the order of a decade.

Concordance of Castle Creek Glacier fluctuations with frontal variation, $b_s$, and $b_a$ of other glaciers in western North America suggests a common climatic driver. Due to the spatial variability of precipitation, it is likely that ablation season temperatures have been the dominant control of glacier recession for these glaciers in recent decades. This conclusion supports the findings that recent glacier recession in western North America and the Northern Hemisphere is primarily caused by anomalously warm temperatures during the ablation season (Rasmussen and Conway, 2004). The synchronous and rapid retreat of Castle Creek, South Cascade, and Blue glaciers in the early 1990s is noteworthy. Clearly, regional climatic conditions drove this retreat.

Climate, glacier geometry, and ice dynamics all influence the terminus position of alpine glaciers. My empirical data indicate a delay of about a decade between changes in $b_w$ and the reaction of the Castle Creek Glacier terminus. This delay differs from the theoretical response time of a glacier to approach a new steady state after a change in
climate. Response time can be approximated as the quotient of mean ice thickness and the net ablation rate at the terminus (Jóhannesson et al., 1989). Using area-volume scaling (Bahr, 1997) to derive average ice thickness (67 m) and recent measurements of net balance rate at the terminus (-3.4 to -4.0 m a\(^{-1}\)), I calculate the response time to be approximately 17-20 years for Castle Creek Glacier. Delay of a decade for accumulation season precipitation defines when the glacier's response to precipitation is maximized, while the method described by Jóhannesson et al. (1989) approximates an approach to the end of terminus response; thus maximum terminus reaction to precipitation is delayed by a decade but theoretical equilibrium response is not achieved for another 10 years. My analysis indicates glacier termini begin to immediately react to ablation season temperatures and \( b_s \). This reaction is superimposed on a longer frontal reaction which is partly forced by past \( b_w \) and filtered by flow dynamics.

Long, continuous records of annual glacier length change form an integral part of regional and global glacier monitoring. Such records can supplement the sparse records of glacier mass balance for remote mountain regions. Annually resolved records of terminus position provide an important dataset to better understand the relation between climate and glacier fluctuations. Measuring seasonal balance and annual frontal variation of the same glacier should be a renewed focus of glacier monitoring as these empirical data provide a means to refine theories proposed for the frontal response time of glaciers.
4. Glacier change in the Cariboo Mountains, British Columbia, Canada (1952-2005)

Publication details:

This chapter has not been peer reviewed; it is in preparation for submission for review and publication in The Cryosphere.

4.1. Abstract

I calculated dimensional change for 33 glaciers in the Cariboo Mountains of British Columbia for the latter half of the twentieth century. All glaciers receded during the period 1952-2005, averaging a surface area loss of $-0.19 \pm 0.05\% \text{ a}^{-1}$. From 1952 to 1985, nine glaciers advanced. Following 1985, rates of recession doubled to $-0.41 \pm 0.12\% \text{ a}^{-1}$. Thinning rates likewise accelerated, from $-0.14 \pm 0.04 \text{ m w.e. a}^{-1} (1952-1985)$ to $-0.50 \pm 0.07 \text{ m w.e. a}^{-1}$ for the period 1985-2005. Temperatures increased from the earlier to the latter period for the ablation (+0.38°C) and accumulation (+0.87°C) seasons, and average precipitation decreased, particularly in the accumulation season (-32 mm, -3.2%). Both increased temperature and decreased precipitation are likely climatic drivers of observed glacier recession and thinning. I show that previous studies may have overestimated recession of Cariboo Mountains glaciers by about 50% due to the presence of late-lying snowcover in aerial photography from the 1980s. My comparison of surface area change with glacier morphometry corroborates previous studies that show primary relations between extent change and surface area. I also find, however, that such
relations were not consistent temporally, with reduced significance or even changing sign for different periods and climatic conditions.

4.2. Introduction

Glaciers are important components of the hydrologic system. They contribute meltwater along with organic and inorganic materials to freshwater systems (Bogdal et al., 2009; Hood et al., 2009; Moore et al., 2009; Marshall et al., 2011). These inputs, along with downstream impacts, are altered when glaciers undergo changes in extent and volume. As glaciers and ice caps (GIC; all glaciers outside of the Antarctic and Greenland ice sheets) lose volume, sea level rises (Radić and Hock, 2011). The largest GIC contributions to sea level rise include the Canadian Arctic Archipelago and Gulf of Alaska regions (Gardner et al., 2011 and Berthier et al., 2010), but recent work demonstrates that, collectively, small glaciers (<1 km²) constitute a significant amount to total GIC volume (Bahr and Radić, 2012).

Recent studies of glacier change use a variety of remotely-sensed products that include aerial photography (Koblet et al., 2010; Tennant et al., 2012), satellite imagery (Paul et al., 2004; Berthier et al., 2010), and laser altimetry (Sapiano et al., 1998; Arendt et al., 2002). Typically these studies employ a combination of geomatic data to assess changes in glaciers for a particular region (Schiefer et al., 2007; Abermann et al., 2009; Bolch et al., 2010). Comprehensive inventories of glacier extent and change (e.g., Bolch et al., 2010), and measurement of glacier volume change of large regions (e.g., Schiefer et al., 2007) rely heavily on satellite borne instruments. Such analysis is thus limited to the last 30 years. Air photos afford the ability to extend our ability to document glacier
change for up to three decades prior to the beginning of the satellite era. In this study I use aerial photogrammetry to investigate the extent and volume change of a subset of glaciers in western Canada for the periods 1952-1970, 1970-1985, and 1985-2005. The use of aerial photography thus provides a method to temporally extend the glacier inventory studies of Schiefer et al. (2007) and Bolch et al. (2010).

The objectives of this study are to: 1) investigate the extent and volume change of a subset of glaciers in the Cariboo Mountains of British Columbia (BC); and 2) assess the climatological conditions that could explain observed glacier change in the study area.

4.2.1. Study area and previous work

The Cariboo Mountains are the northernmost range of the Columbia Mountains of BC, Canada (Fig. 4.1). The climate of the Cariboo Mountains is transitional, wetter than the Rocky Mountains to the east, and drier than the Coast Mountains. Total annual precipitation (1971-2000 climate normals) averages about 1014 and 679 mm on the windward (Barkerville) and lee (McBride) sides of the Cariboo Mountains, and annual temperature averages 1.9 and 4.4°C respectively (Environment Canada, 2012).

Annual length change of Castle Creek Glacier in the Cariboo Mountains is more closely related to changes of glaciers in the Coast Mountains than those of the Rocky Mountains (Beedle et al., 2009). Meltwater from the majority of glaciers in the Cariboo Mountains contributes to the Fraser River; however, some glaciers in the Premier Range (a Cariboo Mountains subrange) contribute to the Columbia River (Fig. 4.1). Maurer et al. (2012), using sediment cores and detrital and in situ fossil wood, found that glaciers of the Cariboo Mountains nearly reached their Holocene maximum extents around 2.73-
Figure 4.1 Study area: the three red rectangles indicate subregions in this study (Castle, Quanstrom, and Premier) and the location of the three maps in Figure 4.2.
2.49 ka and that major retreat of the glaciers did not begin until the early 20th century. Luckman et al. (1987) found that some glaciers of the Premier Range advanced in the 1960s and 1970s in response to decreased temperatures and increased winter precipitation.

Bolch et al. (2010) completed the first comprehensive inventory of glaciers in western Canada, identifying 536 glaciers in the Cariboo Mountains (2005) with a total surface area of 731 km². Cariboo Mountains glaciers lost 7.06 km³ of ice at a rate of -0.58 m water equivalent (w.e.) a⁻¹ during the period 1985-1999 (Schiefer et al., 2007).

4.3. Methods

4.3.1. Imagery and supplemental data

I measured glacier extent and surface elevation change from aerial photographs obtained from the British Columbia Government and Canada National Air Photo Library (Table 4.1). Imagery prior to 2005 was scanned at a resolution of 12-14 μm from diapositives or negatives using a photogrammetric scanner whereas imagery from 2005 was available as digital aerial triangulation (AT) scans (digital photos with available exterior orientation). Ground sampling distance ranges from 0.2 to 1.1 m depending on scanning resolution and image scale. Most photos were taken in late summer, but dates range from mid-July to late-September. I used all images to estimate glacier length and area change, whereas only those from 1946, 1952, 1984, 1985, and 2005 were used to determine surface-elevation change. As photos for each subregion were taken in different
Table 4.1  Aerial photography used to derive glacier extent and thickness change data.

<table>
<thead>
<tr>
<th>Date</th>
<th>Region</th>
<th>Roll ID</th>
<th># of Images</th>
<th>Scale</th>
<th>Contrast</th>
<th>Snowcover</th>
</tr>
</thead>
<tbody>
<tr>
<td>1946-09-10</td>
<td>C</td>
<td>BC320</td>
<td>15</td>
<td>1:31,680</td>
<td>Poor</td>
<td>Good</td>
</tr>
<tr>
<td>1952-09-25</td>
<td>C</td>
<td>A13538</td>
<td>5</td>
<td>1:60,000</td>
<td>Fair</td>
<td>Good</td>
</tr>
<tr>
<td>1952-09-09</td>
<td>Q</td>
<td>A13538</td>
<td>5</td>
<td>1:60,000</td>
<td>Fair</td>
<td>Good</td>
</tr>
<tr>
<td>1955-08-30</td>
<td>P</td>
<td>A14930</td>
<td>7</td>
<td>1:70,000</td>
<td>Poor</td>
<td>Fair</td>
</tr>
<tr>
<td>1967-08-15</td>
<td>C</td>
<td>BC7019</td>
<td>15</td>
<td>1:15,840</td>
<td>Good</td>
<td>Fair</td>
</tr>
<tr>
<td>1970-08-19</td>
<td>P</td>
<td>BC5394</td>
<td>5</td>
<td>1:80,000</td>
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<td>Good</td>
</tr>
<tr>
<td>1971-07-15</td>
<td>Q</td>
<td>A21587</td>
<td>4</td>
<td>1:80,000</td>
<td>Poor</td>
<td>Fair</td>
</tr>
<tr>
<td>1977-07-24</td>
<td>C</td>
<td>A24743</td>
<td>8</td>
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<td>Poor</td>
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<td>C</td>
<td>BC84073</td>
<td>4</td>
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<td>Fair</td>
</tr>
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<td>6</td>
<td>1:20,000</td>
<td>Good</td>
<td>Good</td>
</tr>
</tbody>
</table>

Regions:  C: Castle Creek, Q: Quanstrom, P: Premier Range.
Roll ID:  BC: Provincial aerial photos, A: Federal aerial photos
Contrast:  Subjective determination of image contrast on a scale from poor to excellent.
Snowcover:  Subjective determination of snowcover extent. Poor refers to more snowcover.

years I report a mean area-weighted year: 1952 for photos taken in 1946, 1952, or 1955; 1970 for those of 1967, 1970, or 1971; and 1985 for those of 1984 or 1985. All annual rates of extent or volume change are calculated using the actual duration between images.


I use ClimateWNA (Wang et al., 2012) to assess the variability in temperature and precipitation during the periods for which I measure glacier extent and surface-elevation change. The ClimateWNA v4.72 program (http://www.genetics.forestry.ubc.ca/cfcg/ClimateWNA/ClimateWNA.html) downscales PRISM (Daly et al., 2002) and historical data (Mitchell and Jones, 2005) for western North America, to generate monthly time series of temperature and precipitation for the Cariboo Mountains. I use ClimateWNA output for coordinates of a location that is central to both my subset and the Cariboo Mountains (glacier 31, Fig. 4.2), along with the
median elevation of my subset of glaciers (2470 m). To test the spatial variability of ClimateWNA output, I compare data for this one location with averages of points along the center-line every 100 m in elevation for all 33 glaciers ($n = 212$), and for glaciers of the Castle Creek region ($n = 47$), Quanstrom region ($n = 62$), and Premier region ($n = 103$). I compare monthly temperature and precipitation data from ClimateWNA with measurements at Castle Creek Glacier (Déry et al., 2010) for 2009-2011 to estimate the ability of ClimateWNA to represent conditions at high elevations in the Cariboo Mountains.

To investigate dominant synoptic patterns over the Cariboo Mountains I use the National Centers for Environmental Protection/National Center for Atmospheric Research (NCEP/NCAR) reanalysis data (Kalnay et al., 1996) of 700 hPa geopotential height anomalies. I use 700 hPa geopotential height as an indicator of general conditions in the middle troposphere (~3000 m), an elevation just above the uppermost extent of most glaciers in the Cariboo Mountains. I assess climatic conditions for balance years (October-September), and for ablation and accumulation seasons, defined as June-September and October-May, respectively.

4.3.2. Photogrammetry

I used the Vr Mapping photogrammetry software suite (Cardinal Systems LLC) to create stereo models from aerial photos. Exterior orientation for the models was derived from tie points and common ground control points (GCP) horizontally and vertically distributed throughout the photos (Schiefer and Gilbert, 2007; Barrand et al., 2009). Within each subregion I used common GCPs, which helps reduce systematic positional
errors (Kääb & Vollmer, 2000; Schiefer and Gilbert, 2007; Schiefer et al., 2007). GCPs consist of stable bedrock features or boulders and were obtained from stereo models created from 2005 AT scans.

Eleven check patches, located on stable surfaces around each glacier region (Fig. 4.2), were used to quantify the relative accuracy of each block of stereo models (Fig. 4.3). Each check patch is comprised of 25 individual checkpoints in a five-meter grid; I estimated systematic bias of stereo models from the mean residuals of the check patches. Trend surfaces were created from these 11 mean residuals and used to apply a correction for glacier surface-elevation measurements.

4.3.3. Glacier subset selection

With the exception of 2005, glaciers of the Cariboo Mountains were not comprehensively photographed in any given year, and I thus concentrate my study on three subregions of the Cariboo Mountains with suitable photographic coverage: 1) the Castle Creek Glacier area; 2) the Quanstrom Mountain area; and 3) the Premier Range (Figs. 4.1 and 4.2). I selected my subset of glaciers based on three primary considerations: 1) availability of imagery, both in terms of temporal resolution and spatial scale; 2) snow cover and contrast; and 3) representativeness of all Cariboo Mountains glaciers based on average glacier morphometry as determined by the comprehensive inventory completed by Bolch et al. (2010). Snowcover and poor
Figure 4.2 Cariboo Mountains subregions showing subset of 33 glaciers, including extents for 1952, 1970, 1985, and 2005. Panels show from left to right the Castle, Quanstrom, and Premier subregions. Numerical glacier identification, numbered by 2005 surface area from smallest to largest, corresponds to that of Tables 4.2 and 4.4.
Figure 4.3  Box-and-whisker plots display the maximum, interquartile range, median, and minimum of surface elevation residuals of 275 checkpoints. Checkpoints are used to determine relative accuracy of stereo models and bias correction for measurement of surface elevation change of glaciers in the Castle and Quanstrom regions in three periods.

contrast of older aerial photography reduce potential subregions (Table 4.1). Sun angle with respect to glacier aspect and slope leads to local areas of high reflectivity and an absence of contrast, particularly for some south-facing glaciers. I omitted glaciers with these issues in my selection of a subset.

Within the three Cariboo Mountains subregions, I selected a subset of 33 glaciers representing five size classes (0.1-0.5, 0.5-1.0, 1.0-5.0, 5.0-10.0, and >10.0 km²). From the Bolch et al. (2010) inventory I used the percent total surface area within each size class for all Cariboo Mountains glaciers as a guide for selecting my glacier subset. I used surface area as my primary criterion to select glaciers as many studies find it to be the key morphometric determinant of glacier extent change (e.g., Serandbrei-Barbero et al., 1999; Hoelzle et al., 2003; Paul et al., 2004; Andreassen et al., 2008; Bolch et al., 2010; Paul and Andreassen, 2009). My analysis oversampled glaciers in the largest size class.
as these glaciers likely play a dominant role in regional volume change and meltwater contributions to their respective watersheds (e.g., Arendt et al., 2006; DeBeer and Sharp, 2007; Paul and Haeberli, 2008).

4.3.4. Data collection and analysis

I manually collected glacier extents and point measurements of surface elevation directly from the stereo models. Complete glacier outlines were mapped for 2005, and glacier extents were updated for previous years only below the transient snowline, thus eliminating errantly mapping seasonal snow cover as glacierized area (e.g., Koblet et al., 2010). Delineation of planimetric glacier area above the transient snowline is problematic as seasonal snow masks the glacier margin, and can inflate measurements of glacier surface area and extent change. Mapping of ice divides relied on surface elevation and surface features such as crevasses and runnels, and these divides were held constant in order to compare dimensional changes for a given ice body through time. Extent-change analysis was made for different glacier subsets and periods based on aerial photo coverage and snowcover. Collectively, my analysis allows me to compare area change for 33 glaciers for the periods 1952-1985, 1985-2005, and 1952-2005, and for 26 glaciers in the additional periods 1952-1970, and 1970-1985.

Poor contrast and snowcover limit my analysis of glacier surface elevation change to seven glaciers for the periods 1952-1985, 1985-2005, and 1952-2005. For these glaciers I measured surface elevation on a 100 m grid for glaciers larger than 1 km² and on a 50 m grid for glaciers smaller than 1 km². Poor contrast in the earliest photographs inhibited data collection over portions of glacier accumulation zones for four of the seven
glaciers. In these areas I extrapolated from the three highest elevation bins where I have measurements, assigning the average surface-elevation change of these observations to the bins with missing values. This assumption was necessary only for periods that rely on the earliest year of photography (1952-1985 and 1952-2005), and for less than 0.4% of total surface area for two glaciers, but up to 27% for another two.

To calculate glacier-wide volume change from point measurements, I multiplied average elevation change for each 50 m elevation bin and summed over the entire glacier surface. Elevation bins and glacier surface were derived from epoch-specific glacier hypsometries created from the same point measurements. Where poor contrast led to an absence of measurements in the accumulation zone, I used surface-elevation measurements for that elevation bin from the stereo model of a prior or subsequent year. This assumes that surface elevation and extent underwent negligible change, an assumption that would be inappropriate in the rapidly changing ablation zones, but one that is more likely in the accumulation zone where this assumption was applied. I calculated glacier wide average thickness change based on the average of the two extents that define a given period (e.g., Arendt et al., 2002; Barrand et al., 2010). All values presented in water equivalent (w.e.) are calculated by assuming a density of 900 kg m$^{-3}$ in the ablation zone and 750 kg m$^{-3}$ in the accumulation zone (e.g., Beedle et al., in press); ablation and accumulation zones are defined by each glacier’s median elevation (Table 4.2).

To estimate regional (Cariboo Mountains) glacier extent change, I extrapolated from measurements of my subset of studied glaciers using the average rates of relative
Table 4.2 Morphometric properties of the 33-glacier subset

<table>
<thead>
<tr>
<th>Id (Name)</th>
<th>Region</th>
<th>Length (m)</th>
<th>Area (km²)</th>
<th>Slope</th>
<th>Aspect</th>
<th>Mean Z (m)</th>
<th>Median Z (m)</th>
<th>Min. Z (m)</th>
<th>Max. Z (m)</th>
<th>Range Z (m)</th>
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<td>2587</td>
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<td>2827</td>
<td>946</td>
</tr>
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<td>2565</td>
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<td>3357</td>
<td>1708</td>
</tr>
<tr>
<td>30 (David) P 5228</td>
<td>P</td>
<td>5228</td>
<td>13.533</td>
<td>19.3</td>
<td>SW</td>
<td>2609</td>
<td>2616</td>
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<td>1987</td>
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</table>

Slope and elevation data from TRIM DEM and 2005 extent.
Length and area data from 2005 extent.
Table 4.3 Estimated error in extent for each year and extent change for three successive periods and two cumulative periods.

<table>
<thead>
<tr>
<th>Year</th>
<th>Error (%)&lt;sup&gt;a&lt;/sup&gt;</th>
<th>Period</th>
<th>Error (%)&lt;sup&gt;b&lt;/sup&gt;</th>
</tr>
</thead>
<tbody>
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<td>1952</td>
<td>2.8</td>
<td>1952-1970</td>
<td>3.5</td>
</tr>
<tr>
<td>1970</td>
<td>2.4</td>
<td>1970-1985</td>
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</tr>
<tr>
<td>1985</td>
<td>2.2</td>
<td>1985-2005</td>
<td>2.3</td>
</tr>
<tr>
<td>2005</td>
<td>0.8</td>
<td>1952-1985</td>
<td>3.6</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1952-2005</td>
<td>2.9</td>
</tr>
</tbody>
</table>

<sup>a</sup> Mean glacier extent error from individual glacier buffers and image resolution

<sup>b</sup> Mean RMSE of individual glacier error estimates for extents in two years

extent change by size class for different periods. I used two methods to estimate Cariboo Mountains surface-elevation change: first, the average surface-elevation change of my subset of seven glaciers; and second, the average gradient of surface-elevation change with elevation, defined as the average of all measurements within each 50-meter bin, and integrated with the Cariboo Mountains glacier hypsometry of 1985. I compare extrapolated values with results from recent work that included volume change (Schiefer et al., 2007) and extent change (Bolch et al., 2010) of Cariboo Mountains glaciers.

4.3.5. Error analysis

I estimated error in glacier extent and extent change based on a buffer surrounding each glacier, a method first proposed by Granshaw and Fountain (2006). My buffer of ± five pixels gave buffer widths of ~5.5 to 1.3 m, depending on image resolution. Error in extent change is calculated as the root mean squared error (RMSE) of the error estimates for the years that define a given period. Average errors range from 0.8 - 2.4% for measurements of glacier extent, and from 2.3 - 3.6% for extent change
(Table 4.3). These errors vary depending on image resolution (see Table 4.1) and glacier dimensions.

I used the standard deviation of 275 check points in 11 check patches to estimate error in my measurements of surface-elevation change \( E_{AZ} \). To account for greater uncertainty over surfaces with reduced contrast, I added an error of ± 5 m for measurements above the transient snow line; where measurements are absent due to poor contrast and values are extrapolated, I increased this error to ± 10 m (e.g., Tennant and Menounos, 2013). Error in measurements of surface-elevation change in the ablation zone is 1σ of check points, whereas for the accumulation zone it is the quadrature sum of 1σ of check points \( E_1 \) and the added error of ± 5 or 10 m for reduced contrast \( E_2 \):

\[
E_{AZ} = \sqrt{E_1^2 + E_2^2}
\]  

(4.1)

Converting to water equivalent units through density assumptions imparts an additional error term. To estimate error in the density assumption of 750 kg m\(^{-3}\) for the accumulation zone I used a range of possible values for accumulation-zone density (600-900 kg m\(^{-3}\)). I estimate error in my measurements of surface-elevation in water equivalent units \( E_{AZw,e} \) as:

\[
E_{AZw,e} = \sqrt{E_{AZ}^2 \rho^2 + E_{\rho}^2 \Delta Z^2}
\]  

(4.2)

where \( \rho \) is density expressed as a water-equivalent conversion factor (0.9 in the ablation zone, 0.75 in the accumulation zone), and \( E_\rho \) is the error in my density assumptions, assumed to be 0 in the ablation zone and 0.15 in the accumulation zone.
I estimate volume change error ($E_{VOL}$) after Barrand et al. (2010) as:

$$E_{VOL} = \sqrt{\sum_{1}^{\text{bin}} (E_{\text{bin}} A_{\text{bin}})^2}$$  \hspace{1cm} (4.3)

where $E_{\text{bin}}$ is the error ($E_{\Delta Z e.v.}$) and $A_{\text{bin}}$ is the surface area of each 50 m bin.

Previous work indicated a spatial correlation of 1000 m in photogrammetric DEMs (Rolstad et al., 2009), and others have calculated a number of independent measurements ($n$) assuming a correlation scale of 1000 m (Barrand et al., 2010). I calculated degrees of freedom for each glacier from point measurements of surface-elevation change in epochs with complete spatial coverage. Using the Incremental Spatial Autocorrelation tool in ArcGIS 10.1 I calculated correlation distances that range from 350 to 2,000 m, yielding effective degrees of freedom that range from 1 to 5.

I use cross-validation to test the accuracy of my extrapolation of measured glacier extent and surface-elevation change to the unmeasured glaciers. In my test of glacier extent extrapolation I partition the 26-glacier subset into four scenarios where I withhold 75, 50, and 25% of the individual glaciers as a validation set and retain the remainder as a training set. Set selection is randomized within each size class, ensuring each set retains glaciers of varying size. Each of the four scenarios is applied to observations in five epochs: 1952-1970, 1970-1985, 1985-2005, 1952-1985, and 1952-2005. I use the standard deviation of these 20 test cases ($\pm 4.7\%$) as an estimate of error in my extrapolation of the extent of all Cariboo Mountains glaciers. To estimate error in my extrapolation of surface-elevation change, I partition my seven glacier subsets into two scenarios where I withhold three and four individual glaciers as a validation set and retain...
Figure 4.4 Fraction of total glacierized extent by size class for all Cariboo Mountains glaciers (gray) and my 33-glacier subset (black).

The remainder as a training set. Each scenario is applied to observations in three epochs: 1952-1985, 1985-2005, and 1952-2005. I use the variance of these six test cases to estimate error in my extrapolation of surface-elevation change using two methods: 1) mean annual rate of surface elevation change of the glacier subset (± 0.077 m w.e. a⁻¹); and 2) mean surface-elevation change within each collective 50 m elevation bin for the glacier subset (± 0.089 m w.e. a⁻¹).
4.4. Results

4.4.1. Regional representativeness

As expected, my glacier subset undersamples ice masses of the Cariboo Mountains in all size classes except the largest size class (Fig. 4.4). The bulk of Cariboo Mountains glacier surface area (40%) is comprised of glaciers in the 1.0-5.0 km² size class, whereas only 20% of my subset is of similarly sized glaciers. Glaciers in the largest size class dominate the total surface area of my subset (61%), whereas only 21% of the total surface area of Cariboo Mountains glaciers is comprised of similarly sized glaciers. Eliminating the four of the five largest glaciers from my 33-glacier subset
Figure 4.6 Fraction of total surface area by aspect for all Cariboo Mountains glaciers (black) and my 33-glacier subset (gray).

(glaciers 30-33; Table 4.4) results in a size-class distribution that is nearly identical to that of all glaciers in the Cariboo Mountains.

My glacier subset also oversamples high elevations (Fig. 4.5). Area-altitude distribution of glaciers in the Premier range, for example, are higher than the average elevation of glaciers in the Cariboo Mountains whereas the hypsometry of my glacier subset from the Castle and Quanstrom areas are much more similar to the total glacierized area of the Cariboo Mountains.

In terms of percent of total surface area, glaciers of the study contain a high frequency of north, northeast, and southwest-facing glaciers (Fig. 4.6). In particular the subset inadequately samples glaciers with east and west aspects. Cariboo Mountains glaciers have an average slope of 21.1° whereas my subset has an average slope of 18.7°; a difference in means, however, that is not statistically significant.
Table 4.4 Glacier extent and area change data for glaciers and summed by subregion and size class.

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<th>Extent Successive periods</th>
<th>Cumulative periods</th>
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<td>Area 1970 (km²)</td>
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</tr>
<tr>
<td>24</td>
<td>Q</td>
<td>4.440 4.582</td>
<td>4.555</td>
</tr>
<tr>
<td>25</td>
<td>Q</td>
<td>5.584 No data</td>
<td>5.136</td>
</tr>
<tr>
<td>26</td>
<td>Q</td>
<td>8.119 7.428</td>
<td>7.424</td>
</tr>
<tr>
<td>27</td>
<td>C</td>
<td>11.257 10.348</td>
<td>10.206</td>
</tr>
<tr>
<td>28</td>
<td>P</td>
<td>13.296 13.303</td>
<td>13.253</td>
</tr>
<tr>
<td>29</td>
<td>Q</td>
<td>13.711 13.821</td>
<td>13.394</td>
</tr>
<tr>
<td>30</td>
<td>P</td>
<td>18.392 17.008</td>
<td>16.796</td>
</tr>
<tr>
<td>31</td>
<td>P</td>
<td>18.234 17.042</td>
<td>17.661</td>
</tr>
<tr>
<td>32</td>
<td>C</td>
<td>21.283 20.466</td>
<td>20.038</td>
</tr>
<tr>
<td>33</td>
<td>C</td>
<td>34.960 35.201</td>
<td>35.237</td>
</tr>
</tbody>
</table>

Note: Values in bold indicate glacier advance. Volumes in italics indicate recession (beyond margin of error).

**Extents summed by size class are based on 2005 extent.**

**Summed extents and area change may omit some individual glaciers:** All: 18, C: 5 and 18, Q: none; P: 2, 0.1-0.5: none; 0.5-1.0: 6 and 18, 1.0-5.0: none; 5.0-10: none; 10.0-50: none.
4.4.2. Extent change

Surface area of the 33-glacier subset shrunk by 15.11 ± 1.89 km² (-10.6 ± 2.9%), from 141.96 ± 1.84 km² to 126.85 ± 0.47 km² from 1952-2005 (Table 4.4). Relative area change varied by subregion, with losses of -11.2 ± 2.7%, -16.4 ± 2.0%, and -4.2 ± 3.9% for the Castle, Quanstrom, and Premier regions respectively. Average area change by size class over the same period varied from -6.2 ± 3.3% (0.5-1.0 km²) to -13.9 ± 1.1% (5.0-10 km²) and individual glacier area change varied from -0.5 ± 1.6 to -31.7 ± 1.4%, with 22 glaciers receding beyond margin of error, and 11 glaciers with no discernable change (within margin of error). Absolute extent change is dominated by three of the largest glaciers (glaciers 28, 31, and 33), which comprise 50% of the total subset extent change for the period.

While glaciers on average shrank during the period 1952-2005, their change during intervening epochs was more complex (Table 4.4 and Fig. 4.7). Over the period 1952-1970, for example, five of 26 glaciers advanced, seven receded, and 14 did not significantly change (change did not exceed error margin). Between the years 1970 and 1985, two of 26 glaciers advanced, eight shrank, and the area of 16 glaciers did not change. For the cumulative period 1952-1985, with data for 32 glaciers, nine glaciers advanced, 12 receded, and 11 had no discernable extent change.

Glaciers that advanced were generally shorter, smaller, steeper, and had higher median and minimum elevations than glaciers that receded (Table 4.5). Glaciers that negligibly changed also tended to be steep, but were longer than those glaciers that advanced. Glaciers that changed little were also higher and flowed over a greater
Figure 4.7 Box-and-whisker plots displaying the maximum, interquartile range, median, and minimum of average annual rates of relative area change for (A) 32 glaciers for three periods, and (B) for 26 individual glaciers for four periods. Average annual rates are calculated using actual duration between imaging for each glacier, however this duration differs by subregion (Table 4.1).

After 1985 glaciers throughout the region shrank (Fig. 4.7). For 33 glaciers the average rate of area change increased from -0.05 ± 0.10% a\(^{-1}\) from 1952-1985, to -0.41 ± 0.12% a\(^{-1}\) from 1985-2005. Rates of area change averaged +0.03 ± 0.18% a\(^{-1}\) for 1952-1970, -0.08 ± 0.13% a\(^{-1}\) for 1970-1985, and -0.40 ± 0.10% a\(^{-1}\) from 1985-2005. During the period 1985-2005, 30 glaciers receded, while the area of two glaciers did not change. Total surface area change of the period 1985-2005 was -8.39 ± 1.51 km\(^2\), or -6.2 ± 2.3% (Table 4.4).

For the period 1985-2005 I directly compare extent change of 28 glaciers that were measured in my study and the Bolch et al. (2010) inventory. For the same subset of glaciers, Bolch et al. (2010) report 2005 and 1985 extents which are respectively about 2 and 5% larger than the outlines mapped from the aerial photography of this study. The
Table 4.5 Median glacier morphometry for glaciers that underwent some advance, receded continuously, or exhibited no discernable change (within margin of error) during the period 1952-1985.

<table>
<thead>
<tr>
<th></th>
<th>Count</th>
<th>Length (m)</th>
<th>Area (km²)</th>
<th>Slope (°)</th>
<th>Median Z (m)</th>
<th>Min. Z (m)</th>
<th>Max. Z (m)</th>
<th>Range Z (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Advance</td>
<td>9</td>
<td>1552</td>
<td>0.653</td>
<td>21.6</td>
<td>2449</td>
<td>2169</td>
<td>2733</td>
<td>617</td>
</tr>
<tr>
<td>Retreat</td>
<td>12</td>
<td>3032</td>
<td>3.640</td>
<td>15.2</td>
<td>2403</td>
<td>2075</td>
<td>2774</td>
<td>733</td>
</tr>
<tr>
<td>No Change</td>
<td>11</td>
<td>1949</td>
<td>1.406</td>
<td>20.8</td>
<td>2605</td>
<td>2210</td>
<td>2941</td>
<td>794</td>
</tr>
</tbody>
</table>

Bolch et al. (2010) inventory reports a loss from these 28 glaciers which is about 52% larger than those of this study.

The relation between area change and glacier morphometry temporally varies (Table 4.6). For three periods - 1952-1985, 1985-2005, and 1952-2005 – absolute area change correlates \( (p < 0.05) \) negatively with glacier length and area, indicating that the larger glaciers lost more surface area in all periods. Correlation coefficients between relative area change and both length and area vary from the early to latter period, with negative \( r \)-values of -0.235 and -0.210 for length and area for 1952-1985, and positive relations of 0.328 \( (p < 0.10) \) and 0.316 \( (p < 0.10) \) for 1985-2005. This variability indicates that larger glaciers lost more relative area in the earlier period, whereas small glaciers lost more relative area in the latter period. Average glacier surface slope is correlated with absolute and relative area change in all periods, except for relative area change from 1985-2005, indicating steeper glaciers generally lost less absolute and relative surface area, except during the latter period when there was no correlation. Median glacier elevation is correlated with absolute area change for the period 1952-1985 \( (r = 0.358, p < 0.05) \) and relative area change for the period 1952-2005 \( (r = 0.487, p < 0.05) \) indicating that glaciers with more surface area at higher elevations experienced less area change in these two periods.

Extrapolating from the Bolch et al. (2010) inventory of 2005 \( (731 \text{ km}^2) \) using relative area change of my 26-glacier subset and based on area loss per size class, I
Table 4.6 Correlation table (Pearson product-moment) of relations of glacier area change with respect to glacier morphometry for three periods.

<table>
<thead>
<tr>
<th></th>
<th>Length</th>
<th>Area</th>
<th>Slope</th>
<th>Median Z</th>
<th>Min Z</th>
<th>Max Z</th>
<th>Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>1952-1985 Absolute Area Change</td>
<td>-0.556</td>
<td>-0.650</td>
<td>0.598</td>
<td>0.358</td>
<td>-0.065</td>
<td>-0.215</td>
<td></td>
</tr>
<tr>
<td>1952-1985 Relative Area Change</td>
<td>-0.235</td>
<td>-0.210</td>
<td>0.518</td>
<td>0.323</td>
<td>0.135</td>
<td>0.053</td>
<td>-0.046</td>
</tr>
<tr>
<td>1985-2005 Absolute Area Change</td>
<td>-0.596</td>
<td>-0.765</td>
<td>0.523</td>
<td>0.260</td>
<td>0.513</td>
<td>-0.193</td>
<td>-0.393</td>
</tr>
<tr>
<td>1985-2005 Relative Area Change</td>
<td>0.328</td>
<td>0.316</td>
<td>-0.060</td>
<td>0.282</td>
<td>-0.364</td>
<td>0.412</td>
<td>0.431</td>
</tr>
<tr>
<td>1952-2005 Absolute Area Change</td>
<td>-0.597</td>
<td>-0.731</td>
<td>0.582</td>
<td>0.312</td>
<td>0.427</td>
<td>-0.131</td>
<td>-0.309</td>
</tr>
<tr>
<td>1952-2005 Relative Area Change</td>
<td>-0.008</td>
<td>0.011</td>
<td>0.440</td>
<td>0.487</td>
<td>-0.087</td>
<td>0.310</td>
<td>0.221</td>
</tr>
</tbody>
</table>

Correlation coefficients in bold are significant (α = 0.05)
n = 32 for the periods 1950s-1985 and 1985-2005
n = 33 for the period 1950s-2005

estimate that in 1952, the Cariboo Mountains contained a glacierized area of 824 ± 39 km². By 1970 glacier cover declined to 805 ± 38 km², and the area of glacier cover continued to shrink to 785 ± 37 km² by 1985. Results extrapolated from my 32-glacier subset give nearly identical results to those of the 26-glacier subset, with estimated Cariboo Mountains surface areas differing by only 0.4 and 0.01% in 1952 and 1985 respectively.

4.4.3. Volume change

Over the period 1952-1985, the seven glaciers of my study lost -0.480 ± 0.051 km³ w.e. Glacier loss during the first period (1952-1985) was -0.195 ± 0.059 km³ w.e., and loss accelerated to -0.345 ± 0.048 km³ w.e. during the later period (1985-2005). These volume losses correspond to respective thinning rates of -0.14 ± 0.04 m w.e. a⁻¹ and -0.50 ± 0.07 m w.e. a⁻¹ (Table 4.7 and Fig. 4.8). Net thinning rates over the period 1952-2005 averaged -0.230 ± 0.02 m w.e. a⁻¹. Two glaciers, both of which advanced over the same period, thickened markedly from 1952-1985 (Table 4.7). Both of these glaciers thinned after 1985, but experienced little net change from 1952-2005. Five glaciers continuously thinned and receded.
Table 4.7 Average annual thickness change in meters of water equivalent of seven glaciers for three periods.

<table>
<thead>
<tr>
<th>Glacier Id</th>
<th>Thickness change 1952-1985 (m w.e. a(^{-1}))</th>
<th>Thickness change 1985-2005 (m w.e. a(^{-1}))</th>
<th>Thickness change 1952-2005 (m w.e. a(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.56 ± 0.15</td>
<td>-0.87 ± 0.15</td>
<td>-0.04 ± 0.06</td>
</tr>
<tr>
<td>11</td>
<td>0.27 ± 0.07</td>
<td>-0.57 ± 0.08</td>
<td>-0.00 ± 0.04</td>
</tr>
<tr>
<td>19</td>
<td>-0.09 ± 0.16</td>
<td>-0.82 ± 0.22</td>
<td>-0.39 ± 0.14</td>
</tr>
<tr>
<td>24</td>
<td>-0.10 ± 0.22</td>
<td>-0.12 ± 0.22</td>
<td>-0.20 ± 0.18</td>
</tr>
<tr>
<td>26</td>
<td>-0.02 ± 0.25</td>
<td>-0.41 ± 0.30</td>
<td>-0.14 ± 0.21</td>
</tr>
<tr>
<td>28</td>
<td>-0.08 ± 0.06</td>
<td>-0.59 ± 0.08</td>
<td>-0.22 ± 0.04</td>
</tr>
<tr>
<td>31</td>
<td>-0.24 ± 0.20</td>
<td>-0.44 ± 0.23</td>
<td>-0.27 ± 0.17</td>
</tr>
</tbody>
</table>

Figure 4.8 Box-and-whisker plots display the maximum, interquartile range, median, and minimum of average annual thickness change in meters of water equivalent of seven glaciers for three periods.
Extrapolation of these thinning rates to unmeasured glaciers in the Cariboo Mountains yields volume losses of $-4.027 \pm 2.044$ and $-7.580 \pm 1.167$ km$^3$ w.e. for the periods 1952-1985 and 1985-2005 respectively. I obtain lower estimates of volume change ($-2.312 \pm 2.170$ and $-6.242 \pm 1.315$ km$^3$ w.e) using elevation-averaged surface elevation change measurements from these seven glaciers.

### 4.4.4. Relation to climate

Temperatures generally increased during the period of study, particularly after 1985 (Fig. 4.9). From the earlier period (1952-1985) to the latter (1986-2005) temperatures increased by +0.38 and +0.87 °C for the ablation and accumulation season respectively.

Accumulation season and annual precipitation decreased during the period of study, but little change occurred during the ablation season (Fig. 4.10). From the earlier period (1952-1985) to the latter (1986-2005), precipitation decreased by -32.27 (-3.2%) and -43.57 mm (-3.0%), respectively for the accumulation season and annually.

Geopotential height (700 hPa) anomalies, based on 1952-2005 mean fields, reveal an area of persistent low pressure in western Canada in both the accumulation and ablation seasons for the periods 1952-1970 and 1971-1985 (Fig. 4.11). These low-pressure anomalies were centered over the Yukon Territory and northern BC from 1952-1970, and diminished slightly in the period 1971-1985 with a prominent trough along the Coast Mountains (ablation season) and diffuse area of anomalously low pressure over much of central North America (accumulation season). A marked change in circulation occurred for the later period with anomalously high pressure present during the period...
Figure 4.9 ClimateWNA temperature records of deviation from the long-term (1952-2005) mean of (A) average accumulation season temperature, (B) average ablation season temperature, and (C) average annual temperature. Gray bars and values indicate average deviation from the long-term mean for the periods 1952-1970, 1970-1985, and 1985-2005.
Figure 4.10 ClimateWNA precipitation records of deviation from the long-term (1952-2005) mean of (A) total accumulation season precipitation, (B) total ablation season precipitation, and (C) total annual precipitation. Gray bars and values indicate average deviation from the long-term mean for the periods 1952-1970, 1970-1985, and 1985-2005.
Figure 4.11 Geopotential height (700 hPa) anomalies (m) for the accumulation and ablation seasons for three epochs. Color scale shows magnitude of anomaly. Data for the period 1952-2005 defines the mean fields. Dashed and solid contours respectively denote negative and positive anomalies that are significantly different ($p = 0.05$) from the mean at a given grid point.

1986-2005, with ridges over south central Alaska (ablation season) and off the coast of northern BC and southeastern Alaska (accumulation season).

These synoptic conditions accord broadly with the Pacific Decadal Oscillation (PDO; Mantua et al., 1997), a pattern of Pacific climate variability that persists for multiple decades and plays a prominent role in determining glacier mass balance in northwest North America (e.g., Bitz and Battisti, 1999). The earlier two periods (1950-
1970 and 1971-1985), with lower pressure and reduced glacier recession were largely coincident with a cool phase of the PDO from 1947 to 1976. In contrast, the latest period (1985-2005), with higher pressure and increased glacier recession, coincided with a warm phase of the PDO from 1977 to at least the mid-1990s.

ClimateWNA time series (1952-2005) of monthly temperature and precipitation vary little for a single central location at a median glacier elevation compared with averages of multiple points representing glacier surfaces across the region \(n = 212\). Output from this one location explains 99% of the variance of temperature and 97% of the variance of precipitation of the average of multiple points. ClimateWNA monthly temperature output averages -1.3 °C colder than measurements from two meteorological stations at Castle Creek Glacier for the period 2009-2011. This underestimation of temperature is greater in the ablation season (-1.8 °C) than the accumulation season (-0.8 °C), and also greater at the upper elevation site (2105 m) near the lateral margin of Castle Creek Glacier (-1.5 °C) than the lower elevation site (1803 m) in the glacier forefield (-1.1 °C). Only three of 34 months of precipitation data collected at Castle Creek Glacier compare well (within 15%) of Climate WNA output. No comparison can be made between ClimateWNA precipitation output and measurements at Castle Creek Glacier where failure or gross underestimation of in situ measurements is likely.

4.5. Discussion

4.5.1. Regional representativeness

My glacier subset was a compromise between regional representativeness and the availability of suitable aerial photography. I oversampled large glaciers as they will
contribute most to volume change for a given region. While glaciers of the 1-5 km$^2$ size class dominated glacier volume change in the Cariboo Mountains for the period 1985-2005, I hypothesize that losses from large glaciers (>5 km$^2$) may have been more important for previous epochs. The two largest glaciers for which I calculate volume change, for example, accounted for 94% of the total volume change of seven glaciers from 1952-1985, but 75% from 1985-2005. Selection of a representative subset of glaciers for a region is difficult, especially when glacier change is assessed over multiple epochs.

Most studies of glacier change note increased scatter of percent area change for smaller glaciers (Serandbrei-Barbero et al., 1999; Kääb et al., 2001; Paul, 2002; Paul et al., 2004; DeBeer and Sharp, 2007; Andreassen et al., 2008; Bolch, et al., 2010; Paul and Andreassen, 2009). This scatter may arise from the influence of local, topographic factors (DeBeer and Sharp, 2007; Paul and Andreassen, 2009). Kääb et al. (2001) concluded that this scatter could also be the result of using low-resolution satellite imagery. They note that this scatter may partly arise from the rapid response times of small glaciers resulting in glacier extent change over relatively rapid (a few years) changes in climate. Additionally, I hypothesize that late-lying seasonal snow may inordinately impact the inferred relative area change of small glaciers, and could play a role in the observation of increased scatter for smaller glaciers. To illustrate this effect I compared the scatter (standard deviation) in relative area change of the smallest glaciers as measured for 28 glaciers in both the Bolch et al. (2010) inventory and in my study, which is less affected by late-lying seasonal snow. The standard deviation of relative
area change from 1985-2005 for the Bolch et al. (2010) inventory is 8.1% for the smallest glaciers (<1.0 km²), whereas in my study it is 5.8%.

Previous studies suggest that glaciologists need to further examine intra- and inter-regional mass balance variability (Braithwaite, 2002), and conclude that there are no means of making an a priori selection of a regionally-representative benchmark glacier (Fountain et al., 2009). The paucity of traditional mass-balance and glacier length change measurements limits assessment of regional representativeness. However, widespread use of remotely sensed imagery to derive glacier extents and surface elevations affords an understanding of the spatial variability of glacier extent and volume change on decadal timescales. I demonstrate that comprehensive glacier inventories and assessments of decadal area and thickness change can be used to make an informed a priori selection of a representative subset of glaciers. However, regional representativeness will be variable with respect to glacier morphometry, climate and glacier response. Selection of benchmark glaciers and representative glacier subsets, as well as testing of the regional representativeness of existing benchmark glaciers (e.g., Fountain et al., 2009) should be based on regional glacier morphometry, recent decadal glacier extent and thickness change, and with respect to varying glacier response in different climatic regimes.

4.5.2. Area change

My results indicate net recession of Cariboo Mountains glaciers from 1952-2005, consistent with other studies of glacier change in BC (DeBeer and Sharp, 2007; Brewis, 2012; Tennant et al., 2012; Tennant and Menounos, 2013). While net glacier change of
Figure 4.12  Box-and-whisker plots display the maximum, interquartile range, median, and minimum percent area change for 28 glaciers from 1985 to 2005 for this study and that of Bolch et al. (2010).

my 33-glacier subset over this period is negative (-15.11 ± 1.89 km² or -10.6 ± 2.9%), for 11 of these glaciers recession was within my margin of error. Nine of these 11 glaciers advanced during the period 1952-1985.

My finding of glacier advance in the period 1952-1985 accords with the advance in the Premier Range discussed by Luckman et al. (1987) and with glacier advance elsewhere in BC during this period (Koch et al., 2009; Menounos et al., 2009; Brewis, 2012; Tennant and Menounos, 2013). Studies that did not include imagery from some year between ~1970 and 1985, and only determined net recession over a longer period (e.g., DeBeer and Sharp, 2007; DeBeer and Sharp, 2009; Jiskoot et al., 2009) will not capture this glacier advance. Similarly, work using decadally-spaced images (e.g., this
study) may miss shorter temporal events, such as the minor, short-lived advances documented by Luckman et al. (1987) using images with sub-decadal regularity.

After 1985, recession is dominant and average rates of recession increased eightfold versus the period 1952-1985, and fivefold over the period 1970-1985 (Fig. 4.7). Annual recession of Castle Creek Glacier (glacier 28) increased from about -10 m a\(^{-1}\) in the early-to-mid 1980s to about -40 m a\(^{-1}\) in the early-1990s (Beedle et al., 2009) and Tennant and Menounos (2013) also found increased recession rates for glaciers of the Columbia Icefield of BC and Alberta, which increased after 1979 and again after 2000.

I directly compare glacier extent and extent change (1985-2005) of the Bolch et al. (2010) inventory for 28 glaciers where my mapped ice divides match (Fig. 4.12). This inventory relies on glacier extents mapped by the British Columbia Government Terrain Resource Inventory Management (TRIM) program from 1980s aerial photography and Landsat imagery from 2003-2007. The 2005 extents of this inventory, created through a semi-automated process, average 2% larger than my extents manually digitized from aerial photographs with much higher resolution. However, the Bolch et al. (2010) inventory uses an area-weighted date (2005) as all areas of western Canada were not imaged in the same year, and the actual date of imagery covering the Cariboo Mountains is from 2006. Relative area change for the period 1985-2005 averaged ~0.3% a\(^{-1}\) for this 28-glacier subset, and thus the Bolch et al. (2010) inventory, based on 2006 imagery, will not differ significantly from my measurements from 2005 aerial photos. This provides evidence of the quality of the 2005 inventory and the precision of the method used by Bolch et al. (2010). However, the 1985 TRIM extents, digitized manually by the province of British Columbia from the same aerial photographs I use here, average 5%
larger than my extents. Late-lying snow is prevalent in the 1984 and 1985 photographs, but reference to additional photographs in years with a higher snowline helped determine snowfields from glacierized area. The TRIM mapping included some late-lying snow as glacier, resulting in overestimation of 1985 extents in the TRIM dataset. Errant mapping of late-lying snow thus led to 52% more surface-area loss reported in the Bolch et al. (2010) inventory than I find for the 28 glaciers of this comparison.

Based on extrapolation using a gradient of average surface elevation change, I estimate a net change of -53 km² (-6.8%) for the period 1985-2005 compared to -114 km² (-13.5%) of Bolch et al. (2010). I estimate Cariboo Mountains glacier extent to have been about 785 ± 37 km² in 1985, versus 845 km² reported by Bolch et al., (2010), a difference of 7.7%. The Cariboo Mountains glacierized area for 1952 was 824 ± 39 km², still 2.5% less than the 845 km² of the 1985 TRIM extents used in previous work (Schiefer et al., 2007; Bolch et al., 2010). My results thus demonstrate that glacier loss during the period 1985-2005 was higher than during previous epochs, but that the absolute surface area lost in this period may have been significantly overestimated in previous studies.

4.5.3. Thickness change

Observed thinning rates during the period 1952-2005 for seven glaciers averaged -0.23 ± 0.12 m w.e. a⁻¹, resulting in a net loss of -0.48 ± 0.05 km³ w.e. (Table 4.7). This rate of thinning is considerably less then long-term thinning rates for glaciers of the Columbia Icefield (-0.6 ± 0.3 m w.e. a⁻¹; Tennant and Menounos, 2013) and Alaskan glaciers (-0.48 ± 0.10 m w.e. a⁻¹; Berthier et al., 2010). However, my average thinning
rate is similar to those reported for nine North America glaciers for the period from the mid-1950s to the mid-1990s, which range from +0.01 to -0.61 and average -0.21 m w.e. a\(^{-1}\) (Sapiano et al., 1998).

From 1952-1985 for seven glaciers I find an average thinning rate of -0.14 ± 0.16 m w.e. a\(^{-1}\) with a total volume change of -0.20 ± 0.06 km\(^3\) w.e. The two smallest glaciers (glaciers 1 and 11) thickened and advanced over the same period, however. Brewis (2013) also found that four of six glaciers in the Canoe Basin of the Premier Range, which is about 10 km to the south of my study area (Fig. 4.2C), thickened and advanced during the period 1955-1970.

The observed acceleration of post-1985 thinning accords with other studies of glacier change in northwest North America (e.g., Rasmussen and Conway, 2004; Brewis, 2012; Tennant and Menounos, 2013). Schiefer et al. (2007) found that over the period 1985-1999 glaciers of the Cariboo Mountains thinned by a rate of -0.58 m w.e. a\(^{-1}\) and lost -7.06 km\(^3\) w.e. Using my two different methods of extrapolation described in section 4.3.4, my thinning rates of all glaciers in the Cariboo Mountains yield comparable estimates of ice loss (-6.24 ± 1.32 km\(^3\) and -7.58 ± 1.17 km\(^3\)).

Unlike the period 1985-2005, my methods of extrapolation to Cariboo Mountains glaciers yield significantly different volume change estimates for the period 1952-1985 (-4.03 ± 2.04 km\(^3\) vs. -2.31 ± 2.17 km\(^3\)). I hypothesize that during an epoch of strongly negative mass balance - when ice dynamics is less important for redistribution of mass and thinning is a response to negative surface mass balance - extrapolation from a small subset will perform better than during an epoch where some glaciers advanced. I recommend further study of extrapolation of regional volume change from a subset of
glaciers, but with a more robust sample size to better constrain regional variability of surface-elevation change.

4.5.4. Relations of extent change to glacier morphometry

Glacier surface area is the single morphometric parameter that commonly correlates with extent change (Serandbrei-Barbero et al., 1999; Hoelzle et al., 2003; Paul et al., 2004; Andreassen et al., 2008; Bolch et al., 2010; Paul and Andreassen, 2009). From 1985-2005, a period with increased rates of recession, I find average relative area change of -11.2, -9.3, -7.5, -5.4, and -5.4% for my five progressively larger size classes (Table 4.4). I detect a correlation ($p = 0.10$) between surface area and relative area change from 1985-2005, (Table 4.6) indicating generally more relative area change for smaller glaciers. However, from 1952-1985, a period with reduced rates of recession or advance, this does not apply; average relative area change was -2.5, +3.4, -4.6, -9.0, and -4.1% for the same five size classes, and sign of the r value (not significant) is negative. It appears that the magnitude of glacier change is not simply related to dimensional attributes of a given glacier; they apparently vary with differing climatic regimes.

Studies rarely find a significant correlation between most other morphometric parameters and glacier extent change (Paul, 2002; Granshaw and Fountain, 2006; Andreassen et al., 2008; Bolch et al., 2010; DeBeer and Sharp, 2009; and Paul and Andreassen, 2009). These parameters include slope, aspect, and median, minimum, and maximum elevation. Slope and absolute area change are correlated in all periods. Slope correlates with relative area change from 1952-1985, and on a net basis from 1952-2005, but not for the latter period 1985-2005 (Table 4.6). This relation, indicating that the
steeper glaciers lost less absolute and relative surface area, may be spurious, however, as slope is highly correlated with glacier surface area and length (e.g., Hoelzle et al., 2003).

I also find that median, minimum, and maximum glacier elevations (along with elevation range) and area change are related, but the strength of these associations is not constant through time (Table 4.6). Glaciers that advanced from 1952-1985 were generally at higher elevations than those that receded (Table 4.5). For this epoch, I find a correlation between median elevation and absolute area change indicating that glaciers with higher median elevations receded less or advanced. This relation, however, does not hold for the period from 1985-2005. I also find correlation between minimum elevation and elevation range for both absolute and relative area change, but only in the period 1985-2005. A more robust study with a larger sample size is required to assess whether the temporal changes I observed in this study are present for other mountain ranges.

4.5.5. Relations to climate

Mean annual and seasonal temperatures increased from 1952-2005, particularly after 1985, whereas precipitation leading to snow accumulation (accumulation season primarily) generally decreased (Figs. 4.9 and 4.10). It is likely that observed glacier loss resulted from increased (decreased) temperatures (precipitation), a finding in agreement with previous studies for glaciers in the Cariboo Mountains (Luckman et al., 1987 and Brewis, 2012). Other studies have concluded that increasing temperatures are largely the cause of recent increased glacier recession of North America glaciers (Rasmussen and Conway, 2004; Arendt et al., 2009), however these studies focused largely on maritime glaciers. The role of decreased precipitation in the Cariboo Mountains may be a local
phenomenon, or perhaps reflect a difference between conditions in maritime and more continental climates.

There was little temperature change, or decreased temperatures from 1952-1970 to 1971-1986, but following 1985, temperatures increased in the Cariboo Mountains (Fig. 4.9). After 1985, increased annual temperature in the Cariboo Mountains occurred in both the accumulation and ablation seasons, but change during the accumulation season is double that of the ablation season.

Precipitation decreased consistently over the three periods of analysis for the accumulation season and annually (Fig. 4.10). Decreased annual precipitation is primarily an accumulation season phenomenon, whereas ablation season precipitation changed little or even increased. Lower accumulation season temperatures from 1971-1985 may have counteracted decreased precipitation, leading to snow accumulation comparable to that of the period 1952-1970. Some, but not all, glaciers thickened and advanced between 1952-1985, a result of both individual glacier hypsometry leading to more retained accumulation, and length and slope angle that lead to a more rapid terminus response.

ClimateWNA records of temperature and precipitation indicate alignment of wetter and cooler periods with anomalously low-pressure (1952-1970 and 1971-1985) and drier and warmer periods with anomalously high-pressure (1985-2005) (Fig. 4.11). This shift to drier and warmer conditions, to atmospheric circulation favoring high-pressure over the northwest North America, and with coincident glacier recession and thinning after the late-1970s or 1980s has been noted previously (e.g., Bitz and Battisti, 1999; Hodge et al., 1998; Rasmussen and Conway, 2004). My findings for glaciers of
the Cariboo Mountains accord with this earlier work and further illustrate the extent of this climate change of the late-20th century and resultant glacier recession and thinning.

Glaciers of the Cariboo Mountains should strongly recede for the remainder of the 21st century. Recent work projects warming of 1.8°C to 2.7°C above the 1971-2000 mean by the 2050s for the Columbia Basin of BC, an area that includes a portion of the Premier Range (Murdock et al., 2013). Assuming a lapse rate of 0.006 °C m⁻¹, such warming would result in a rise in the ELA of 300 to 450 m. Given current glacier hypsometry (Fig. 4.5), an assumed steady-state accumulation-area ratio (AAR) of 60% indicates a present steady-state ELA for Cariboo Mountains glaciers of approximately 2,375 m. A rise in the ELA of from 300 to 450 m would reduce the AAR from 60% to between 12 and 5%, leaving glaciers of the Cariboo Mountains grossly out of balance and greatly diminished by the mid-21st century. Such a decline is concordant with projections that glaciers of the eastern slopes of the Rocky Mountains will lose 80-90% of their volume by 2100 (Marshall et al., 2011).

4.6. Conclusions

Glaciers of the Cariboo Mountains shrank during the period 1952-2005, losing an average of -0.19 ± 0.05% a⁻¹ of their original surface area. This shrinkage is separated into two distinct periods: 1) from 1952-1985 when many glaciers advanced and net recession averaged only -0.05 ± 0.10% a⁻¹, and 2) from 1985-2005 when all glaciers receded and rates of recession accelerated (by a factor of eight) to an average of -0.41± 0.12% a⁻¹. Thinning rates also increased from -0.143 ± 0.043 m w.e. a⁻¹ (1952-1985) to -0.500 ± 0.070 m w.e. a⁻¹ (1985-2005). Temperatures increased from 1952-2005,
primarily after 1985, whereas accumulation season precipitation decreased over the period of study, indicating that increases (decreases) in temperature (precipitation) explain the observed pattern of glacier change.

The inventory of Bolch et al. (2010) likely overestimated the recession of Cariboo Mountains glaciers from 1985-2005. For 28 common glaciers my extents average 2% smaller than the 2005 extents generated with the semi-automated method of Bolch et al. (2010) and 5% smaller than the 1985 extents of the TRIM dataset. This overestimation in 1985 – likely due to mapping of late-lying seasonal snow as glacierized area in the TRIM dataset – results in recession in the Bolch et al. (2010) inventory that is 52% greater than what I report here.

My results indicate that relations between glacier change and dimensional attributes of the glaciers of this study are not stable through time; this non-stationarity may by related to ice dynamics and the complex way glaciers respond to changes in surface mass balance.
5. **Thesis conclusions**

My thesis addressed two primary objectives. First, I applied new and contemporary technologies and refined existing methods to measure seasonal-to-annual glacier change. Second, I documented twentieth century glacier fluctuations in the Cariboo Mountains and the climatic and morphometric factors which were responsible for these fluctuations. Within this final thesis chapter, I discuss the progress gained, the study limitations, and the knowledge gaps that remain. I conclude with recommendations for future work that will help the glaciological community make steps towards filling these knowledge gaps.

5.1. **Progress gained**

In chapter two I evaluated different methods of measuring annual mass balance as they apply to temperate glaciers. Photogrammetric and GPS methods can estimate annual glacier mass balance with errors that are comparable to the glaciological method, and with reductions in time, effort and expense. These two geodetic methods can also be used to resume research at glaciers with interrupted or discontinued series of dimensional change, expand mass change studies to new regions, or to better quantify errors in mass balance records.

Photogrammetric mass balance measurements are advantageous in that they provide a method to readily calculate mass change for multiple glaciers in a region, especially for large, inaccessible glaciers. The spatial resolution of aerial photography (< 1 m) is generally superior to most satellite imagery (~ 5-30 m), and results in vertical accuracy that makes possible annual measurements of glacier elevation change. The
2011 imagery of Castle Creek Glacier revealed the additional facility of multispectral aerial photography, and in particular the potential to largely resolve the long-standing problem of poor contrast in glacier accumulation zones.

I show that the GPS methodology, previously thought to be unusable without a detailed understanding of vertical surface velocity, can be used to generate measurements of annual glacier mass balance. High precision GPS can be used in tandem with glaciological methods to improve our understanding of specific mass balance, particularly in regard to constraining the roles of densification, internal and subglacial mass balance, advection of topography, and spatial variability. I demonstrated how synchronous at-a-point measurements using GPS and glaciological mass-balance methods could be used to estimate vertical surface velocity. This method can be used to measure submergence in the accumulation zone, which is difficult to measure with the majority of remote-sensing methods.

The photogrammetric and GPS methods do not suffer from the primary biases within the glaciological method, namely those of internal and basal mass balance, and hold promise in improving our understanding of biases in the glaciological method. As geodetic methods effectively include internal and basal mass balance, and as biases within DEMs are relatively simple to constrain, decadal glacier volume change from geodetic methods is often used to verify and correct glaciological measurements (e.g., Thibert et al., 2008; Zemp et al., 2010). I contend that geodetic methods might also be used to constrain error in glaciological measurements of glacier wide mass balance on a sub-decadal, and perhaps even annual basis. Furthermore, I show that geodetic methods – particularly the high-precision GPS method studied here – might also be used to
constrain biases within glaciological measurements at-a-point. This may help constrain measurements of internal and basal mass balance, of which current understanding is limited, and estimated to vary from millimeters to decimeters per annum (e.g., Cogley and Adams, 1998; Thibert et al., 2008) or, in locations with high precipitation, up to meters per annum (e.g., Franz Josef Glacier, New Zealand; Alexander et al., 2011). The GPS method does not suffer from biases inherent in either the glaciological or photogrammetric methods, although it may suffer from a bias associated with vertical surface velocity in some applications. Vertical surface velocity is a confounding factor in assessing at-a-point geodetic mass balance. However, future testing at sites where flow dynamics are minimized - such as for small downwasting glacier remnants (e.g., Burroughs Glacier remnant, Glacier Bay, Alaska) or for polar glaciers with little flow (e.g., glaciers of the Dry Valleys of Antarctica).

In chapter three, I used high-resolution aerial photography of a series of push moraines to reconstruct the longest annual record of length change for a North American glacier. Historical aerial photography enabled me to reliably date each moraine and the intervening periods of recession.

Statistical relations between climatic forcing and nearly 50 years of annual extent change of Castle Creek Glacier allowed the distinction between a lagged glacier ‘response’ and both immediate and delayed glacier ‘reaction’, adding nuance to our knowledge of how glaciers respond to changes in climate, a discussion that has largely been framed by the idea of a theoretical response time (e.g., Jóhannesson et al., 1989) or, in other words, a time period during which glaciers have ‘memory’ of a prior climatic regime or perturbation (Greene, 2005). I find annual length change of Castle Creek
Glacier to strongly correlate with summer temperature at zero lag. This result is important as it suggests that glacier length directly responds to changes in surface air temperature. The results of this study also indicate that, as expected, that the glacier has a delayed reaction to changes in precipitation. I find a lagged peak in correlation between winter precipitation and terminus behavior. However, this timing is when terminus response to precipitation is maximized, and does not coincide with theoretical response time, which would occur later.

Anomalously high rates of Castle Creek Glacier recession in the early-1990s are synchronous with rapid retreat elsewhere such as the South Cascade and Blue glaciers of Washington, suggesting a common driver. Similar maxima of retreat are further evidence of the importance of an immediate terminus response to temperature and ablation, as it would be exceedingly unlikely to find coincidence of termini behavior in response to spatially variable precipitation that is further modulated by glacier dynamics.

Recent work has demonstrated that kilometer-scale glacier extent changes can be driven by the 'noise' of natural interannual variability and the glacier-specific 'memory' of prior perturbations (Huybers and Roe, 2009; Roe and O'Neal, 2009). A primary assumption of the model used in these studies, based on the work on theoretical response time by Jóhannesson et al. (1989), is that all accumulation and ablation anomalies act immediately to drive glacier length change. Other work has shown that this model, and slight derivations of it, gives realistic glacier variations on timescales of decades to centuries (e.g., Harrison et al., 2001; Oerlemans, 2005). However, my results indicate more complexity in glacier response to climate forcing with immediate and lagged reactions to temperature and precipitation respectively, both occurring before theoretical
response time, or memory time. An immediate response of the terminus to changes in surface air temperature indicates that geologic records of glacier fluctuations within a given region would be expected to preserve some degree of past temperature variability.

My analysis of annual Castle Creek Glacier recession reveals that response of a glacier to climate is complex and related to terminus change that occurs both instantaneously and delayed. To summarize, I find: 1) the Castle Creek Glacier terminus ‘reacts’ immediately to temperature and associated ablation; 2) the terminus reaches a maximum ‘reaction’ to retained accumulation-season precipitation at a lag that is glacier specific and in relation to flow dynamics; 3) both of these terminus ‘reactions’ occur prior to theoretical glacier ‘response’ or the culmination of a glacier’s ‘memory’ of a prior climate; and 4) correlation between records of annual glacier length change is unlikely in the absence of an immediate terminus reaction, given the modulating influence of glacier-specific flow dynamics.

In chapter four, I used photogrammetry to determine extent and volume change of a subset of glaciers in the Cariboo Mountains. This analysis showed no detectable change in average glacier area during the period 1952-1985, but nine glaciers advanced during this period. In contrast, all glaciers receded during a period (1985-2005) characterized by warm, dry conditions.

Glaciers that advanced during the period 1952-1985 were small, short and steep ice masses that extended over a small elevation range. The median elevation of these glaciers and the elevation of their toes were higher than glaciers that receded. It appears that these small, steep glaciers responded more quickly to cool, wet conditions which
characterized the period 1952-1985. They thus appear to be more sensitive indicators of climate change than their larger, less steep counterparts.

In the Monashee Mountains, some 200 km to the south, DeBeer and Sharp (2009) found that glaciers overall did not retreat during the period 1951-2001. That study concluded that these small ice bodies might not be especially sensitive to changes in climate. In contrast, I found that the smallest glaciers were more responsive than the larger ice bodies in the Cariboo Mountains – small, steep glaciers underwent an advance and later retreat while large glaciers experienced overall retreat. This underscores the importance of remote-sensing studies on a decadal or even sub-decadal basis, and stratification by size class, in order to use glaciers as an indicator of climate change.

I also found that the relation between glacier morphometry and glacier change varies with time and climate. From the cool, wet period 1952-1985 - with no discernable net glacier change, to the warmer, drier period 1985-2005 - when all glaciers receded, relations between relative area change and length, area, slope, min Z, max Z, and Z range lost or gained statistical significance, or changed sign altogether. Characterization of this variability is important as it directly affects the sensitivity of a glacier to climate change and the selection of a representative subset of glaciers for long-term study.

5.2. Study limitations

I detail the primary study limitations of my thesis here with the intent of providing guidance for future efforts.

The results of chapter two depend heavily on the implementation of the three mass-balance methods I compared. In retrospect, Castle Creek Glacier was not an ideal
test site for this study, with its size (9.5 km²) and mid-glacier icefall leading to less than optimal densities of point measurements for both the glaciological and GPS methods.

Unfortunately, planned 2010 aerial photography did not occur due to inclement weather. A lack of 2010 photography limited my analysis of photogrammetric annual mass balance to only one year instead of three. Timing of the aerial photography also necessitated correction using a temperature-index model in order to make comparisons with the glaciological and GPS methods. Better temporal alignment of image acquisition and field work would have improved the analysis.

Limited measurements hindered a rigorous analysis of the GPS method for glacier mass balance. Terrain complexity and base station power failures limited radio communication between the base station and the rover, preventing adequate measurements of elevation change in the accumulation zone. The relatively remote nature of the site, and lack of a field station with means to recharge the automotive batteries used to power the base radio limited my ability to thoroughly assess the capabilities of the GPS methodology.

In chapter three, I derived an annual record of length change from a preserved series of annual push moraines. The extent of the preserved push moraines varies from 4 to 54% of the total terminus width, and averages 26%. Glacier length change is not constant across the terminus, and this variability in push moraine length likely imparts error in obtaining a representative length change for the glacier. Averaging glacier length change across 4% of a glacier terminus, however, will likely be more representative than measuring the maximum length of the glacier, a common methodology, but one which yields a point measurement for the glacier.
My analysis of the climatic drivers of Castle Creek Glacier annual recession relies heavily on the Prince George meteorological station, which lies ~180 km to the northwest. This reliance on data from a distant station is not optimal, and data from a closer location would likely give improved results, particularly in regards to analysis of precipitation, which is highly spatially variable.

My analysis of Cariboo Mountains glacier change in chapter four is limited by the glacier subsets used to assess extent and volume change. The quantity of glaciers in the subsets is largely a product of the availability and quality of historical aerial photographs. Aerial photograph quality is less important for measurement of glacier extent than for glacier surface elevation, but photo availability (spatial and temporal) limits the glaciers that can be measured. While I measured extent for 33 glaciers I could only measure volume change for seven of these glaciers. This reduced subset is due to poor contrast in the accumulation zone of glaciers in the earliest photography.

Results of Chapter 4 are also limited by the dates of aerial photograph acquisition. To measure the glacier subsets studied here necessitated the use of variable acquisition years to define a common epoch for comparisons from glacier to glacier and within Cariboo Mountains sub regions. Average annual rates of change within any given epoch in this study are unlikely to markedly change if image acquisition dates differ by up to a few years, however significant error might result with shorter epochs and/or a greater difference in acquisition year.
5.3. Knowledge gaps

The magnitude of both error and bias remain primary knowledge gaps in measurements of glacier mass balance. Measurement error remains an issue in the glaciological and photogrammetric methods, whereas it is minimized in the GPS method. Spatial variability of ablation, accumulation, and vertical surface velocity are key uncertainties whenever limited point measurements are used to represent large spatial areas, such as with the glaciological and GPS methods. Density conversion imparts an error in the geodetic methods and remains an important error term.

The magnitudes of internal and basal mass balance remain largely unknown; these processes significantly vary from glacier to glacier. These biases may render records of glaciological mass balance significantly in error and may do so with magnitudes that are highly site specific. Appropriate ground control remains important in constraining bias in photogrammetric methods, and the role of densification in geodetic methods is a problem that needs further investigation. Photogrammetry also suffers from biases associated with manual operator error and with modeling of melt between the date of imaging and the end of the balance year. While I hypothesize that the GPS method largely avoided the biases of the other two methods, it will be important to further study the role that vertical surface velocity may play in imparting bias in a study that relies on a limited subset of point measurements.

There is a need for additional records of annual glacier length change. Records of glacier extent and length change serve as proxy records of climate (e.g., Oerlemans, 2005), and provide the ability to determine glacier sensitivity to climate forcing, both for reconstruction of paleoclimate (e.g., Menounos et al., 2009), and prediction of future
glacier change, particularly with regard to feedback mechanisms related to changing glacier morphometry (e.g., Haeberli et al., 2007). Records of annual glacier length change, particularly for glaciers with measurements of annual or seasonal mass balance, enable detailed analyses of dimensional glacier response to mass-balance forcing. It would be informative to have the capability to compare empirical records of glacier reaction with models of theoretical glacier response. Unfortunately there are few glaciers for which records of both mass balance and length change exist on an annual basis. It will be important to further investigate the roles of immediate and lagged reactions to mass-balance forcing, and to do so for glaciers that are more representative of the range of morphometries within a region. Measurements of annual glacier length change are relatively simple and can be readily made where safety concerns make glaciological mass-balance measurements impractical. One means of filling this data gap is by examination of annual push moraines. However, if such geomorphic features are found, accurate dating of moraine formation and annual length change also require adequate available historical imagery. A concerted effort should be made to continue existing length-change records, reestablish interrupted series, and to begin new efforts to monitor annual glacier length change, particularly where field studies of glacier mass balance are currently undertaken.

Shortcomings remain in the use of aerial photogrammetry to measure glacier change. Errors in photogrammetric mapping of glacier extents and surface elevations remain a challenge, particularly with regard to the use of lower resolution imagery, and with poor contrast due to seasonal snow cover and sun angle. While high-resolution multispectral imagery may largely solve the problem of contrast in glacier accumulation
zones, sun angle with respect to fresh snow covered surfaces will remain a concern. The performance of multispectral imagery with automated photogrammetric methods has yet to be fully assessed. Operator bias in manual aerial photogrammetry remains a concern, and further testing should be done with respect to glaciological applications, and with relation to the performance of automated methods.

Significant advances have been made in recent years, but for many regions there are limited opportunities to study decadal glacier change in the pre-satellite era. Furthermore, the temporal resolution of some multi-decadal studies has resulted in net change across many decades, where the decadal variability was likely more complex than the multi-decadal mean. Such studies provide an understanding of general net change, but do little to inform our understanding of glacier response to climatic forcing and relations to glacier morphometry.

The demonstration in my thesis that relations between glacier morphometry and change may vary temporally with respect to climate identifies a new gap in understanding. This variability, when present, will impart challenges in the selection of representative glacier subsets – with the ideal subset likely changing temporally, and with the forecasting of glacier response with future climate change.

5.4. Recommendations

I conclude with ten recommendations for future work that will help address the knowledge gaps mentioned above:

1. Pair geodetic and glaciological measurements of annual and seasonal mass balance to further constrain error and bias within each methodology. One should do so at scales from glacier-wide to specific measurements. Studies are needed of the potential error imparted by
advection of topography in both the GPS and photogrammetric methods. Use of this method to study a stagnant, down-wasting glacier (with limited ice flux and accumulation zone) might yield constraints on the magnitude of internal and basal mass balance.

2. **Conduct further tests of the GPS methodology at a site with an established field research station.** Testing on larger glaciers will necessitate assessment of appropriate thresholds for base-line distances. Use a stadia rod to maximize measurement precision. Test sampling along longitudinal flow-lines, on a grid, and with respect to measurement density.

3. **Pursue the photogrammetric method of measuring mass balance with new platforms (e.g., unmanned aerial vehicles) and technologies (e.g., multi-spectral sensors, structure from motion photogrammetry, and LiDAR).** Compare manual and automated photogrammetric methods for glaciological applications.

4. **Combine at-a-point GPS and glaciological measurements to investigate temporal and spatial variability in vertical surface velocity.** Further testing should be done with regard to measurements of submergence. Use this method to help define the role of ice flux in glacier volume change.

5. **Pursue annual monitoring of glacier length change, particularly for those glaciers where annual or seasonal mass balance is measured.** Such monitoring is relatively simple, particularly when the glacier is already monitored for mass balance. Partnerships with groups and individuals within close proximity to glaciers (e.g., park rangers, mountaineers, local clubs) might provide both valuable data and also important outreach and engagement opportunities.

6. **Use high-resolution aerial photography to search for series of annual push moraines, and, where possible, extract from them records of annual glacier length change.** The growing online availability of freely accessible high-resolution imagery (e.g., via GoogleEarth™) affords an expedient means of searching glacier forefields for sites with annual push moraines.

7. **Analyze dimensional glacier response to mass-balance forcing for glaciers where records exist of annual mass balance and extent change.** Compare such empirical records and analysis of glacier reaction with models of theoretical glacier response.

8. **Continue to pursue decadal pre-satellite era remote sensing of glacier change in unstudied regions to investigate temporal and spatial variability of glacier response to climate change.** Such studies should be
done on a decadal basis where possible to capture short-term glacier response.

9. **Use a robust sample size to investigate the temporal variability of relations between glacier morphometry and change.** Such a study should be done on a multi-temporal basis across periods with variable climatic forcing, and for regions with different climatic regimes (e.g., maritime vs. continental).

10. **Use comprehensive inventories to determine the regional representativeness of existing benchmark glaciers and to identify ideal glaciers or glacier subsets for future study.** Assess regional representativeness for regions with historical photographic coverage that covers a bulk of the glacierized area – allowing for large sample sizes and flexibility in determination of representivity – and with a temporal resolution that allows assessment of decadal change.
References


