The Relative Importance of Calving and Surface Ablation at a Lacustrine Terminating Glacier

A Detailed Assessment of Ice Loss at Bridge Glacier, British Columbia

by

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B.Sc., The University of British Columbia, 2010

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF

MASTER OF SCIENCE

in

The Faculty of Graduate and Postdoctoral Studies

(Geography)

THE UNIVERSITY OF BRITISH COLUMBIA

(Vancouver)

August 2014

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Abstract

Bridge Glacier is a lacustrine calving glacier located in the southern Coast Mountains and terminates in a 6.2 km² proglacial lake. The glacier has retreated more than 3.55 km up-valley since 1984, the majority of the retreat having occurred since 2003. While surface melt may have contributed to the retreat, calving allowed for an additional annual volume of ice loss. The relative contributions from surface melt and calving to the total volume of ice loss is examined for the 2013 melt season. Surface melt is quantified using on-glacier meteorological data to drive a distributed energy balance model. The calving flux is quantified using field measurements of lake bathymetry, terminus area change, and ice thickness. Calving flux estimates are completed by daily measurements of terminus surface velocity derived from manual feature tracking using oblique time lapse camera imagery. Calving accounts for 23% of the total ice loss in the 2013 melt season, suggesting that surface melt is the main driver of mass loss at Bridge Glacier.

Data from the 2013 field season is used to inform historical calving flux and surface melt estimates from 1984 to 2013. The calving flux is minor until 1991, at which point the glacier terminus achieves flotation, and begins to discharge large tabular icebergs. Calving was characterized by large, multi-annual retreats, alternating with periods of relative stability. The calving flux peaked from 2005 to 2010, when it was roughly equal to the mass loss due to surface melt. Calving was a much smaller contributor of mass loss from Bridge Glacier, except for a transient high-calving period in the late 2000s. Looking forward, Bridge Glacier will retreat into shallower water where the terminus will no longer float, and calving losses should decrease substantially. Although calving losses will become an increasingly minor portion of the mass balance, future retreat is expected at Bridge Glacier due to a legacy of dynamic thinning brought about by its transient calving phase.
Preface

This research was made possible with the help of many people. Technical and academic advice from Dr. Michele Koppes and Dr. R. Dan Moore significantly improved fieldwork planning, focus, and the writing and content of this thesis. Alistair Davis and Melanie Ebsworth assisted with field work. Data collection, field work, and data analyses were improved by advice and support from Lawrence Bird.

A paper is currently being written that encompasses Chapters 3, 5, and 6. It is currently titled “The Relative Contributions of Calving and Surface Ablation to Ice Loss at a Lacustrine Terminating Glacier” and is co-authored by M. Koppes and R.D. Moore.
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Acknowledgements

Although only my name appears on this thesis, what is written would not have been possible without the help of many people. I would like to thank Michele Koppes for taking me on as a student, and for guiding me through two exciting years of science and discovery, from field work, to computer work, to actually trying to make sense of, and communicate, the findings. I would like to also thank Dan Moore for providing another guiding voice in the process, giving a broader context to the work, new ideas, and many hours of guitar playing. This work was supported by a Discovery Grant from Natural Sciences and Engineering Research Council (NSERC) to Dr. Koppes and Dr. Moore as well as a Graduate Support Initiative scholarship to M. Chernos.

I would like to thank the Geography Department and its faculty for not only two great years as a master student, but four more as an undergraduate. The courses and people here have taught me to think critically, and have given me a new appreciation and passion for the science that governs our natural world. I would like to thank Dr. Brett Eaton, who initially convinced me after a Geography field course, that I should be pursuing a BSc. (not a BA. for which I was originally registered), and helped navigate the inner workings of the university to make it happen. I would also like to thank the professors for whom I was a teaching assistant (Drs. Khaled Hamdan, Simon Donner, Greg Henry, Mike Bovis, Andreas Christen, and Dan Moore) who shared with me their passion for teaching, and their field. Finally, I would like to thank Mike Church his guidance, passion, and perspective on science and research through work on the Fraser River and beyond.

I would also like to acknowledge Lawrence Bird, for his support navigating Bridge Glacier field work, Geography Building construction, R tutorials, field gear construction, and the spatial distribution of UBC pubs. I would also like to thank Alistair Davis for his field work acumen, great field stories, and pretty clever Glacier AWS tripod design.

I would like to thank my family for their support and encouragement, and for giving me perspective on my journey to become the Master of Glaciers (or something like that). Finally, I would like to thank Melanie Ebsworth, who has graciously put up with many hours of typing and incessant conversations about glaciers. Her support and patience has never been more evident than when our hikes get side-tracked to get “just another view” of a glacier.

To my family and friends, colleagues, and supervisory committee:

Thank you, and happy trails!
To Bridge Glacier,
and everyone else’s Bridge Glacier,
wherever that may be.
Chapter 1

Introduction: Calving Glaciers and Their Relationship to Climate

1.1 Motivation

Since the 19th century Little Ice Age, glaciers across the globe have been shrinking at an accelerated rate. Although this retreat has been irregular, a general trend of 20th century retreat is pervasive, and well correlated with an increase in global mean temperatures (Oerlemans, 2005). The reduction in glacial area has raised considerable concern about potential changes in the timing, volume, and duration of summer streamflow (Marshall et al., 2011; Stahl et al., 2008). These changes have major implications for hydroelectric projects, agriculture, and water quantity and quality for general consumption. Similarly, glacier retreat, and subsequent hydrological changes, also has significant ramifications for eustatic sea level rise (Dyurgerov et al., 2002; Barry, 2006). While recent glacier retreat is well documented (Kaser et al., 2006), the projection of future glacier retreat is critical to the management of water resources, projection of future channel morphology (Cowie et al., 2014), and the establishment of ecological and aquatic habitats (Milner and Bailey, 1989).

Due to their direct relationship with air temperatures and precipitation, glaciers serve as important high altitude weather stations (Kaser et al., 2006). As barometers of a changing climate, fluctuations in glacier mass balance, area, or length can be related to air temperature (Oerlemans, 2005), supplementing a global network of temperature records, much of which is biased towards lower elevation sites (Christy et al., 2000; Oerlemans, 2005). Finally, glacier fluctuations provide an additional record of a changing global climate, independent of other climatic indicators such as tree rings, sediment cores, and isotope records (Oerlemans, 2005; Dyurgerov et al., 2002).

Glacier fluctuations are predominantly a product of changing climatic conditions, although glaciers that terminate in bodies of water respond at least partially independently
1.2. Glacier Mass Balance

of climate on decadal timescales (Warren and Kirkbride, 2003; Post et al., 2011). This de-
coupling of the climate-glacier signal is due to the fact that calving, the process by which
icebergs are mechanically discharged from the glacier terminus, can be an important addi-
tional source of ice loss (Benn et al., 2007a). While calving glaciers have a more complex
climatic signal than one from glaciers that terminate on land (Van der Veen, 2002; Motyka
et al., 2003), their inherent instability suggests that they have the potential to contribute
disproportionately to eustatic sea level rise (Meier and Post, 1987; Dyurgerov and Meier,
2005), making them an important ‘grey area’ in our understanding of how glaciers respond
to climate.

While previous work has focused on the mechanisms and rate of calving in both marine
and lacustrine glacier systems (Benn et al., 2007b; Post et al., 2011), more work is needed
to assess the role and importance of calving on glacier mass balance. The goal of this study
is to quantify the contributions of ice loss from calving and surface melt over a single melt
season. This study then aims to use field data from the melt season to inform estimates
of how the relative importance of calving and surface melt has changed over the transient
calving phase of a lacustrine calving glacier.

1.2 Glacier Mass Balance

Before 1950, there were few in situ observations of glacier mass change, while before 1900,
direct glacier observations were extremely sparse (Rasmussen and Conway, 2004; Oerle-
mans, 1994), limited to anecdotal evidence and artwork (Nussbaum et al., 2007). It was
only with the inception of the International Hydrological Decade in 1965 that a standard
methodology was instituted to quantify glacial change in Canada, and consistent observa-
tions began in the southwestern Canadian Cordillera. Today there are continuous records
from 1965 to present of Peyto Glacier in the Rocky Mountains, as well as similar length
records from Place Glacier and Helm Glacier in the southern Coast Mountains (Demuth
et al., 2006; Moore and Demuth, 2001). These records complement a long-term dataset
at South Cascade Glacier in Washington state, USA, from 1959 to present (Rasmussen
and Conway, 2004; Beedle et al., 2009), among other North American datasets in Al-
berta, northern British Columbia, and Alaska (Pelto, 1987; Rasmussen and Conway, 2004;
Dyurgerov and Meier, 2005; Braithwaite, 2002).

Glacier change is quantified by measuring the net mass balance, a glacier’s annual
change in volume, as measured in units of water equivalent melt. Mass balance can be
partitioned into seasonal balances, where the net balance for the year ($b_n$) is equal to the sum of the winter ($b_w$) and summer ($b_s$) balances:

$$b_n = b_w + b_s.$$  
(1.1)

In many regions, including North America, the winter balance is a period of net accumulation, while the summer is a period of net loss (ablation).

Land-terminating glaciers provide an relatively unambiguous measure of climatic variability due to their strong response to atmospheric indicators. Glacier mass balance is measured using ablation stakes (Ostrem and Brugman, 1991; Braithwaite, 2002), in conjunction with snow depth and density measurements. Although the most direct way to measure glacier mass balance involves manual measurements, simple predictive models have shown that changes can be well predicted by winter precipitation, summer temperatures (see Table 1.1), and glacier geometry (Moore and Demuth, 2001; Leteguilly, 1988). Air temperatures have consistently been shown to produce accurate estimates of summer ablation; however, they lack the explanatory power to describe the main drivers of melt. While the energy pathways to generate melt are more complex (termed the energy balance), air temperatures are still an effective proxy for melt primarily because they are significantly correlated with many of the meteorological variables used in energy balance modelling.


<table>
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<th>$R^2$</th>
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<td>$b_n$</td>
<td>$P_w, T_s$ (Vancouver, BC)</td>
<td>Place Glacier, BC</td>
<td>0.82</td>
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<td>$b_n$</td>
<td>$\min T_s$ (Jasper, AB)</td>
<td>Peyto Glacier, AB</td>
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<td>Place Glacier, BC</td>
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<td>$b_n$</td>
<td>$P_{march}$ (Prince George, BC)</td>
<td>Castle Creek Glacier, BC</td>
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<td>$b_w$</td>
<td>NDJF 500 mb heights</td>
<td>Peyto Glacier, AB</td>
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<td>Gulkana, Wolverine, AK</td>
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<tr>
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1.3 Glacier Flow

In glacial environments, flow is regulated by the driving force of gravity, controlled by the up-glacier mass, and resisted by friction due to drag at the base and sidewalls of the glacier (Nye, 1952). The driving force responsible for down-valley flow is made possible by glaciers gaining mass in the upper reaches of the glacier (accumulation area), while losing mass in the lowest reaches (ablation area). This allows for a mass imbalance causing the glacier to re-adjust its geometry and flow downslope.

For land-terminating glaciers, flow is achieved through a combination of deformation and sliding, resisted partially by lateral and basal drag. The driving and resisting forces are controlled by topography, subglacial hydrology and mass balance (Benn and Evans, 2010). Seasonal variations in glacier velocity are related to melt rates, where higher melt rates increase the amount of water at the glacier bed, in turn reducing basal drag (Bartholomew et al., 2010; Zwally et al., 2002). Sub-annual variations in glacier flow speed are commonly observed for land-terminating glaciers; however, over annual timescales, flow via deformation and sliding is roughly equal (Nye, 1952; Benn and Evans, 2010).

In calving glacier systems flow is complicated by regions where the glacier is in contact with a water body. Water at the bed of the glacier leads to a significant reduction in basal drag, or the complete removal, if the terminus is floating. This reduction in friction results in a substantial increase in velocity in the floating terminus (Van der Veen, 2002; Rignot and Kanagaratnam, 2006; Meier and Post, 1987).

Calving itself has been shown to increase velocity as well, due to the excess ice loss from the terminus enhancing the longitudinal mass imbalance. Similar to a large ablation year, significant ice loss from the terminus increases the average glacier slope, in turn enhancing the driving force and accelerating glacier flow (Van der Veen, 1996; Kirkbride and Warren, 1999; Benn et al., 2007a). Variations in velocity following calving events have been documented in Greenland (Joughin et al., 2004), as well as Alaska, South America and New Zealand (Pelto and Warren, 1991; Brown et al., 1982). The connection between calving and velocity suggests that calving has the potential to enhance velocity, in turn enhancing calving, leading to a runaway effect.
1.4 Glacier Calving

For glaciers that terminate in lacustrine or marine waters, calving can be an important additional component of the mass balance ($b_n$, Equation 1.1), as it leads to greater ice loss than possible through surface ablation alone (Van der Veen, 2002). The calving rate, $U_c$, is calculated as

$$U_c = U_T - \frac{dL}{dt}$$

(1.2)

where $U_T$ is the terminus-averaged surface velocity, and $\frac{dL}{dt}$ is the change in glacier length with time.

The calving rate can be calculated from field measurements (or estimates) of the surface velocity near the terminus, and the change in terminus position (Benn et al., 2007a).

Calving glaciers are observed in both marine and fresh water settings. Calving rates are roughly an order of magnitude greater in marine (tidewater) settings (Venteris, 1999; Haresign, 2004), suggesting that while land-terminating glaciers respond most readily to climatic changes, and marine systems are relatively insensitive, lacustrine glaciers occupy a ‘middle ground’. Differences between marine and lacustrine glaciers can be attributed to some processes only observed in marine environments, such as tides, greater water circulation, and water density differences (Rignot et al., 2010; Benn et al., 2007b), while other factors, such as flow speed and longitudinal strain rates, vary only in magnitude (Van der Veen, 2002; Warren and Kirkbride, 2003).

Multiple studies have found significant empirical correlations between terminus water depth and calving rates (Skvarca et al., 2002; Warren and Kirkbride, 2003; Haresign, 2004). The relationship between calving rate and water depth applies in a variety of settings, from grounded termini to buoyant or near-floating termini across the globe. However, these ‘calving laws’ do not provide a physically based explanation for calving dynamics, nor does it account for observed seasonal fluctuations (Van der Veen, 2002). Furthermore, regional variations in the calving rate (Figure 1.1), as well as greater calving rates in marine environments, indicate that water depth does not provide a full explanation for calving variability.

One of the reasons water depth is important for glacial stability is the potential for deep water to promote terminus buoyancy due to the relative density difference between ice and water. In many cases, flotation of the terminus has coincided with large-scale retreat (Meier and Post, 1987; Warren and Sugden, 1993). Flotation can produce immediate calving due
to torque, while under low velocities a floating terminus may develop. If calving is not immediate, the terminus becomes more sensitive to small perturbations in water level and strain, eventually leading to terminus disintegration and substantial retreat (Boyce et al., 2007).

1.4.1 Calving Mechanisms

Calving events are driven by an interplay of forces acting on the terminus. Longitudinal strain, terminal force imbalance, subaqueous melt, and buoyant torque have been identified as the primary mechanisms responsible for individual calving events (Benn et al., 2007b).

Longitudinal strain near the terminus, the product of large velocity gradients, has been identified as a primary driver of calving events. If velocity gradients are high enough, strain on the glacier leads to crevasse formation. If a crevasse is located relatively close to the terminus, it becomes a ‘preferential line of weakness’. This weakness provides the fracture margin for calving events (see Figure 1.2), producing large tabular icebergs in floating terminus environments.

Calving can also result from an outward force imbalance at the terminus. Above the
1.4. Glacier Calving

Figure 1.2: Calving along a ‘preferential line of weakness’, Bridge Glacier, June 20, 2013.

waterline, outward-acting cryostatic pressure is opposed only nominally by the atmosphere. Below the waterline, cryostatic pressure is greater than the resisting hydrostatic pressure, except at the base of the ice front (Paterson, 2000). This imbalance is greatest at the waterline, where the resisting water pressure is zero (Figure 1.3), leading to the generation of small volume, high frequency calving events, where ice ‘topples over’ into the water body.

In cases where melt of the ice face is greater below the waterline than above, undercutting will occur. This uneven melt progressively increases the force imbalance at the terminus. Conversely, in environments where the water temperature is low or strongly stratified, an ‘ice foot’ can develop (Dykes et al., 2011; Rohl, 2006). The ice foot is less dense than the surrounding water, and buoyant forces cause it to fail, shooting to the surface in dramatic fashion (Skvarca et al., 2002; Motyka et al., 2003).

Although the calving mechanisms have been presented as separate entities, it is rare for a situation to exist without multiple mechanisms acting on the terminus. However, the
dominant mechanism differs depending on the current glacier, water body and climatological conditions. The mechanisms acting on the terminus have significant implications for the magnitude and character of calving events.

In floating terminus environments, a hierarchy in driving calving mechanisms exists. Longitudinal strain is the only mechanism that creates a physical margin along which failure can propagate, making it a first order control on the calving rate. The magnitude of events triggered due to strain is often orders of magnitude greater than other mechanisms. These large, relatively infrequent events supersede the low volume, high frequency events of terminus force imbalances, even often occurring on overlapping areas.

Conversely, for grounded termini, longitudinal strain is less important in determining the calving flux, since glacier flow patterns do not allow crevasses to penetrate the full depth of the terminus (Benn et al., 2007a). Instead, the terminus force imbalance becomes the main driver of individual calving events. This change in driving mechanisms affects the magnitude and character of calving. While floating termini in some environments experience low frequency, large tabular calving events, calving driven by a force imbalance is defined by low magnitude, high frequency events.

Water temperature also plays a significant role in determining the character and magnitude of calving events due to its influence on terminus geometry. High water temperatures promote undercutting, which in turn enhances the terminus force imbalance, promoting
further high frequency, low magnitude events. Alternatively, low water temperatures inhibit subaqueous melt, and if above-water melt is sufficient, can allow for the development of an ice-foot. Ice feet fail due to buoyant torque, and supply low frequency, moderate to large magnitude events in floating terminus environments.

Although water temperature has an effect on terminus geometry, its influence is often dampened by terminus ice velocity. Low velocities allow for longer exposure of the ice to the water body, allowing greater melt rates, and a quicker establishment of instability (Rohl, 2006). High glacier terminus velocities dampen this effect by limiting the amount of time the ice face is exposed to warmer waters before being transported to deeper water, where the glacier calves more readily. The relative importance of individual mechanisms have been related to terminus velocities (Figure 1.4), where fast glaciers respond more strongly to topographic controls, while slower glaciers are more affected by lacustrine processes.

![Figure 1.4: The relative importance of calving controls related to glacier flow speed at the terminus. Adapted from Haresign (2004).](image)

### 1.5 An Intricately Coupled System

Attempts to formulate a portable ‘calving law’ to predict calving rates have proven elusive due to significant variability in calving environments (see Table 1.2). Lacustrine Patagonian
1.5. An Intricately Coupled System

Glaciers Upsala and Leon can be characterized by high flow speeds, high lake temperatures, and deep waters, promoting high calving rates. Conversely, small lacustrine glaciers Maud, Grey, Godley, Hooker, and Ruth in the Southern Alps (New Zealand) have small water depths, low flow speeds, and low lake temperatures, all of which restrict calving losses. While some studies have found topography to be the major control on calving rates and terminus position (Koppes et al., 2011), others have attributed changes to flow gradients (Tsutaki et al., 2011), water temperature (Dykes et al., 2011; Rohl, 2006; Robertson et al., 2012; Motyka et al., 2003), and buoyancy (Boyce et al., 2007).

Table 1.2: Characteristics of selected major lacustrine calving glacier studies. $D_w$ is the mean water depth, $T_w$ is the mean water (depth averaged or range) temperature, $U_c$ is the calving rate, and $U_T$ is the terminus averaged flow speed. Citations: a: Boyce et al. (2007), b: Motyka et al. (2003), c: Warren and Kirkbride (2003), d: Dykes et al. (2011), e: Warren and Aniya (1999), f: Stuefer et al. (2007), g: Haresign (2004)

<table>
<thead>
<tr>
<th>Location</th>
<th>Year</th>
<th>$D_w$ (m)</th>
<th>$T_w$ ($^\circ$C)</th>
<th>$U_c$ (ma$^{-1}$)</th>
<th>$U_T$ (ma$^{-1}$)</th>
<th>Source</th>
</tr>
</thead>
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<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>Mendenhall</td>
<td>1997 - 2004</td>
<td>45 - 52</td>
<td>1 - 3</td>
<td>12 - 431</td>
<td>45 - 55</td>
<td>a, b</td>
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<tr>
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<td></td>
<td></td>
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<td></td>
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</tr>
<tr>
<td>Maud, Grey,</td>
<td>1995</td>
<td>4 - 20</td>
<td>1.7 - 4.3</td>
<td>14 - 88</td>
<td>5 - 151</td>
<td>c</td>
</tr>
<tr>
<td>Godley, Ruth, Hooker</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tasman</td>
<td>1995</td>
<td>10</td>
<td>0.5</td>
<td>28</td>
<td>11</td>
<td>c</td>
</tr>
<tr>
<td></td>
<td>2000 - 2006</td>
<td>50</td>
<td>1 - 10</td>
<td>78</td>
<td>69</td>
<td>d</td>
</tr>
<tr>
<td></td>
<td>2006 - 2008</td>
<td>153</td>
<td>1 - 10</td>
<td>227</td>
<td>218</td>
<td>d</td>
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<tr>
<td><strong>Patagonia</strong></td>
<td></td>
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<td></td>
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<td></td>
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</tr>
<tr>
<td>Upsala West, Ameghino, Grey, Nef</td>
<td>1995</td>
<td>153 - 325</td>
<td>2 - 7</td>
<td>355 - 2020</td>
<td>370 - 1620</td>
<td>e</td>
</tr>
<tr>
<td>Perito Mereno</td>
<td>1995 - 2006</td>
<td>175</td>
<td>5.5 - 7.6</td>
<td>510</td>
<td>535</td>
<td>e, f</td>
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<tr>
<td>Leon</td>
<td>2001</td>
<td>65</td>
<td>4.5 - 7.0</td>
<td>520 - 1770</td>
<td>520 - 1810</td>
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<tr>
<td><strong>Iceland</strong></td>
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<tr>
<td>Fjallsjokull</td>
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<td>75</td>
<td>1.5 - 3.0</td>
<td>582</td>
<td>258</td>
<td>g</td>
</tr>
</tbody>
</table>

Along with regional heterogeneity, there is substantial variability in the temporal controls on calving rates. Over long timescales (decadal or longer), climatic warming has been suggested as the dominant control on glacial mass balance and calving rates (Warren and Kirkbride, 2003; Motyka et al., 2003). Sustained negative mass balance for multiple years results in thinning, most pronounced near the terminus, which can initiate buoyancy-driven retreat from deep to shallower water. Over annual timescales there is a seasonal increase in
the calving rate in the late summer, attributed to surface ablation and thinning (Haresign, 2004; Van der Veen, 2002). Thinning acts in concert with an observed increase in lake level (and most likely subglacial meltwater enhanced velocity), increasing the probability of the terminus achieving flotation.

For lacustrine glaciers, temporal variability is further accentuated by the transient ‘life-cycle’ of calving retreat over an over-deepened lake or freshwater fjord. Controls, mechanisms, and calving rates vary as the glacier retreats from deep waters, where geometry and submarine melt play dominant roles, and calving rates are high (Meier and Post, 1987; Rignot et al., 2010), to shallower waters, where flow gradients and water level play a larger role and calving losses are significantly smaller (Tsutaki et al., 2011; Boyce et al., 2007). Eventually, the terminus retreats to dry land, where calving is no longer a significant source of ice loss, allowing the glacier to thicken and possibly re-advance (Van der Veen, 2002), or continue to retreat due to a legacy of dynamic thinning and continued climatic warming (Oerlemans, 2010).

1.6 Research Gaps and Study Goals

While advances have been made in predicting the magnitude and timing of calving in water-terminating glaciers, there remains many questions about their nature, timing, and importance. There is an insufficient theoretical understanding of the drivers of calving to make reliable predictions and quantify ‘calving laws’. Similarly, a robust empirical framework has so far proven elusive due to a complex set of interconnected mechanisms, diverse range of environments, and relative paucity of study sites worldwide. In order to better understand each unique calving system, a better understanding of the broad commonalities in lake-terminating glacier response is required.

Although understanding lacustrine glaciers is of critical importance for better watershed management and for isolating the climatic signal in calving glaciers, few lacustrine glaciers are studied worldwide. Most notably, work in North America has focused on Mendenhall Glacier in Alaska (Motyka et al., 2003; Boyce et al., 2007), while international efforts have focused on Tasman Glacier in New Zealand (Warren and Kirkbride, 2003; Dykes et al., 2011; Dykes and Brook, 2010) and a few Patagonian glaciers, most notably Perito Mereno Glacier (Warren and Sugden, 1993; Warren and Aniya, 1999; Stuefer et al., 2007). Bridge Glacier presents another valuable study site to supplement the worldwide database, and provides information from a mountain range where no other calving glacier is currently
1.7. Study Location

Bridge Glacier (50°48′11″N, 123°38′40″W) is located in the Pacific Ranges of the Coast Mountains of southwestern British Columbia, Canada (Figure 1.5). The glacier is an outlet
of the Lillooet Icefield, one of the largest icefields in British Columbia, and is located roughly 175 km north of Vancouver, Canada. Bridge Glacier has an area of 81 km$^2$, extending from an elevation of over 2900 m at Bridge Peak to 1390 m, where it terminates in a large proglacial lake. The lake has grown from under 2 km$^2$ in 1972 to over 6 km$^2$ in 2013. The glacier has experienced large tabular calving events since the early 1990s, indicative of a floating terminus. The far (east) end of the lake houses numerous large (several hundred m$^2$) icebergs, which have been present for multiple (in some cases more than 10) years.

The region is characterized by a large precipitation gradient, and Bridge Glacier lies on the divide between the heavy precipitation coastal Pacific Ranges, and the arid interior Chilcotin Ranges. Synoptic air flow is predominantly from the west, leading to the highest elevation, most westerly reaches experiencing heavy snowfall, while the eastern flank of the glacier is more arid. Manual snow surveys at the eastern edge of Bridge Lake show mean
May 1 SWE of 600 mm (http://a100.gov.bc.ca/pub/mss/stationlist.do), while the region is characterized by cool moist winters, and warm dry summers.

Bridge Glacier is also a particularly valuable site to study due to its importance for industry, and as a site of previous glaciological work. Bridge River, which begins as an outlet of the proglacial lake in which Bridge Glacier terminates, supplies water for a hydroelectric complex consisting of 3 dams. The project has been operational since 1960, and currently supplies 6 - 8% of British Columbia’s electricity (https://www.bchydro.com/community/recreation_areas/bridge_river.html). Bridge Glacier was also a site of Environment Canada annual mass balance surveys from 1977 until the program was discontinued in 1985 (Mokievsky-Zubok et al., 1985). More recently, the glacier has been the location of several glacio-meterological, hydrological and paleo-environmental studies (Ryder, 1991; Stahl et al., 2008; Shea et al., 2009; Allen and Smith, 2007).

1.8 Thesis Structure

This thesis aims to shed light on the research goals (Section 1.6) by examining the pertinent concepts thematically. Chapter 2 presents an introduction to Bridge Glacier and outlines its recent history from 1972 to 2012. The chapter reconstructs Bridge Glacier’s retreat rate, and uses a linear inverse model to estimate the climatic-driven retreat, giving a first order estimate of the relative importance of calving.

Chapter 3 estimates surface melt during the 2013 melt season. Surface melt is modelled using a distributed energy balance model that spatially distributes meteorological variables. The model follows, and evaluates, recent work on modelling katabatic flow to account for the relatively pronounced effect of katabatic winds on temperature, humidity and wind speed across the glacier. Finally, this chapter outlines the main meteorological variables driving surface melt, and estimates how sensitive Bridge Glacier is to changing climatic conditions.

Chapter 4 presents a method for producing glacier velocity measurements from a time lapse camera. The method requires only a single camera, open source software, and basic geometry to manually measure features on-glacier. The chapter then evaluates the potential and observed errors in the method and compares results to in situ measurements. Finally this chapter presents daily and seasonal displacements from 2 sites near the terminus of Bridge Glacier.

Chapter 5 quantifies the calving flux for the 2013 season. The volume of ice discharged
from the terminus is reconstructed from satellite imagery, time lapse camera derived ter-
minus flow speeds, \textit{in situ} ice front height observations, and lake bathymetry. Individual
calving events, captured using time lapse cameras, are also related to meteorological and
limnological variables, and glacier velocities. The character and driving forces of calving
events are examined and related to the current terminus environment. Finally, the relative
importance of calving during the 2013 melt season is explored.

Chapter 6 synthesizes the main findings and implications from the 2013 melt season. This chapter explores how these findings relate to historical rates of calving and surface melt, what these findings mean for the future of Bridge Glacier, and how the relative importance of calving is be expected to change over the coming decades. Finally, this chapter examines how findings from Bridge Glacier relate to other lacustrine calving glaciers in order to identify broad commonalities and unique qualities in each system.

Chapter 7 summarizes the main findings from this study, and provides potential avenues of future work.
Chapter 2

A First Order Estimate of the Influence of Climate on Retreat

2.1 Introduction

Glaciers serve as valuable barometers of a changing climate due to the fact that variations in their length, area and mass can be tied directly to regional and global air temperatures. While this strong correlation provides valuable information for remote locations where weather station records do not exist, the climatic signal in glacier systems is obscured when calving is present. This is due to the fact that calving allows for an additional mechanism of ice loss, leading to larger volumes of ice loss than would be otherwise possible through surface melt alone. In order to isolate the climatic signal in calving glacier systems, and accurately predict future retreat, a better understanding of the relative importance and influence of calving is needed.

In other lacustrine calving systems, dramatic retreat has been found to coincide with terminus floatation (Warren and Kirkbride, 2003; Boyce et al., 2007; Dykes et al., 2011). In many cases, terminus flotation may be achieved through thinning near the terminus due to successive years of high melt rates. Enhanced thinning due to a run of hot summers has been observed at Mendenhall Glacier (Motyka et al., 2003). This climate induced thinning led to increased instability and propensity to calve, and eventually to the collapse and widespread retreat into shallower waters (Boyce et al., 2007). Similar findings have been made at Tasman Glacier, New Zealand (Warren and Kirkbride, 2003; Dykes and Brook, 2010), and in Patagonia (Skvarca et al., 2002), suggesting that retreat from climatic warming may enhance calving rates over decadal time scales.

While a few studies have examined lake-terminating glacier retreat (Haresign, 2004), the data are limited both by the number of glaciers observed and temporal extent, where most studies have focused on sub-decadal responses (Benn et al., 2007b). In order to
better understand the long-term impact of climate on glacier retreat, the full range of calving behaviour needs to be taken into account, from calving in lake-contact termini, to grounded termini, and to buoyant termini. The goal of this chapter is to broadly outline the relationship between calving and climate in a lacustrine terminating calving glacier. First, this research aims to reconstruct and quantify retreat at Bridge Glacier, British Columbia, between 1972 and 2012. Second, this chapter aims to quantify the role of calving on retreat by comparing observed rates to the expected climatic response. Finally this chapter will make a first order estimate on the relationship between climate and calving, and how that varies through the ‘life-cycle’ of a lacustrine calving glacier.

2.2 Methods

Terminus positions were obtained from Landsat imagery available using LandsatLook Viewer (http://landsatlook.usgs.gov/) through MSS, TM, ETM+ and ETM+ SLC OFF sensors. Bridge Glacier imagery was obtained between 1972 and 2012; however, before 1984 usable imagery was sparse, and terminus positions are missing from 1978 to 1981, as well as 1983. The terminus position for each year was recorded using imagery taken at the end of the melt season (August 26 to October 16), using the last available image before the glacier became snow-covered. Notable exceptions were 1972, 1973, 1975 and 1977, when imagery was only available for mid-August dates, and 1982, when a December image was the only image that season that was not cloud-covered. Three points were taken for each season’s terminus and were connected to delineate the shape of the terminus. Retreat was calculated by finding the average down-glacier change between the two annual termini positions.

Climate data were obtained from Environment Canada for Whistler, BC (50°08’ N 122°57’ W, elevation = 659 m, ID #1048898), and Vancouver International Airport, BC (49°12’ N 123°11’ W, elevation = 4 m, ID #1108447), the two closest stations with relatively complete records for the period of study (http://climate.weatheroffice.gc.ca/). Whistler had monthly records of precipitation and temperature since 1977; however, data since 2007 had not been verified or calibrated, and contained gaps and physically improbable values in the record. The Vancouver International Airport data spaned the period from 1937 to present, but contained incomplete entries for September 2010 as well as for January to May 2012. Annual winter precipitation was computed by summing precipitation between November and April (inclusive). Mean summer temperatures were obtained
by averaging mean monthly temperatures between May and October (inclusive).

2.3 Results

2.3.1 Retreat and Climate

Figure 2.1: Landsat scenes from late fall 1985, 1995, 2005, and 2012. Note the substantial thinning and disintegration of the terminus once it reaches flotation.

Landsat imagery (see http://mattchernos.files.wordpress.com/2013/03/bridge_retreat.gif, Figure 2.1) showed a relatively stable terminus position between 1972 and 1984. During this period the proglacial lake was relatively small and did not contain large icebergs. In 1989, icebergs were consistently observed in the lake, suggesting that calving was occurring. Larger icebergs, however, only began to appear in the lake in the summer of 1991. In 1994, and again in 1997, two large calving events led to a more substantial retreat. The glacier entered a period of relative stability from 1997 until 2004, at which point large crevasses formed near the terminus and became water-filled. In the summer of 2005, the terminus disintegrated and retreated 469 m, the largest calving rate during the
2.3. Results

study period. After 2005, the glacier remained at a relatively stable position until 2010, when it underwent another substantial retreat toward its present day margin.

2.3.2 Climate

For Whistler and Vancouver, annual winter precipitation is significantly correlated with the May 1 snow water equivalent for Bridge Glacier terminus, obtained between 1995 and 2012 by the Water Stewardship Division of the Government of British Columbia Ministry of Environment (Station #1C39, http://a100.gov.bc.ca/pub/mss/stationlist.do). The relation with Whistler \((n = 13)\) has a larger \(r^2\) value than that for Vancouver (Figure 2.2); however, caution is required since the sample size is larger for the Vancouver dataset \((n = 18)\).

![Figure 2.2: Relationship between winter precipitation at Whistler and Vancouver, BC and Bridge Glacier May 1 SWE.](image)

Values for both Vancouver and Whistler are significant predictors of mean annual flow (1979-2010) recorded at the outlet of Bridge Glacier (Figure 2.3), indicative of summer melt at Bridge Glacier. Hydrometric data was obtained from Environment Canada gauging station 08ME023 (50°51′22″ N, 123°27′01″ W, http://www.wsc.ec.gc.ca/applications/H20/HydromatD-eng.cfm). Vancouver is a better predictor of Bridge River mean annual
2.3. Results

flow than Whistler, suggesting it is a strong predictor of summer melt. Due to the fact that Vancouver has a longer climate record than Whistler, and a strong correlation with Bridge Glacier climatology, it has been used for all further analyses in this study.

![Graphs showing correlation between summer mean temperature and mean annual flow for Whistler and Vancouver.](image)

Figure 2.3: Mean Summer Temperature (May - October) for Whistler and Vancouver related to Mean Annual Flow at Bridge Glacier’s outlet.

Winter precipitation (Figure 2.4a) during the period oscillated in short, 3 to 5 year cycles of high and low snowfalls, with relatively large variations. Average winter precipitation for the period was 819 mm, but varied up to 400 mm in individual years. Four consecutive heavy snowfall years were observed between 1980 and 1984, while from 2000 to 2009 only two seasons experienced above average precipitation. In general, the relative lack of a long term trend or shift in precipitation patterns suggest that its impact on retreat would have been minimal, a conclusion echoed in other studies (Oerlemans, 2005; Post et al., 2011).

Summer temperatures (Figure 2.4b) show a general trend of warming during the period. A long run of warmer than average summers was observed from 1987 to 1998, with only two years falling below normals. From 1999 to 2001, three consecutive summers were cooler than normal, followed by 3 of the 5 warmest summers in the record from 2003 to 2005. Both winter precipitation and summer temperature patterns are significant predictors of Bridge Glacier equilibrium line altitudes.

The equilibrium line altitude (ELA, Figure 2.4c) showed two strongly below average seasons in 1974 and 1976. The ELA was anomalously high for 9 out of 11 seasons, between
2.3. Results

Figure 2.4: a. Vancouver winter precipitation anomaly ($\bar{x} = 819$ mm) b. Vancouver summer temperature anomaly ($\bar{x} = 14.8^\circ$C), c. Bridge Glacier equilibrium line altitude ($\bar{x} = 2089$ m), d. Bridge River mean annual flow anomaly ($\bar{x} = 10.7$ m$^3$s$^{-1}$), e. Annual retreat rate of Bridge Glacier, dashed line is loess-smoothed retreat (span = 0.5).

1985 and 1999. Save one anomalously low ELA season in 2000, only minor deviations from the mean were observed in the last 20 years. The ELA trend is well correlated with both temperature and mean annual flow anomalies.
2.3. Results

Mean annual flow from the gauging station at the outlet of Bridge Lake (Figure 2.4d.) follows the general pattern of summer temperatures and is significantly correlated with both Whistler and Vancouver, however Vancouver is a stronger predictor. Four consecutive low flow years were observed between 1983 and 1985, while between 1989 and 1998, 9 of the 10 years were above average. The period from 2003 to 2006 was characterized by another run of 4 consecutive high flow years.

Bridge Glacier’s retreat (Figure 2.4e.) was relatively consistent until 1991 when calving began to have a larger impact on terminus position. Bridge Glacier’s post-1991 retreat is characterized by large retreat years followed by periods of relative stability. The rate of retreat between 1972 and 1991 was relatively low (21 ma\(^{-1}\)), but accelerated substantially after 1991, with an average rate of 144 ma\(^{-1}\). The rate of retreat accelerated again since 2009, retreating over 1200 m in 3 years.

2.3.3 Inverse Linear Model

In order to estimate the relative roles of calving and climate, Bridge Glacier’s retreat was modelled using an inverse linear method (Oerlemans, 2005, 2007). The model treats the glacier as a one-dimensional (longitudinal) flowline that responds only to changes in climatic conditions. By modelling retreat solely as a function of climatic forcing, we can project how the glacier would have responded had it not been affected by calving. Through the use of this model, the influence of climate can be quantified, while deviations from modelled results can be hypothesized to be due to calving influence.

Table 2.1: Parameters used in the linear inverse model. Both \(c\) and \(\tau\) fit well within the range found in Oerlemans (2005).

<table>
<thead>
<tr>
<th>Model Parameters</th>
<th>Symbol</th>
<th>Value</th>
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<td>Slope</td>
<td>(\theta)</td>
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<tr>
<td>Glacier Reference Length</td>
<td>(L)</td>
<td>18 500 m</td>
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<tr>
<td>Mean Annual Winter Precipitation</td>
<td>(P)</td>
<td>819.31 mm/year</td>
</tr>
<tr>
<td>Mass Balance Gradient</td>
<td>(\beta)</td>
<td>0.00543 K/m</td>
</tr>
<tr>
<td>Climatic Sensitivity</td>
<td>(c)</td>
<td>2366.36 m/K</td>
</tr>
<tr>
<td>Response Time</td>
<td>(\tau)</td>
<td>139.03 years</td>
</tr>
</tbody>
</table>

In the linear inverse model, the retreat rate (\(\frac{dL}{dt}\)) is calculated as

\[
\frac{dL}{dt} = -\frac{1}{\tau} [c T'_(t) + L'_(t)]
\]  

(2.1a)
where $T'$ is the annual summer temperature anomaly ($^\circ$C), $c$ is the glacier’s individual climate sensitivity (m/K), and $\tau$ is response time in years, and $L'$ is the glacier’s length, relative to a reference length (see Table 2.1). The relationship can be rewritten to solve for $L'_t$ as

$$L'_t = (1 - e^{-\frac{t}{\tau}})cT'_t.$$

(2.1b)

which implies that when the time ($t$) is equal to the response time ($\tau$), roughly two-thirds of the climatic forcing has manifested itself as a change in glacier length.

In order to reconstruct retreat rates, a response time and climatic sensitivity are calculated as

$$\tau = \frac{13.6}{\theta}(1 + 20\theta)^{-0.5}L^{-0.5}$$

(2.2a)

$$c = \frac{2.3P^{0.6}}{\theta}$$

(2.2b)

where, following Oerlemans (2005), $\theta$ is the average glacier slope and $L$ is the glacier reference length in metres. The mass balance gradient (K/m), $\beta$, is calculated as

$$\beta = 0.006P^{0.5}$$

(2.2c)

and is a function of $P$, the average annual winter precipitation (mm/year).

Values of response time and climate sensitivity were derived using 1972 as the reference year (see Table 2.1). The derived response time for Bridge Glacier is 139 years, which is relatively long, but consistent with other low slope, large glaciers (Oerlemans, 2005). Climate sensitivity was increased by 50% because the accumulation area is roughly four times the ablation area, a method that is consistent with other studies, and has been confirmed in numerical simulations (Oerlemans, 2007, 2011). Therefore, the value of $c$ is well within the range of average values derived for glaciers around the world (Oerlemans, 2005).

The model is generally robust to changes in parameters (Figure 2.5). Sensitivity analyses suggest that changes in climate sensitivity and response time have a minor impact on the amplitude of response, and cannot account for the far larger observed retreat rates.

Since modelled results from Equation 2.1b are noisy, and retreat rates fluctuate by as much as 571 m, length changes are smoothed (Oerlemans, 2007) using local regression techniques (loes in the R statistical package). The loess-smoothed modelled annual retreat predicts a cumulative retreat of 613 m between 1972 and 2012. Modelled retreat
2.4 Discussion

Figure 2.5: Modelled and observed retreat at Bridge Glacier. The model is run with varied response time ($\tau$) and climatic sensitivity ($c$) parameters.

agrees quite well with observed rates until 1991, at which point observed retreat rates increase substantially while modelled retreat increases only gradually. This departure between observed and modelled retreat coincides with the onset of large calving events in Bridge Lake. In the summer of 2005, large calving events further increases the rate of terminus retreat, while the modelled rates do not vary significantly, leading to a further departure between the two retreat rates.

2.4 Discussion

The general agreement between the observed and modelled results up to 1991 suggests that, between 1972 and 1991, Bridge Glacier’s retreat was directly related to climatic conditions.
2.4. Discussion

Between 1972 and 1991, anomalously hot summers produced retreat consistent with the derived response times and climate sensitivity. Since 1991, however, observed retreat was at least partially independent of climatic forcing. The acceleration of retreat, asynchronous to climate trends, suggests that Bridge Glacier’s retreat was less directly influenced by climatic conditions once calving began. After the onset of calving, the glacier exhibited a staggered retreat, characterized by larger annual variations than could be expected from climate-driven retreat alone, followed by periods without any significant retreat.

Although calving modifies the relationship between climate and glacier retreat, climate may still play an important role in the long term retreat of calving glaciers. Many studies have found that flotation is a key variable in determining the stability and retreat of water-terminating glaciers (Meier and Post, 1987; Van der Veen, 2002); however, the mechanisms of thinning are diverse.

For Bridge Glacier, calving only had a measurable effect on the terminus position once it retreated into deeper water and achieved floatation, a finding is supported by the large, infrequent tabular icebergs produced from the terminus. Similar observations have been made at Mendenhall Glacier in Alaska (Motyka et al., 2003; Boyce et al., 2007), where series of high melt summers preceded flotation and large scale retreat. Climatic indicators for Bridge Glacier suggest strongly negative mass balance years from 1986 to 1991 (Figure 2.4), which would have increased the rate of glacier thinning and created the conditions necessary for terminus floatation, and enhanced the rate of calving.

While it is possible that other factors could have affected the terminus thinning during the period leading up to the floatation of Bridge Glacier’s terminus, these scenarios are much less probable. It is possible that increased basal water due to enhanced melt rates (or an earlier, more prolonged melt season) would have allowed for greater flow speeds, enhancing strain rates and dynamic thinning. It is also possible that thinning could have been produced by a decrease in flow speeds in the lowest reaches of the glacier, leading to a lower supply of ice to the terminus, and ice stagnation. Both of these conditions, however, require a systematic shift in the dynamics of the glacier system before the onset of calving. While they cannot be ruled out completely, it is unlikely that these these scenarios can fully explain the significant increase in Bridge Glacier’s retreat rate since they do not correspond with the observed climatic trend of warming preceding the increased rate of retreat.

Once calving begins to have a significant impact on the retreat rate of the glacier it also plays a role in driving for glacier dynamics. Once Bridge Glacier reaches flotation, calving becomes a more important source of ice loss. Increased ice loss, leading to terminus
2.5. Conclusion

In this chapter, the retreat of Bridge Glacier has been reconstructed and has been related to climate through a simulation of its linear response to climatic forcing. From 1972 to 2012, Bridge Glacier has retreated 3.55 km. The retreat rate was modest up until 1991, with less than 600 m of cumulative retreat. Since 1991, however, retreat accelerated, culminating in retreat rates of over 400 ma$^{-1}$ between 2009 and 2012. The post-1991 retreat is irregular,
characterized by large tabular calving events and high retreat years, and periods of relative terminus stability. The high magnitude, low frequency tabular character of observed calving events suggest that calving is primarily due to terminus floatation (Benn et al., 2007b).

Observed retreat follows modelled retreat projected from climatic conditions until 1991; at which point retreat rates far exceed modelled results. Bridge Glacier was retreating in sync with climate until 1991, however, since then the retreat rate is driven primarily by calving. Modelled results suggest that calving is responsible for an additional 2.5 - 3.0 km of retreat, suggesting that since 1991, calving has been the dominant control on the terminus position.

Although these findings emphasize the relative importance of calving in Bridge Glacier’s recent retreat, climatic records and modelling suggest that an initial climatic warming is necessary to create the conditions necessary for calving to become an important component of retreat. A series of consecutive high ablation summers is necessary to thin the lowest reaches of the glacier enough to allow the the terminus to become buoyant. Once the terminus becomes buoyant, however, retreat can continue independent of the larger climatic trend due to the vulnerability of the terminus to calving. Terminus flotation and increased calving rates in turn promote higher terminus velocity, accelerating terminus thinning and instigating a positive feedback loop where a calving-driven retreat can continue independent of climatic conditions.

This first order estimate of calving at Bridge Glacier suggests that while the inverse linear model employed in this chapter attributes the vast majority of observed retreat in the last 40 years to calving, calving itself acts within the initial bounds of climate. Within the ‘life-cycle’ of a lacustrine calving glacier, climatic forcing is necessary to create the conditions necessary to promote further calving retreat.

This work suggests that calving is responsible for the majority of the observed retreat during this period; however, the same implication does not necessarily hold true for Bridge Glacier’s mass balance during the period. Beyond the implication that terminus buoyancy is critical to determining the calving rate, this work does not elucidate any of the controls of calving itself on Bridge Glacier. In order to better constrain the role and importance of calving to the overall ‘health’ of Bridge Glacier, a quantitative account of the volumetric ice loss from both mechanisms is required. These questions, among others, are the overarching concepts that drive research in subsequent chapters.
Chapter 3

A Distributed Energy Balance Model for Ice Ablation

3.1 Introduction

Since the Little Ice Age, the cryosphere, and mountain glaciers in particular, have been receding at an accelerated pace (Barry, 2006). Although irregular, this retreat is pervasive across regions within Canada and across the globe, and is well correlated to rising mean global temperatures (Oerlemans, 2005). This retreat has significant ramifications for future volume and timing of water resources, as well as future eustatic sea level rise (Dyurgerov and Meier, 2005), and requires a detailed predictive system to ensure proper management over short and long timescales.

Numerical melt models allow for a quantification of meltwater generation from glacier and snow surfaces by relating surface melt to meteorological or climatological variables. Melt models vary substantially in complexity and data requirements and have been shown to have high predictive value over a range of temporal and spatial scales (Hock, 2005). Empirical frameworks, such as Positive Degree Day models (Braithwaite and Olesen, 1989) and Temperature Index models (Hock, 2003), which relate melt to air temperature, can provide very good daily to seasonal predictions. Empirical temperature models are desirable in many cases due to few data requirements and their ability to out-perform physically based models in locations where detailed reliable meteorological data cannot be obtained (Shea, 2010).

3.1.1 Surface Energy Balance Modelling

Surface energy balance models offer an alternative to empirical models by explicitly accounting for the fluxes of energy available for melt at the glacier surface. While energy balance models have higher predictive power over short timescales, they also require sub-
3.1. Introduction

stantially more input variables, increasing the complexity of the model (Hock, 2005). Furthermore, the transferability of energy balance model input variables is limited, necessitating detailed in situ meteorological observations to accurately capture melt processes (Anslow et al., 2008; MacDougall and Flowers, 2011). Notwithstanding, energy balance models are preferable in many cases because they also offer the added benefit of quantifying the relative importance of individual melt processes, allowing for future projections of the sensitivity of glacier melt to a changing climate.

Surface energy balance models are most commonly made at a single point on the glacier (Munro, 2006; Brock et al., 2000). Typically, a weather station is installed at a location within the ablation area that measures all relevant energy balance variables. While this method is logistically superior, it does not account for spatial variability in relevant meteorological variables, and relies on the representativeness of a single point to accurately capture the main drivers of melt over the glacier.

As an extension of point energy balance models, distributed energy balance models present a more complex model that can account for the spatial variation in melt across the glacier. The energy available for melt is calculated for each grid point on the glacier by interpolating or distributing meteorological parameters spatially. With this increase in spatial complexity, distributed models have larger data requirements, including a digital elevation model, and a routine to spatially distribute turbulent flux components.

**Turbulent Heat Parameter Distributions**

Sensible and latent heat fluxes are important components of melt energy supply, and are controlled by temperature and vapour pressure gradients, as well as wind speed. These turbulent fluxes distribute predominantly as a function of elevation; however, in glacierized basins this spatial and altitudinal distribution is disrupted by the development and presence of katabatic winds (Shea and Moore, 2010). Air directly above the glacier surface is cooled, creating denser, less buoyant air, and initiating down-valley flow. As air flows downslope it is further cooled, creating a persistent down-glacier wind (Munro, 2006). Katabatic flows create a sharp, shallow thermocline (between 5 and 20 m above the surface) where temperatures have been observed to be significantly cooler than the ambient air temperature (Oerlemans, 2010; Shea and Moore, 2010).

During the melt season, winds over the glacier surface are predominantly characterized by a down-glacier katabatic flow (Oerlemans, 2010; Oerlemans and Reichert, 2000), while
3.2. Instrumentation and Field Methods

observations in the Coast Mountains found that the presence of katabatic flow is dependent on a threshold temperature (Shea and Moore, 2010). Using piecewise regression, katabatic effects were found only when the ambient air temperature was greater than 4°C. This observation was corroborated on Pasterze Glacier, Austria, where higher downslope wind speeds correlate well with higher ambient air temperatures (Greuell and Smeets, 2001).

While many studies have used a constant wind speed measured at on-glacier weather stations to distribute wind speed across the glacier (Hock and Holmgren, 2005; MacDougall and Flowers, 2011), this distribution scheme does not take into account the inherent complexity of katabatic flow. Expected spatial variability in katabatic wind speed magnitudes, as well as terrain sheltering and exposure across the glacier, suggest that a constant wind speed across the glacier is unrealistic, and will ultimately reduce the accuracy of turbulent heat calculations. Similarly, temperature is commonly distributed using a $-6 \, ^\circ \text{C} \, \text{km}^{-1}$ lapse rate (Hock, 2005), or an average measured seasonal lapse rate (Anslow et al., 2008; MacDougall and Flowers, 2011). However, the cooling of on-glacier air temperatures by katabatic flow ultimately suggests that standard atmospheric lapse rates are not suitable to distribute on-glacier air temperatures (Munro, 2006; Shea and Moore, 2010).

As an alternative, recent work has focused on modelling the magnitude and extent of katabatic winds across a range of glacier sizes (Shea, 2010). Using terrain morphometry, katabatic wind is shown to be stronger in the lowest reaches of the glacier, and can significantly depress temperature and vapour pressure (Shea and Moore, 2010), and dominate the character of valley winds (Petersen and Pellicciotti, 2011). Since katabatic wind can significantly enhance turbulent heat fluxes (Munro, 2006), a full evaluation of katabatic forcing is required in order to accurately spatially and temporally model melt for glaciers with a strong katabatic influence.

This chapter addresses three main themes. First a distributed energy balance model is used to calculate the volume of ice lost at Bridge Glacier during the 2013 melt season. Secondly, this model elucidates the main drivers of melt during the 2013 melt season, and examines how those drivers vary both spatially and temporally. Finally, the work of Shea and Moore (2010) is used to account for the strong influence of katabatic wind on the distribution of turbulent heat flux parameters, and compare the melt projections of this model against more commonly used turbulent parameter distribution schemes.
Figure 3.1: Bridge Glacier study area, instrumentation, and 2013 terminus position. Contour intervals are 100 m.

### 3.2 Instrumentation and Field Methods

Between June 20 and September 12, 2013, data were collected from 3 automatic weather stations (AWS) (see Figure 4.1). The majority of the data were taken from an AWS installed on-glacier (Glacier AWS, 50°49'02" N 123°33'35" W, 1607 m a.s.l.). A second AWS (Ridge AWS, 50°50'33" N 123°31'23" W, 1639 m a.s.l.) was used for ambient temperature data, and was shielded from strong, persistent katabatic flow observed throughout the melt season, measuring ambient meteorological conditions. A third weather station was located at the shore of the lake (Lake AWS, 50°50'25" N 123°30'07" W, 1393 m a.s.l.), along a terminal moraine, and was used to measure incoming longwave radiation. All meteorolog-
3.2. Instrumentation and Field Methods

Table 3.1: Variables measured at on-site automatic weather stations. All variables were measured at 1.75 m above the surface.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Symbol</th>
<th>Units</th>
<th>Accuracy</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Glacier AWS</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Reflected Radiation</td>
<td>$K \uparrow$</td>
<td>$\text{Wm}^{-2}$</td>
<td>&lt;1%</td>
</tr>
<tr>
<td>Temperature</td>
<td>$T_g$</td>
<td>°C</td>
<td>0.21°C</td>
</tr>
<tr>
<td>Relative Humidity</td>
<td>RH%</td>
<td>%</td>
<td>2.5%</td>
</tr>
<tr>
<td>Wind Speed</td>
<td>$u$</td>
<td>$\text{ms}^{-1}$</td>
<td>4%</td>
</tr>
<tr>
<td>Wind Direction</td>
<td>$u_d$</td>
<td>°</td>
<td>5°</td>
</tr>
<tr>
<td>Atmospheric Pressure</td>
<td>$P$</td>
<td>Pa</td>
<td>3.0 mbar</td>
</tr>
<tr>
<td>Rainfall Rate</td>
<td>$R$</td>
<td>mm hr$^{-1}$</td>
<td>1.0%</td>
</tr>
<tr>
<td>Melt Rate</td>
<td>$M$</td>
<td>Pa</td>
<td>0.05 m d$^{-1}$</td>
</tr>
<tr>
<td><strong>Ridge AWS</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Temperature</td>
<td>$T_{amb.}$</td>
<td>°C</td>
<td>0.2°C</td>
</tr>
<tr>
<td>Incoming Shortwave Radiation</td>
<td>$K \downarrow$</td>
<td>$\text{Wm}^{-2}$</td>
<td>&lt;1%</td>
</tr>
<tr>
<td><strong>Lake AWS</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Incoming Longwave Radiation</td>
<td>$L \downarrow$</td>
<td>$\text{Wm}^{-2}$</td>
<td>&lt;1%</td>
</tr>
</tbody>
</table>

*Approximate value, see Appendix A.1 for further information.

\[\text{The authors gratefully acknowledge M. Demuth, C. Hopkinson, and B. Menounos for the LiDAR data used in this study, which was collected as part of a C-CLEAR effort to develop LiDAR environmental applications, and funded in part by the Western Canadian Cryospheric Network (WCCN).}\]
3.3 Energy Balance Modelling Methods

Glacier surface melt \((M)\) is expressed in metres of water equivalent melt per day and is calculated as

\[
M = \frac{Q_M}{L_f \rho_i}
\]  

(3.1a)

where \(Q_M\) is the sum of available energy at the surface (Wm\(^{-2}\)), \(L_f\) is the latent heat of fusion, and \(\rho_i\) is the density of ice. Energy supplied to the glacier surface is taken as positive, while energy flux away from the surface is negative. The sum of available melt depth for each day, allowing daily melt rates to be obtained (see Appendix A.1 for further information). Using the pressure transducer data, the date of the ablation stake (A) at Glacier AWS melting out was determined to be August 28 (the date at which the pressure transducer also reached the surface). This date is assumed constant for all other stakes, although it is likely that there is an uncertainty of up to two to three days for individual stakes.
3.3. Energy Balance Modelling Methods

Energy can be further divided as

\[ Q_M = Q^* + Q_H + Q_E + Q_R \]  \hfill (3.1b)

where the net radiation \((Q^*)\), sensible heat flux \((Q_H)\), latent heat flux \((Q_E)\), and rain heat flux \((Q_R)\) can all contribute melt energy to the glacier surface. It is assumed that all energy fluxes occur at the glacier surface (Oerlemans, 2010; Munro, 2006), and subglacial melt is neglected. All calculations were completed using R Statistical Software unless otherwise noted.

Table 3.2: Variables used in this study.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Variable</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>(I_0)</td>
<td>Solar Constant</td>
<td>1366.5 W m(^{-2})</td>
</tr>
<tr>
<td>(\varphi)</td>
<td>Glacier Latitude</td>
<td>50.81(^\circ)</td>
</tr>
<tr>
<td>(\psi_a)</td>
<td>Clear Sky Transmissivity</td>
<td>0.75</td>
</tr>
<tr>
<td>(P_o)</td>
<td>Mean Sea Level Pressure</td>
<td>1013.15 hPa</td>
</tr>
<tr>
<td>(\Gamma_P)</td>
<td>Pressure Lapse Rate</td>
<td>-0.12 hPa m(^{-1})</td>
</tr>
<tr>
<td>(\Gamma_T)</td>
<td>Temperature Lapse Rate</td>
<td>-0.006 °C m(^{-1})</td>
</tr>
<tr>
<td>(\varepsilon_{\text{terrain}})</td>
<td>Terrain emissivity</td>
<td>0.95</td>
</tr>
<tr>
<td>(\varepsilon_{\text{ice}})</td>
<td>Ice emissivity</td>
<td>0.98</td>
</tr>
<tr>
<td>(\alpha_{\text{terrain}})</td>
<td>Terrain albedo</td>
<td>0.17</td>
</tr>
<tr>
<td>(g)</td>
<td>Gravitational constant</td>
<td>9.81 m s(^{-1})</td>
</tr>
<tr>
<td>(c_a)</td>
<td>Specific heat capacity of air</td>
<td>1006 J kg(^{-1})K(^{-1})</td>
</tr>
<tr>
<td>(c_w)</td>
<td>Specific heat capacity of water</td>
<td>4200 J kg(^{-1})K(^{-1})</td>
</tr>
<tr>
<td>(k)</td>
<td>von Karman constant</td>
<td>0.41</td>
</tr>
<tr>
<td>(z_o)</td>
<td>Roughness length of momentum</td>
<td>2.5 \times 10^{-3} m</td>
</tr>
<tr>
<td>(z)</td>
<td>Measurement height</td>
<td>1.75 m</td>
</tr>
<tr>
<td>(L_v)</td>
<td>Latent heat of vapourization</td>
<td>2.26 \times 10^6 J kg(^{-1})</td>
</tr>
<tr>
<td>(L_f)</td>
<td>Latent heat of fusion</td>
<td>3.34 \times 10^6 J kg(^{-1})</td>
</tr>
<tr>
<td>(\rho_i)</td>
<td>Density of ice</td>
<td>917 kg m(^{-3})</td>
</tr>
<tr>
<td>(\rho_w)</td>
<td>Density of water</td>
<td>1000 kg m(^{-3})</td>
</tr>
</tbody>
</table>

3.3.1 Net Radiation

Net radiation \((Q^*)\) is calculated as the sum of incoming (\(\downarrow\)) and outgoing (\(\uparrow\)) shortwave \((K)\) and longwave \((L)\) radiation as follows:

\[ Q^* = (K \downarrow - K \uparrow) + (L \downarrow - L \uparrow). \]  \hfill (3.2)
3.3. Energy Balance Modelling Methods

To spatially distribute net radiation, it is rewritten as

\[ Q^* = (S \downarrow + D \downarrow)(1 - \alpha_{ice}) + (L \downarrow - L \uparrow) \]  \hspace{1cm} (3.3)

where shortwave radiation is separated into direct (S) and diffuse (D) components.

**Shortwave Radiation**

Direct radiation (Wm\(^{-2}\)) for each point, \(S \downarrow\), is calculated as

\[ S \downarrow = (1 - \frac{D \downarrow}{K \downarrow})K \downarrow \frac{K_{ex_{i,j}}}{K_{ex}} \]  \hspace{1cm} (3.4)

by relating the ratio of direct shortwave radiation at Glacier AWS and observed incoming shortwave radiation to the ratio of potential direct radiation at the grid point \((K_{ex_{i,j}})\) and at Glacier AWS \((K_{ex})\). The potential solar radiation at the top of the atmosphere is calculated as

\[ K_{ex} = I_o \left(\frac{R_m}{R}\right)^2 \cos \theta \]  \hspace{1cm} (3.5a)

where \(I_o\) is the solar constant, \(R_m\) and \(R\) are the mean and instantaneous Sun-Earth distances, and \(\theta\) is the angle of incidence (see Appendix A.2 for solar geometry calculations). Potential direct radiation was corrected for slope geometry, and the angle of incidence \((\theta)\) where

\[ \cos \theta = \cos Z \cos \beta + \sin Z \sin \beta \cos(\Omega - A) \]  \hspace{1cm} (3.5b)

and \(Z\) is the elevation angle, \(\beta\) is the grid cell slope, \(\Omega\) the solar azimuth, and \(A\) is the aspect.

Diffuse shortwave radiation for all cells when \(K_{ex}\) is greater than 0 is calculated as

\[ D = K \downarrow \left(\frac{D \downarrow}{K \downarrow}\right)\phi + \alpha_{terrain}K \downarrow (1 - \phi) \]  \hspace{1cm} (3.6)

where \(\phi\) is sky view factor for each grid cell (Hock and Holmgren, 2005; MacDougall and Flowers, 2011).

Due to the complications and heterogeneity involved in measuring the albedo \((\alpha)\) for the surrounding terrain, a constant value of 0.17 was assumed, which is well within the range for dark, rocky surfaces (Oke, 1988). The errors associated with this assumption should be minor, since the sky view factor for most grid cells is greater than 0.9.
3.3. Energy Balance Modelling Methods

The ratio of incoming diffuse to shortwave radiation is calculated as

\[ \frac{D}{K} = \begin{cases} 
0.929 + 1.134 \frac{K_{\downarrow}}{K_{ex}} - 5.111 (\frac{K_{\downarrow}}{K_{ex}})^2 + 3.106 (\frac{K_{\downarrow}}{K_{ex}})^3 & 0.15 < \frac{K_{\downarrow}}{K_{ex}} \leq 0.80 \\
1.0 & \frac{K_{\downarrow}}{K_{ex}} > 0.80 \\
0.15 & \frac{K_{\downarrow}}{K_{ex}} \leq 0.15 
\end{cases} \quad (3.7) \]

where \( \frac{K_{\downarrow}}{K_{ex}} \) is the ratio of observed to potential shortwave radiation at the weather station. The method follows empirical relationships derived by Collares-Pereira and Rabl (1979), and re-confirmed with a different dataset by Hock and Holmgren (2005).

In order to spatially distribute incoming shortwave radiation, each grid point is modelled as either shaded or sunny. A shading algorithm was implemented that calculates the maximum horizon angle for each grid point within a 10 km² window, using 10° azimuth bins. At each time step, if the horizon angle is greater than the elevation angle (Z), the grid point is shaded, and only receives diffuse radiation. For times when the horizon angle is smaller than elevation angle, the grid point receives both direct and diffuse radiation.

Sky view factor was calculated using SAGA GIS software and a 25 m lidar DEM. The algorithm,

\[ \phi = \frac{1}{2\pi} \int_{0}^{2\pi} \cos^2 H d\Omega, \quad (3.8) \]

integrates the maximum horizon angles (H) for each grid cell, for each azimuth angle (1° interval). A maximum search window of 10 km² was implemented to reduce computation time.

**Longwave Radiation**

Incoming longwave radiation (\( L_{\downarrow} \)) was measured directly at Lake AWS. In order to distribute longwave radiation across the glacier, it is scaled by the sky view factor (\( \phi \)) as

\[ L_{\downarrow} = L_{\text{in measured}} \frac{\phi_{i,j}}{\phi_{\text{aw}}} + L_{\text{terrain}} (1 - \phi_{i,j}) \quad (3.9) \]

where additional longwave input is supplied by the surrounding terrain (\( L_{\text{terrain}} \)). Terrain temperature is assumed to equal air temperature, and the Stefan-Boltzmann equation is used with an emissivity (\( \varepsilon_{\text{terrain}} \)) of 0.95 (Oke, 1988).
3.3. Energy Balance Modelling Methods

Outgoing longwave can be calculated as

\[ L^{\uparrow} = \varepsilon_{\text{ice}} \sigma T^4 + (1 - \varepsilon_{\text{ice}}) L^{\downarrow} \quad (3.10) \]

where an ice emissivity (\( \varepsilon \)) of 0.98 is used, \( \sigma \) is the Stefan-Boltzmann constant, and \( T \) is the ice temperature. It is assumed that during the melt season the glacier surface was constantly at the melting point (273.15 K).

**Albedo and Net Radiation**

Reflected shortwave radiation was measured on-glacier, in the ablation area, throughout the melt season using a gimbal joint designed to keep the pyranometer pointed directly downwards. Due to difficulties ensuring a pyranometer attached to the glacier weather station would remain level, incoming shortwave measurements were taken from the off-glacier Ridge AWS. The differences in shading were found to be minor. To eliminate small discrepancies in shading, uneven cloud patterns, and low solar angle errors (Oerlemans, 2010), ice albedo (\( \alpha \)) is calculated as

\[ \alpha_{\text{daily}} = \frac{\int K^{\uparrow} dt}{\int K^{\downarrow} dt} \quad (3.11) \]

for all daylight hours, and is assumed constant throughout the day. The weather station was located on bare ice for the entire study, and since this study only concerns ice loss, snow albedo was not considered.

**3.3.2 Turbulent Heat Fluxes**

Sensible heat (\( Q_H \)) and latent heat (\( Q_E \)) are important components of melt energy availability, and can be expressed according to the bulk transfer approach as

\[ Q_H = \rho_{\text{air}} c_a C u (T_a - T_g) \quad (3.12) \]

\[ Q_E = \rho_{\text{air}} L_v C u \left( \frac{0.622 (e_a - e_g)}{P} \right) \quad (3.13) \]

where \( c_{\text{air}} \) is the specific heat capacity of air, \( C \) is a turbulent transfer coefficient, \( u \) is the windspeed, \( T \) is air temperature, \( L_v \) is the latent heat of vaporization, \( e \) is the vapour pressure, and \( P \) is the atmospheric pressure at the surface (see Table 3.2). The gradients for
3.3. Energy Balance Modelling Methods

Air temperature and vapour pressure are denoted where the subscript $g$ refers to on-glacier, and $a$ refers to off-glacier ‘ambient’ conditions.

In order to reflect the significant forcing of katabatic winds on turbulent flux variables, glacier morphometry is used to model the extent and magnitude of katabatic forcing. Using flow path lengths, defined as the average flow distance to a given point from an upslope summit or ridge, statistical relationships are formulated to spatially distribute temperature and wind speed across the glacier. Flow path lengths (FPL) for the glacier were calculated using a 25 m lidar DEM and a routine in the Terrain Analysis - Hydrology module of SAGA GIS (Quinn et al., 1991), and is shown in Figure 3.3. Since downslope flow drives katabatic forcing, flow path lengths allow for a quantification of the magnitude of relative forcing across the glacier (Shea and Moore, 2010).
Distributing Temperature

The magnitude of katabatic forcing was modelled as a function of the temperature difference ($\Delta T$) between on-glacier (Glacier AWS) and off glacier (Ridge AWS) (outside the katabatic boundary layer) weather stations. Temperature differences were separated into upslope (northeasterly) and downslope katabatic (southwesterly) periods, based on the wind directions of Glacier AWS. Linear regressions of both groups against ambient (off-glacier) temperature (Figure 3.4, $T_a$, from Ridge AWS) show a strong linear increase in $\Delta T$ with increasing ambient temperatures for katabatic flows, indicating the magnitude of katabatic forcing increases with increasing ambient temperatures. Conversely, $\Delta T$ does not significantly vary as a function of ambient temperatures during upslope flow, although ambient temperature above 10 °C during these periods was rare. The elevation of both weather stations are within 100 m, and small corrections to potential temperature, using a $-6°C km^{-1}$ lapse rate, did not produce a meaningful difference in the linear fit.

Figure 3.4: Katabatic on-glacier cooling, as a function of ambient temperature from Ridge AWS.
3.3. Energy Balance Modelling Methods

On-glacier temperature for each grid point is modelled as a function of the katabatic temperature depression where

\[ T_g = T_a - (k_1 T_a + \Delta T^*) \]  

(3.14a)

and \( k_1 \) is the magnitude of katabatic forcing for each point on the glacier, \( T_a \) is the off-glacier (ambient) air temperature, and \( \Delta T^* \) is the threshold temperature differential at which katabatic flow is observed.

The magnitude of katabatic forcing for each point on glacier is calculated by using statistical coefficients and glacier flow path lengths (FPL) where

\[ k_1 = \beta_1 \exp(\beta_2 FPL) \]  

(3.14b)

and the empirical coefficient \( \beta_2 \) is taken from Shea (2010). The coefficient is derived using data from multiple elevations and multiple glacier basins in the southwestern Coast Mountains (see Table 3.3). The coefficient \( \beta_1 \) is calculated as

\[ \beta_1 = \frac{m_T}{\exp(\beta_2 FPL_{aws})} \]  

(3.14c)

where FPL and the degree of katabatic forcing \( (m_T) \) values are extracted from the location of Glacier AWS.

The coefficients \( m_T \) and \( \Delta T^* \) can be derived from linear regression (Figure 3.4), where

\[ \Delta T = m_T T_a + \Delta T^* \]  

(3.14d)

and the difference between on-glacier and off-glacier air temperatures \( (\Delta T) \) is regressed against ambient air temperatures (from Ridge AWS).

In the rare case that wind direction is upslope, temperatures are distributed using on-glacier temperature, \( T_g \), and a standard temperature lapse rate of \(-6^\circ C \text{ km}^{-1}\).
3.3. Energy Balance Modelling Methods

Table 3.3: Statistical coefficients used in FPL modelling.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\beta_1$</td>
<td>$3.90 \times 10^{-1}$</td>
</tr>
<tr>
<td>$\beta_2$</td>
<td>$4.43 \times 10^{-5}$*</td>
</tr>
<tr>
<td>$\beta_3$</td>
<td>$6.70 \times 10^{-2}$</td>
</tr>
<tr>
<td>$\beta_4$</td>
<td>$-3.39 \times 10^{-1}$*</td>
</tr>
<tr>
<td>$\Delta T^*$</td>
<td>$-1.67 ^\circ C$</td>
</tr>
<tr>
<td>$u^*$</td>
<td>$1.08 , \text{m s}^{-1}$</td>
</tr>
</tbody>
</table>

*From Shea and Moore (2010)

Distributing Vapour Pressure

Vapour pressure is calculated from measured relative humidity and saturation vapour pressure ($e_s$). Saturated vapour pressure is calculated using Teten’s formula, where

$$
\begin{align*}
    e_s &= \begin{cases} 
        6.11 \times 10^{\left(\frac{7.57}{237.7}T\right)} & \text{if } T \geq 0 \\
        6.11 \times 10^{\left(\frac{9.57}{265.57}T\right)} & \text{if } T < 0 
    \end{cases} \\
    &\quad (3.15a)
\end{align*}
$$

and $T$ is the air temperature. Vapour pressure ($e$) can be related to relative humidity ($RHI\%$) as follows:

$$
    e = e_s \frac{RHI}{100}. \quad (3.15b)
$$

Since only a weak relationship was found between on-glacier vapour pressures and off-glacier temperatures, relative humidity measured at Glacier AWS is held spatially constant across the glacier for each timestep, and saturation vapour pressure is calculated from distributed on-glacier air temperatures.

Distributing Wind Speed

Glacier wind speed was distributed in a similar fashion to air temperatures, as a function of katabatic forcing and ambient temperatures, following Shea (2010). On-glacier wind speed is regressed against ambient air temperature. For downslope winds, wind speed increases linearly with increasing ambient air temperature, while upslope wind speeds show no significant change with ambient temperature (see Figure 3.5).

On-glacier wind speed for each grid point is modelled as a function of off-glacier tem-
3.3. Energy Balance Modelling Methods

Figure 3.5: Wind speed as a function of ambient (off-glacier) air temperature.

Temperature where

\[ u_g = u_1 T_a + u^* \]  \hspace{1cm} (3.16a)

and \( u_1 \) is the magnitude of katabatic forcing for each point and \( u^* \) is the katabatic wind speed when the off-glacier temperature is 0°C.

The magnitude of katabatic forcing on wind speed for any point on the glacier calculated using statistical coefficients and glacier flow path length as

\[ u_1 = \beta_4 + \beta_3 \log(FPL) \]  \hspace{1cm} (3.16b)

where the empirical coefficient \( \beta_4 \) is taken from Shea and Moore (2010). The coefficient \( \beta_3 \) is calculated as

\[ \beta_3 = \frac{m_u - \beta_4}{\log(FPL_{aws})} \]  \hspace{1cm} (3.16c)
3.3. Energy Balance Modelling Methods

where FPL and the degree of katabatic forcing on wind speed \( (m_u) \) values are extracted from the location of Glacier AWS.

The coefficients \( m_u \) and \( u^* \) can be derived from linear regression (Figure 3.5), where

\[
  u_g = m_u T_a + u^*
\]

and the on-glacier wind speed is regressed against off-glacier air temperatures. When the wind direction is upslope, on-glacier wind speed is held constant, using measured wind speeds from Glacier AWS.

Physical Limitations of Flow Path Length Modelling

This model has the major drawback in that \( FPL < 1 \) m results in a negative wind speed. In order to mitigate this drawback, points with no katabatic forcing \( (u_1 \leq 0) \) are assumed to have the same wind speed as the off-glacier Ridge AWS. The vast majority of these points occur near ridges and peaks, and are unlikely to be affected by katabatic flow. Secondly, the physical meaning of \( u^* \) breaks down at low temperatures, theoretically producing katabatic winds when the ambient air temperature is colder than the glacier surface \( (T_{amb.} < 0^\circ C) \). Since sub-zero temperatures were not observed during the study period, the potential of a non-linear fit at low temperatures cannot be tested. It is likely, however, that the surface of the glacier would be less than 0°C during events with low ambient air temperatures, and would not be melting substantially.

Similar limitations exist in distributing on-glacier temperatures. The calculated value of \( \Delta T^* \) is \(< 0^\circ C\), suggests that a katabatic temperature depression occurs even when glacier air temperatures are higher than ambient values. This phenomenon is not observed during the study period, and seemingly contradicts expected katabatic behavior. However, forcing the linear fit through the expected value of \( \Delta T^* = 0^\circ C \) substantially reduces the linear fit. Similar to wind speed limitations, it is likely that katabatic forcing is uncommon at lower temperatures during the summer melt period, an assumption supported by temperatures of roughly and below 5°C, exhibiting predominantly upslope wind directions.

Finally, while \( \beta_1 \) and \( \beta_3 \) were recalculated, and could be compared with values derived from Shea and Moore (2010), variables \( \beta_2 \) and \( \beta_4 \), which scale FPL magnitude spatially, could not be tested with the current 2013 field dataset since it requires data from multiple elevations, and were therefore assumed to be the same.
3.3. Energy Balance Modelling Methods

Turbulent Transfer Coefficient

Using the bulk aerodynamic method (Moore, 1983), the turbulent transfer coefficient, $C$ (dimensionless), is calculated using a stability correction ($\Theta$) and surface roughness lengths ($z_o$ and $z_x$) as follows:

$$ C = \Theta k^2 \left[\log \left(\frac{z}{z_o}\right) \log \left(\frac{z}{z_x}\right)\right]^{-1}. \quad (3.17a) $$

The magnitude of the stability correction is calculated as

$$ \Theta = \begin{cases} 
(1 - 5.2 R_b)^2 & R_b > 0 \\
(1 - 16.0 R_b)^{0.75} & R_b < 0 
\end{cases} \quad (3.17b) $$

where positive bulk Richardson Numbers ($R_b$) indicates a stable atmosphere. The bulk Richardson Number is calculated as

$$ R_b = \frac{g (T_a - T_o)z}{T_g + 273.15 u^2} \quad (3.17c) $$

where $g$ is the gravitational constant, and $z$ is the height of measurement (m).

The roughness length for momentum ($z_o$) is taken as 2.5 mm for ice (Munro, 1989), while the roughness length for temperature and vapour pressure ($z_x$) is calculated as $z_o/300$ following Hock (1998). Both roughness lengths are assumed constant in space and time within the ablation area.

Rain Heat Flux

Although relatively minor, energy supplied to the surface due to rain is calculated as

$$ Q_R = \rho_w c_w R T_R \quad (3.18) $$

where $R$ is the rainfall rate (ms$^{-1}$), $T_R$ is the temperature of rain, which is assumed the same as the ambient (off-glacier) air temperature, and $\rho_w$ and $c_w$ are the density and specific heat of water (see Table 3.2).

3.3.3 Snowline Retreat

Since an adequate spring snow survey was not logistically possible, and proxy snow surveys were considered unreliable due to large precipitation gradients in the region (Stahl
3.3. Energy Balance Modelling Methods

Figure 3.6: Modelled snowline retreat from Landsat data. Dashed line is loess smoothed retreat.

et al., 2008; Shea and Moore, 2010), snowline retreat was reconstructed using Landsat images obtained from the LandsatLook Viewer (http://landsatlook.usgs.gov/). Multiple measurements were taken for each scene (9 unique dates throughout the study period), and averaged to produce a basin-wide snowline elevation. Interpolation between points was achieved using the loess smoothing function in R (Figure 3.6). Substantial variability exists in the snow coverage in the north and south channels; however, the errors associated with this interpolation should be relatively minor in the overall summer ablation. In order to calculate the volume of ice melt, any melt above the snowline, at each time step, is set to 0.

3.3.4 Complementary Melt Rates and Meteorology

As an alternative method of deriving and constraining 2013 melt, melt rates are also modelled using a daily temperature index model (TIM) (Hock, 2003; Shea et al., 2009) as follows:

\[ M = DDF \times T_a. \]  

(3.19a)
3.4 Results and Discussion

The degree day factor (DDF) is calculated as

\[ DDF = \frac{\sum M}{\sum T_a^+} \]  

where \( T_a^+ \) is the positive off-glacier air temperature (from Ridge AWS), and \( M \) is the measured seasonal ablation from ablation stakes near Glacier AWS.

3.4 Results and Discussion

3.4.1 Volumetric Ice Melt

Throughout the study period of June 20 to September 12, 2013, the model predicted ice melt of 1.0 m (w.e) near the ELA, to 5.9 m (w.e) near the terminus. Melt is greatest along the main tongue, displaying the signature of enhanced sensible heat flux due to persistent katabatic flow (Figure 3.7). The southernmost tributary glacier shows relatively low melt rates relative to its elevation, due predominantly to the fact that it is sheltered from high winds, and its north-facing aspect allows for substantial shading throughout the melt season.

The volume of ice loss during the study period was 0.124 km\(^3\). By the end of the melt season, the ablation area was 27.7 km\(^2\), accounting for 34% of the total glacier surface area (81.3 km\(^2\)).

3.4.2 Relative Energy Contributions

During the study period, net radiation (\( Q^* \)) was the primary driver of ice melt (Figure 3.8). Due to the relatively low measured albedo of ice (values between 0.4 and 0.2), net radiation is strongly controlled by incoming shortwave radiation (\( K \downarrow \)). The relative importance of \( Q^* \) decreases throughout the melt season from roughly 80% of melt energy to less than 60% by September. This decrease is driven primarily by the decrease in sunlight hours later in the season and a decrease in the incidence angle.

Sensible heat flux represents between 20% and 40% of the melt energy supplied to the ice surface during the melt season. Driven primarily by peaks in wind speed and temperature, the relative importance of \( Q_H \) coincides with the warmest days in late July and early August. While a significant decrease in \( Q^* \) is observed during storm events, both sensible and latent heat show only very minor decreases in supplied energy, suggesting that
3.4. Results and Discussion

Figure 3.7: Modelled ice melt (water equivalent) for the study period from June 20 to September 12, 2013.

Inclement weather does not significantly change the supply of turbulent heat to exposed ice on the glacier.

Latent heat flux was a relatively minor source of energy throughout the melt season. The relatively dry air on the lee side of the Coast Mountains made for little condensation on the glacier surface in July and early August. Evaporation was predicted for several days in mid-July, leading to latent heat transfer away from the glacier surface, indicating a low air moisture (roughly 5.98 hPa). Late August rain storms supplied high moisture content in the air, leading to slightly elevated latent heat flux and notable rain events supplying additional energy to the glacier surface.
3.4. Results and Discussion

3.4.3 Modelled Glacier Temperature Distribution

During hot, sunny days (see Figure 3.9), modelled temperature shows a large depression of up to 14 °C on-glacier, relative to off-glacier values. The magnitude of katabatic forcing is substantially higher further down-glacier, while the uppermost reaches of the glacier are relatively insensitive to katabatic forcing, and reflect temperatures that would be expected from a standard environmental lapse rate. The magnitude of this forcing is scaled by off-glacier temperatures ($T_{amb.}$), where warmer ambient temperatures create a stronger katabatic forcing, and a larger temperature depression on-glacier (see Equation 3.14a).

3.4.4 Comparative Model Performance

Although there was insufficient data to fully ground-truth the distributed energy balance modelled (DEBM) melt, the model was tested against sparse data from several sources to constrain the range of uncertainty in predicted melt. Modelled results were compared against a point energy balance model using the same modelling methodology, and derived solely from observations at Glacier AWS (i.e. no FPL distributed $T$, $e$, and $u$). The DEBM
3.4. Results and Discussion

was also evaluated against daily melt measured at Glacier AWS using a pressure transducer in a water-filled borehole. Summer cumulative melt was further constrained by 6 ablation stakes.

A comparison of cumulative melt (Figure 3.10) shows that the Point EBM predicts the highest melt (5.10 m w.e.), while both measured (4.39 m w.e.) and DEBM (4.67 m w.e.) agree relatively well after 67 days (June 20 - August 28). All three methods reproduce a sharp decline in the melt rate for several days in late June. However, measured values show a substantial decrease in the melt rate in mid-July that is not reproduced in either energy balance model. This mid-July decrease in the melt rate corresponds with very large observed diurnal variations in the measured borehole depth (see Appendix Figure A.1), suggesting that the borehole may not have been refilling completely during the period, obscuring the melt rate signal. All three methods also show a mid-August decrease in the melt rate, although the measured value shows significantly more scatter, most likely also due to changes in the water level within the borehole.

The Point EBM produces slightly higher cumulative melt and a consistently higher
3.4. Results and Discussion

Figure 3.10: Cumulative melt at Glacier AWS from the Distributed Energy Balance Model (DEBM), a point Energy Balance Model, and from measured daily melt taken from a pressure transducer in a borehole.

melt rate throughout the summer study period (Figure 3.11). Net radiation is consistently slightly higher in the Point model than the DEBM, due to the fact that the point model does not take into account the surface slope at Glacier AWS, leading to a slightly higher estimate of incoming shortwave radiation.

Evaluation of the sensible and latent heat fluxes from the DEBM and Point EBM gives a sense of how well the FPL reproduces turbulent heat parameters at Glacier AWS. While sensible heat is well reproduced by the DEBM model, latent heat is less well reproduced, showing a consistently lower estimate than from the Point EBM. This suggests that holding relative humidity constant across the glacier, and calculating vapour pressure using FPL distributed temperatures, consistently underestimates the vapour pressure at Glacier AWS. Although further in situ data are necessary to confirm this finding, it is likely that this underestimate is greater near the terminus, where the FPL model predicts the strongest katabatic winds and lowest air temperatures, in turn producing the lowest vapour pressures (Shea and Moore, 2010).

Summer melt measured at the ablation stakes is slightly less than that predicted from the DEBM; however, the discrepancy is less than 0.20 m w.e. (3.5%) for all stakes except
3.4. Results and Discussion

D (Figure 3.12). Although all ablation stakes are assumed to have melted out at the same time, it is possible that there was up to several days between each stake melting out, leading to underestimates of the total melt during the period. Furthermore, surface heterogeneity, such as small meltwater pools forming at the surface, or patches of dirty ice (which would lower the albedo) were observed across the glacier, and could have affected the melt rate at individual stakes, including stake D, which shows substantially less melt than nearby stakes (12%). Overall, total melt at the stakes was well reproduced by the DEBM, suggesting that the model accurately captures the main drivers of melt in this part of the ablation area.
3.4. Results and Discussion

3.4.5 Comparison of FPL Model Uncertainty

In order to test the sensitivity of the distributed energy balance model to the choice of turbulent parameter distribution, the flowpath length model (see Sections 3.3.2 and 3.3.2) was compared with other frequently employed temperature and wind speed distribution models (e.g. MacDougall and Flowers (2011); Hock and Holmgren (2005)). The FPL model is tested against turbulent heat flux calculations obtained by using a spatially constant wind speed (measured at Glacier AWS) distributed across the glacier, and by using a standard temperature lapse rate (see Table 3.2) to distribute measured on-glacier temperatures (from Glacier AWS).

Comparing wind speed distribution models, the FPL model predicts measurably less total melt than using a constant wind speed (Figure 3.13). While the main (southern) channel of Bridge Glacier seems relatively insensitive to the choice of wind speed distribution model, the model predicts over 1.5 m of additional melt for the north-flowing channel for a constant wind speed. This additional ice melt prediction is due to the fact that the FPL model predicts significant sheltering in this side channel, shielding it from the much greater wind speeds experienced near the terminus and Glacier AWS. Although not tested,
it is highly likely that wind speeds are much higher at Glacier AWS than further up-glacier due to a combination of being located in a constrained area with high sidewalls (allowing winds to be funnelled), and close to the terminus, where katabatic winds are expected to be strongest. This FPL model also suggests that, if using a constant wind speed across the glacier, the model will be very sensitive to the location of the wind speed measurements.

Temporally, both sensible and latent heat fluxes increase across the glacier using a consistent wind speed throughout the melt season (Figure 3.14). This is most observable during late July through early August, coinciding with the highest off-glacier air temperatures and highest observed on-glacier wind speeds. Since a constant wind speed model does not take into account any potential for sheltering or spatial variation in wind speeds, it likely overestimates observed wind speeds in sheltered areas.

The FPL model for katabatic temperature distribution predicts more melt overall, compared to a constant temperature lapse rate distribution. Near the terminus there is little change in modelled melt, except for immediately at the terminus, where a constant temperature lapse rate predicts and additional 0.3 m w.e of melt throughout the summer (roughly

Figure 3.13: Plots of the difference in ice melt (in m w.e.) between spatially constant windspeed and temperature lapse rate distributions, and FPL modelled temperature and windspeeds.

3.4. Results and Discussion
Above Glacier AWS, higher melt is predicted using the FPL model, with a predicted difference of up to 1 m (18%). This discrepancy is due to the fact that the FPL model predicts that the magnitude of katabatic forcing generally decreases with increasing elevation, leading to slightly warmer temperatures further up-glacier. Distributing temperatures by using an on-glacier weather station and standard temperature lapse rate will predict much lower temperatures than would be expected from distributing temperatures with off-glacier weather stations, even in areas such as peaks and ridges, where katabatic cooling would not be expected. By using on-glacier temperatures from an AWS in an area strongly affected by katabatic cooling, a standard temperature lapse rate will lead to a significant underestimate of higher elevation temperatures, and subsequently sensible heat fluxes. It also suggests that due to the potential for differential katabatic forcing, the location of an on-glacier weather station will ultimately have a significant effect on modelled temperature distributions using a standard lapse rate.

Sensible heat flux derived from a constant temperature lapse rate is smaller than predicted from a FPL model throughout the melt season (Figure 3.14). This trend is most apparent when katabatic forcing is greatest, due to high ambient temperatures and on-glacier wind speeds. During cool or rainy weather, the difference in sensible heat flux from both models is minor, suggesting that katabatic forcing is negligible during poor weather. This assumption is supported by less frequent observations of downslope winds during these periods.

Since the relative humidity was held constant across the glacier, changes in the latent heat flux are themselves a product of air temperatures. Since the constant temperature lapse rate model predicts very low temperatures at higher elevations during strong katabatic flow conditions, as a by-product, it also predicts very dry conditions, most apparent during the warm, high katabatic flow period of late July through mid-August. Since the constant temperature lapse rate most likely underestimates air temperatures at higher elevations, it is likely that it also underestimates vapour pressure at those elevations. This suggests that using a constant temperature lapse rate coupled with a constant relative humidity across the glacier is sensitive to measurement location, and has the potential to significantly underestimate the latent heat flux, specifically at higher elevations, where katabatic forcing is expected to be less significant.
3.4. Results and Discussion

3.4.6 Climatic Sensitivity

Research has shown the length of time ice is exposed is by far the most critical factor in determining cumulative ice melt during the summer (Moore and Demuth, 2001). In order to further position the 2013 study period within a longer term context, the DEBM was re-run with climatic perturbations to simulate the relative importance of climatic variability. The snowline position throughout the summer is altered, by delaying the retreat by one and two weeks, as well as accelerating it by two weeks. Climatic variability is also tested by increasing ambient temperatures for each timestep by 1°C.

In all cases, the absolute changes in total volumetric ice loss are smaller than the calving flux (see Chapter 5). The absolute change in volume of ice loss due to a 2 week delay snowline retreat; however, is comparable (72.5%) to the calving flux (Figure 3.15). While snowline retreat (and subsequent ice exposure) has long been demonstrated as one of the most important controls on glacier mass balance, the magnitude of this variability is well within the expected annual observed variability in spring snowpack and snowline retreat.

Figure 3.14: Changes in turbulent heat fluxes using different models for spatially distributing temperature and wind speed.
3.4. Results and Discussion

Figure 3.15: Calving loses relative to climatic variability scenarios affecting the melt rate. The DEBM is modified by increasing ambient air temperatures by 1°C, as well as delaying the snowline retreat by 1 and 2 weeks, as well as accelerating it by 2 weeks.

In order to assess the relative sensitivity of Bridge Glacier to future changing climatic conditions, the DEBM also run with increased air temperature (Figure 3.16. Throughout the summer, the ambient temperature (at Ridge AWS) was increased by 1°C for each timestep. Results show that this increase in off-glacier air temperature yields an additional melt of over 0.25 m w.e. in the lower bench of the glacier, and an additional $5.73 \times 10^{-3}$ km$^3$ of ice loss.

Higher ambient temperatures increase melt directly due to higher on-glacier temperatures, while it also indirectly enhances melt by intensifying katabatic flow. This feedback is borne out in the model by enhanced melt along the largest flow path lengths along the main southern tongue of the glacier, while sheltered areas show a minimal increase in melt. This feedback mechanism also suggests that ambient temperatures have a stronger effect on melt through their ability to enhance wind speeds rather than air temperatures themselves.

While an increase in ambient air temperature does produce a meaningful increase in ice melt, the impact is comparatively modest when taking into account the magnitude of changes in other melt variables. This increase in ambient air temperature only results in
3.4. Results and Discussion

Figure 3.16: Predicted additional melt from a 1°C increase in air temperature during the 2013 melt season.

a 5% increase in cumulative melt. While this increase is significant, it could also be easily achieved by a small increase in incoming radiation, due to fewer cloudy days, or a few additional melt days.

Additionally, since it has been shown that sensible heat flux has a larger impact on snow rather than ice (Munro, 2006), an increase in air temperatures would accelerate the snowline retreat and provide additional ice melt on top of what has been estimated here. Ultimately, however, this test shows that the Bridge Glacier’s climatic sensitivity is relatively large, and small changes in summer air temperatures, winter precipitation, or decreased cloudiness can all provide significant additional ice loss.
3.5 Conclusions

Between June 20 and September 12, 2013, up to 5.9 m w.e. of ice was lost in the ablation area, leading to a total melt volume of 0.124 km$^3$. The melt rate was highest in July through mid-August, corresponding with warmest temperatures, and predominantly clear-sky conditions. Modelled melt during the study period corresponds well with measured melt from both ablation stakes, and cumulative daily melt from a pressure transducer in a water-filled borehole at Glacier AWS.

Climatic sensitivity analyses suggest that given the current geometry of Bridge Glacier, the volumetric ice loss variability is in the range of ±0.03 km$^3$. Ice loss is most strongly controlled by the length and timing of ice exposure, while it is only moderately sensitive to changes in mean air temperature.

During the study period, net radiation, driven primarily by incoming shortwave radiation, is the main source of melt energy across the ablation area. Net radiation accounts for almost 80% of the ice melt in June and July, becoming less important throughout the summer as the number of sunlight hours decreases. Net radiation is very variable, showing large decreases during summer storms, a trend that is not mirrored in latent and sensible heat fluxes.

Contrasting with many other DEBM temperature and wind speed distribution models (MacDougall and Flowers, 2011; Hock and Holmgren, 2005; Anslow et al., 2008), this formulation considers observed katabatic forcing (Petersen and Pellicciotti, 2011; Jiskoot and Mueller, 2012). Katabatic flow exerts a significant decrease in on-glacier air temper-
3.5. Conclusions

ature and an increase in on-glacier wind speed, while both trends are strongly enhanced by high off-glacier air temperatures. Using flow path lengths, the magnitude of katabatic forcing is modelled, allowing for an account of katabatic influence on air temperature. The model produces results that allow for significant cooling in the lowest on-glacier reaches, while producing high elevation results that are in line with what would be expected from off-glacier standard temperature lapse rates.
Chapter 4

Measuring Glacier Velocity with Time Lapse Imagery

4.1 Introduction

Measuring glacier velocity has long been an important part of glacier study, and is an integral variable in calculating calving fluxes in tidewater and lacustrine systems (Meier and Post, 1987), strain rates and estimating basal water pressures. Original measurements of glacier velocity involved repeat manual surveys of on-glacier targets. While this technique benefits from a simple design and low instrumentation requirements, it is less desirable from a glaciological perspective because it provides a lower time resolution since the temporal resolution depends on the frequency of visits to the glacier.

With the increased reliability in Differential Global Positioning Systems (DGPS), providing accuracy of up to 10 cm, the temporal resolution of a velocity dataset can be substantially augmented by installation of a DGPS on-glacier (Danielson and Sharp, 2013). While this technique can yield high accuracy measurements, translating into hourly to sub-daily resolutions, it only provides displacements for one location, and is limited by the high costs of instrumentation and maintenance.

The use of remotely sensed images, such as Landsat or SPOT, to track glacier flow overcomes the spatial constraints of point measurements. However, it too suffers from low temporal resolution, dictated by the frequency of repeat images. In environments where heavy clouds and consistent precipitation patterns are expected, the number of usable images is further reduced due to cloud-cover obscuring the glacier surface.

As a relatively low cost alternative, time lapse photography offers a remote option that can provide sub-daily velocity measurements over an array of targets distributed across the field of view of the camera. The use of time lapse photography to calculate glacier velocity has been in practice for several decades, and has yielded valuable data, albeit
with significant challenges, such as camera displacement, inclement weather, and poor battery performance (Krimmel and Vaughn, 1987; Harrison et al., 1992). However, recent technological advances increasing image quality and digital storage, as well as the design of automatic camera triggers and robust weather enclosures, have made repeat photography a convenient and reliable method of glaciological observation.

Monoscopic (single) cameras are the most basic time lapse set-up that can yield displacement values, although without additional surveyed points, displacement values are only relative to the camera. To overcome this limitation, ground control points must be surveyed, or estimated, in order to calculate true displacements. Another method that has been implemented to avoid this complication involves using a DEM to orthorectify each image (Corripio, 2004), yielding a distributed velocity time series from a monoscopic time lapse dataset (Rivera et al., 2012). This method is reliant on having an accurate DEM, which is unfeasible in many instances due to substantial glacial thinning. Since errors in glacier surface topography will propagate into orthorectification, and subsequently, measured displacements, values derived using a DEM are, in some cases, no more accurate than assuming a planar glacier surface.

To overcome this limitation, stereoscopic (two overlapping images) photography presents an alternative that allows for the derivation of real world coordinates directly. This method is preferable for some applications (Sund et al., 2011; Rivera et al., 2012) and has the added benefit of estimating glacier thinning rates. However, this method requires additional geometric and geographic inputs, a full account of lens distortion to accurately produce direct displacements, as well the increased cost of deploying two systems.

More recent advances have employed computer vision algorithms instead of manual image-pixel tracking (Ahn and Howat, 2011). Automated algorithms can provide greater spatial coverage, at a fraction of the time required for manual pixel tracking, and in some cases can provide more accurate solutions (Eiken and Sund, 2012). However, the dynamic nature of image lighting, changing glacial features throughout the season, and inclement weather end up reducing their effectiveness and reliability unless substantial image pre-processing is applied (Ahn and Howat, 2011).

The purpose of this chapter is to present a method for producing glacier velocity measurements from a monoscopic time lapse camera using manual feature tracking, outlining a low-cost method with minimal external input requirements. Secondly, this chapter will present results from two monoscopic cameras and quantify the errors in this method. Finally, this chapter aims to produce daily velocities at two locations on Bridge Glacier that
can be used to quantify the calving flux and strain rates (see Chapter 5).

4.2 Study Site and Instrumental Setup

Three cameras were installed at Bridge Glacier between June 17 and September 14, 2013. Two cameras were installed roughly perpendicular to glacier flow, while a third was located with a full view of the terminus, roughly 2 km from the terminus, and was not used to measure glacier velocities. One camera (referred to as ‘Terminus Camera’) was installed perpendicular to the ice front along Bridge Lake, and captured the floating terminus of the glacier. A second camera (referred to as ‘Nunatak Camera’) was installed perpendicular to flow on top of a prominent (former) nunatak, with a view along the south (main) branch of the glacier (see Figure 4.1 and 4.3). Both cameras were programmed to shoot every 30 minutes between 6 am and 9 pm, producing 34 images per day.

Figure 4.1: Bridge Glacier extent at the end of the 2013 study period, and the location of the two tracking time lapse cameras, Terminus Camera and Nunatak Camera. Contour intervals are 40 m asl.
4.2. Study Site and Instrumental Setup

The time lapse camera setup was designed ready-made by Harbortronics. A Canon EOS Rebel T3i camera with a 10-20 mm wide-angle lens was installed in a custom-made enclosure powered by two lithium batteries, which are recharged by a solar panel. The camera is controlled using custom command line software designed to program a DigiSnap 2000 intervalometer.

(a) Nunatak Camera, September 12, 2013. (b) Velocity Stake survey, June 20, 2013.

Figure 4.2: Photos of time lapse camera and velocity stake

Cameras were installed on June 17, 2013. On July 18, 2013, the cameras were checked, and the SD card was changed. On September 14, 2013, the cameras were uninstalled and removed from the field. During the June and July visits, the location and elevation of each camera was checked using a handheld Garmin GPS. Other geographic parameters (Table 4.1) were derived from BC TRIM and Google Earth SPOT data.

Each camera produced over 3000 images during the study period. Time lapse images were culled before feature tracking was applied. Daily images from Nunatak Camera at 18:00 hrs were selected. Terminus Camera images were selected bi-daily, at 06:00 and 21:00 hrs. Images from both cameras were manually checked, and several images were replaced by successive images in the time series due to fog, poor lighting, water droplets, or even in one case, a marmot peering into the inclosure. Images were selected to minimize
4.2. Study Site and Instrumental Setup

(a) Nunatak Camera view. The blue line represents 595 m at the centreline points (1030 m from the camera).

(b) Terminus Camera view. The blue line represents a scale of 690 m at the ice front (970 m from the camera).

Figure 4.3: Camera views and tracked points from images taken on June 17, 2013. Points in the foreground correspond to ground control points.
4.3 Measuring Glacier Velocity from Time Lapse Images

Table 4.1: Time lapse camera variables.

<table>
<thead>
<tr>
<th></th>
<th>Terminus Camera</th>
<th>Nunatak Camera</th>
</tr>
</thead>
<tbody>
<tr>
<td>Glacier Slope</td>
<td>0.060</td>
<td>0.075</td>
</tr>
<tr>
<td>Flow Direction (δ)</td>
<td>10.6°</td>
<td>13.0°</td>
</tr>
<tr>
<td>Focal Length</td>
<td>15 mm</td>
<td>16 mm</td>
</tr>
<tr>
<td>Horizontal Angle of View</td>
<td>100.4°</td>
<td>96.7°</td>
</tr>
<tr>
<td>Height Above Glacier (h)</td>
<td>65 m</td>
<td>130 m</td>
</tr>
</tbody>
</table>

the effects of changing solar angles, and all replaced images were within 2 hours of the ‘ideal’ time. Typical camera views are shown in Figure 4.3, along with targets tracked throughout the study period.

Six stakes were also installed on the glacier, two of which were within the viewshed of the Nunatak Camera. These stakes were surveyed at visits in June, July and September with a Garmin eTrex Legend Cx commercial grade GPS. The GPS is WAAS enabled and has been tested to be accurate up to 1 m.

4.3 Measuring Glacier Velocity from Time Lapse Images

The overall procedure for measuring glacier velocity from oblique time lapse images is outlined in Figure 4.4.

4.3.1 Tracking Features

Features were tracked manually using Tracker (http://www.cabrillo.edu/~dbrown/tracker/), an Open Source program that produces X and Y pixel values for a tracked point in successive frames. While Tracker has an automatic tracking feature, it was found to be unreliable due to the relatively large changes in target lighting and shape throughout the season. Changing shadows (due to time of day, weather, seasonality), changing feature shape (due to melt or movement), and disappearance of the target altogether (calving or melt) made tracking a single feature through the whole time series difficult, and sometimes impossible. In the case that a feature was lost, a nearby feature was selected for the remainder of the dataset (the time-step of the change was flagged so as not to produce an artificial displacement).

In order to pick the best possible feature to track, a few simple guidelines were followed:

- Tracked features were selected to have a substantially different colour than the back-
4.3. Measuring Glacier Velocity from Time Lapse Images

Figure 4.4: Time lapse image processing workflow

- ground ice. Rocks or crevasses worked best since the black features stuck out against the grey/white glacial ice in a wide range of lighting conditions.
- The middle of a feature was tracked. Shadows tended to distort the edge of features, making it harder to discern shadow from feature edge.
- Features were selected near the centre of the image to minimize the potential for image distortion, which was not accounted for in this study.
- Late afternoon images had the least amount of light angle changes throughout the season, while early morning and evening pictures went from direct to diffuse sunlight as the days got shorter, while lengthening shadows also made features more difficult to discern. The lighting for midday images was found to be most affected by changes
in cloud cover.

In order to account for camera motion due to winds, temperature changes, and other factors shifting the camera orientation (Ahn and Howat, 2011; Eiken and Sund, 2012; Ahn and Box, 2010), several ground control points were tracked. Ground control points were selected as distinguishable features that would not be expected to move, such as large rocks or lichen on a rock in the foreground. The pixel displacement for each ground control point was checked manually, and a mean background pixel displacement was computed. This baseline value was then subtracted from each tracked feature pixel displacement.

### 4.3.2 Converting pixel displacement into distance displacement

Converting a pixel displacement into distance displacement was achieved by creating a pixel to distance scale (m/pixel). By knowing the location of two control points (CP in Figure 4.5), the distance between them perpendicular to the camera \(x\) can be computed (Harrison et al., 1992) or measured as follows:

\[ x' = x \frac{y'}{y}. \]  

(4.1)

In this study, initial values of \(x\) and \(y\) were measured using Google Earth. Using the distance from the line \(x\) to the camera \((y)\), and from the tracked feature point to the camera \((y')\), a scale can be created where \(x'\) is proportional to \(x\), which can be related to the number of pixels in that distance to calculate a distance per pixel translation factor.

Assuming that the glacier flow is perpendicular to the camera (parallel to and at line \(x'\)), we can reference points (in real world distance) with respect to \(x'\), where 0 corresponds with the centre of the frame (the intersection of lines \(x'\) and \(y'\)). In this convention, positive displacement is attributed to movement from left to right, and was corrected since glacier flow was in the opposite direction.

### 4.3.3 Calculating the horizontal distance from camera

In order to accurately calculate the down-glacier displacement in distance units (Equation 4.1), an adequate distance per pixel ratio must be calculated. This calculation relies on an accurate measurement of the distance between the camera and the tracked point \((y_i)\). Estimations of this distance can be aided and constrained by glacier features that can be seen from georeferenced remotely sensed data (such as medial moraines, or large crevasses).
4.3. Measuring Glacier Velocity from Time Lapse Images

Figure 4.5: Camera geometry for converting pixel displacement into pixel values.

However, errors in the distance between the camera and tracked feature will scale linearly with measured displacement (Eiken and Sund, 2012), making an accurate calculation of \( y_i \) more important.

By assuming a planar glacier surface, geometric relationships can be used to calculate the distance of any point on the glacier along the line \( y \) (Figure 4.6). By knowing the elevation of the camera above the glacier plane (\( h \)), and the angle of the point of interest from horizontal (\( \alpha \)), referred to as the zenith angle, a target’s distance from the camera (\( y_i \)) can be calculated as follows:

\[
y_i = \frac{h}{\tan \alpha_i}.
\]  
(4.2)

In order to solve Equation 4.2, the zenith angle (\( \alpha \)) must be calculated for the point of interest. This can be accomplished by having two known points, and their distances from the camera (\( y_1, y_2 \)), as well as their corresponding Y pixel values (\( Y_1, Y_2 \)). The zenith angle (\( \alpha \)) is first calculated for each known point as follows:

\[
\alpha = \tan^{-1} \frac{h}{y}.
\]  
(4.3a)
4.3. Measuring Glacier Velocity from Time Lapse Images

Figure 4.6: Camera depth geometry, where \( y_0 \) is horizontal to the camera.

Since the zenith angle scales linearly with Y pixel values (\( Y \)), the zenith angle for any Y pixel location (\( Y_i \)) can be calculated as

\[
m = \frac{Y_2 - Y_1}{\alpha_2 - \alpha_1}
\]

where \( m \) is the slope of the regression line. Individual pixel points can be converted into zenith angles as

\[
\alpha_i = \frac{Y_i - b}{m}
\]

using the intercept of the regression line, \( b \).

Finally, zenith angles for tracked points can be converted into distances from the camera using Equation 4.2. This relationship suggests that the degree of horizontal foreshortening is strongly dependent on the height of the camera above the glacier plane (\( h \)). For relatively low heights, the degree of foreshortening is great, and a small decrease in the zenith angle will result in a large increase in distance from the camera. Since picture quality is confined to a finite number of pixels, the accuracy of any pixel measurement in the \( y \) direction decreases substantially with low camera heights and low zenith angles (Figure 4.7).
4.3. Measuring Glacier Velocity from Time Lapse Images

4.3.4 Correcting from non-perpendicular flow

In the likely case that flow is not perfectly perpendicular to the camera (represented as $v$ in Figure 4.5), knowledge of the target’s X pixel location and direction of flow yields a corrected displacement value ($d$). The raw displacement value ($d'$) can be corrected, following Eiken and Sund (2012), where

$$d = \frac{d'}{\cos \delta + \sin \delta \tan \theta}$$ \hspace{1cm} (4.4)

and $\theta$ is the angle of the target relative to the centre of the camera frame and $\delta$ the direction of glacier flow relative to perpendicular to the camera.

4.3.5 Data Processing

Displacement values were filtered, where negative (up-glacier) displacements and values greater than a specified threshold (1.5 $\text{md}^{-1}$ in this study) were classified as erroneous. Velocities were computed by dividing the displacement by the individual time interval.
between images. Finally, if multiple points were tracked in the same approximate location, measurements for each time step were compared. A final velocity was calculated by using a trimmed mean. In each case, if the number of velocity measurements for a time-step was greater than half the total possible measurements, and the spread was larger than 3 standard deviations, the value furthest from the mean was removed. This process was repeated a second time, and second outlier was removed if the spread was still larger than 3 standard deviations.

Outliers were predominantly substantially larger than the mean tracked velocity. These outliers are most likely due to the fact that if a feature is not captured in the previous frame, the displacement captures 2 (or more) days, artificially overestimating the daily velocity. The majority of these values were removed by applying a threshold (1.5 md$^{-1}$ is roughly 3 times the average flow speed). However, some outliers remained, and were most often visibly divergent from the majority of tracked points.

4.4 Results and Discussion

4.4.1 Ground Control Points

![Noise from Stationary Nunatak Control Points](image1)

![Noise from Stationary Terminus Control Points](image2)

(a) Nunatak Camera GCP noise  
(b) Terminus Camera GCP noise

Figure 4.8: Displacements for stationary Ground Control Points for the two cameras. Note: the anomalously large displacement in late July is due to the camera being moved when the SD cards were changed.

The displacements for the ground control points show significantly larger displacements in the nunatak dataset (Figure 4.8). The average displacement in the terminus data was in the order of ±1-2 pixels, while the nunatak data showed an average displacement of ±4-5
4.4. Results and Discussion

pixels. The nunatak camera was substantially more exposed, and had a larger wind drag due to a wood backing on the tripod, making it more susceptible to shaking. The terminus camera, on the other hand, was sheltered by a large boulder and was located in a small depression.

Ground control point displacements on Terminus Camera were found to have a diurnal pattern, with movements downglacier during the afternoon, and upglacier in the morning. This pattern was most strongly observed during periods of high pressure, sunny weather, coinciding with the ideal conditions that promote strong katabatic winds, suggesting that wind was the main factor responsible for camera stability.

4.4.2 Terminus Velocity

Tracked points from Terminus Camera showed relatively consistent velocities of roughly 0.4 m·d\(^{-1}\) throughout the study period, corresponding with total displacements of just under 34 m. Between 6 and 14 frames were discarded from each target due to anomalously high, or negative daily displacement values (see Table 4.2).

Table 4.2: Terminus tracked points. All points are located along the terminus ice front, roughly perpendicular to glacier flow, 972 m from the camera.

<table>
<thead>
<tr>
<th>Point</th>
<th>Discarded Frames</th>
<th>Displacement (m)</th>
<th>Flow Speed (ma(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>13</td>
<td>34.28</td>
<td>140.60</td>
</tr>
<tr>
<td>b</td>
<td>14</td>
<td>31.97</td>
<td>131.13</td>
</tr>
<tr>
<td>c</td>
<td>13</td>
<td>33.56</td>
<td>137.62</td>
</tr>
<tr>
<td>d</td>
<td>6</td>
<td>33.70</td>
<td>138.20</td>
</tr>
<tr>
<td>e</td>
<td>14</td>
<td>32.52</td>
<td>133.38</td>
</tr>
<tr>
<td>f</td>
<td>12</td>
<td>34.30</td>
<td>140.65</td>
</tr>
<tr>
<td>g</td>
<td>9</td>
<td>36.40</td>
<td>149.27</td>
</tr>
<tr>
<td>h</td>
<td>9</td>
<td>33.91</td>
<td>139.08</td>
</tr>
<tr>
<td>Mean</td>
<td>11</td>
<td>33.83</td>
<td>138.74</td>
</tr>
</tbody>
</table>

Raw velocity data were very noisy, and were smoothed using a running mean (span of 5) in order to attempt to isolate seasonal variations. Average velocity measurements show little variability throughout the summer (Figure 4.9). A small increase in velocity is apparent throughout July and mid-August. In early September, terminus velocity decreases roughly 0.1 m·d\(^{-1}\).
4.4 Results and Discussion

This relatively steady summer velocity profile is consistent with what would be physically expected from a floating terminus. Since there is no basal friction, seasonal changes in meltwater channels, and subsequently basal water pressure, do not affect the terminus. Absent changes in basal friction (or in this case lateral friction as well, since the terminus does not abut valley side-walls), we would expect ‘plug flow’, where ice velocity is determined primarily by the channel gradient, which did not measurably change throughout the study period.

4.4.3 Nunatak Velocity

Tracked points along the glacier centreline from Nunatak Camera showed very similar average velocities during the study period, with substantially higher peaks, and more variability. Similar to the terminus dataset, raw daily velocity measurements were noisy, and had to be smoothed using a running mean to isolate weekly and seasonal trends (Figure 4.10). A peak is located at the beginning of the dataset (June 18-20) and then a short period of low velocities at the end of June. July 1-12 is marked by well above average velocities, reaching running 5 day averages of over 0.85 md$^{-1}$. From July 20 onwards,
the velocity time series settled into a variable pattern of relatively low values, oscillating between 0.3 and 0.5 \( \text{m day}^{-1} \). Between 6 and 15 frames were discarded from each tracked point due to anomalously high, or negative displacement values (see Table 4.3).

![Nunatak centerline velocity time series](image)

Figure 4.10: Nunatak centerline velocity time series. The grey lines represent each tracked target, while the bold line is the trimmed mean. Data are smoothed into a running mean (5 time steps).

Tracked points along the glacier centreline from Nunatak Camera showed very similar average velocities during the study period, with substantially higher peaks, and more variability. Similar to the terminus dataset, raw daily velocity measurements were noisy, and had to be smoothed, using a running mean to isolate weekly and seasonal trends (Figure 4.10). A peak is located at the beginning of the dataset (June 18-20) and then a short period of low velocities at the end of June. July 1-12 is marked by well above average velocities, reaching running 5 day averages of over 0.85 \( \text{m day}^{-1} \). From July 20 onwards, the velocity time series settled into a variable pattern of relatively low values, oscillating between 0.3 and 0.5 \( \text{m day}^{-1} \). Between 6 and 15 frames were discarded from each tracked point due to anomalously high, or negative displacement values (Table 4.3).

The increase in velocity in late June is consistent with classical glacier flow theory where an increase in basal water pressure builds up early in the melt season. High basal water pressure reduces friction at the bed, allowing for faster flow. As the summer progresses, and more melt water is stored in the glacier, sub-glacial channels are scoured,
4.4. Results and Discussion

Table 4.3: Nunatak centreline velocity time series. All points are from near the centerline, 897 m to 980 m from the camera, roughly perpendicular to glacier flow.

<table>
<thead>
<tr>
<th>Point</th>
<th>Target Distance (m)</th>
<th>Discarded Frames</th>
<th>Displacement (m)</th>
<th>Flow Speed (ma$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>897</td>
<td>10</td>
<td>33.61</td>
<td>137.85</td>
</tr>
<tr>
<td>h</td>
<td>911</td>
<td>6</td>
<td>34.04</td>
<td>139.60</td>
</tr>
<tr>
<td>d</td>
<td>920</td>
<td>12</td>
<td>32.67</td>
<td>133.98</td>
</tr>
<tr>
<td>g</td>
<td>946</td>
<td>14</td>
<td>37.81</td>
<td>155.06</td>
</tr>
<tr>
<td>b</td>
<td>951</td>
<td>13</td>
<td>32.66</td>
<td>132.31</td>
</tr>
<tr>
<td>c</td>
<td>958</td>
<td>12</td>
<td>33.84</td>
<td>138.79</td>
</tr>
<tr>
<td>f</td>
<td>976</td>
<td>13</td>
<td>30.97</td>
<td>127.01</td>
</tr>
<tr>
<td>e</td>
<td>980</td>
<td>7</td>
<td>33.74</td>
<td>138.36</td>
</tr>
<tr>
<td>Mean</td>
<td>942</td>
<td>11</td>
<td>33.62</td>
<td>137.87</td>
</tr>
</tbody>
</table>

 eventually allowing the stored basal water to drain, decreasing the basal water pressure, and subsequently, flow speeds.

Since Nunatak Camera was located high enough above the glacier surface, it was possible to track points distributed across the width of the glacier. Tracked points show very low velocities very close to the north edge of the glacier (Table 4.4). Velocities increased with distance from the camera (southward across the glacier), except for the two furthest points (b and l), which were both within 150 m of the south margin of the glacier, which showed slightly lower flow speeds than those close to the centreline.

Table 4.4: Nunatak Camera distributed points. All points are within 200 m of the centre of the image. The glacier extends from 337 m to 1698 m from Nunatak Camera.

<table>
<thead>
<tr>
<th>Point</th>
<th>Target Distance (m)</th>
<th>Discarded Frames</th>
<th>Displacement (m)</th>
<th>Flow Speed (ma$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>c</td>
<td>422</td>
<td>22</td>
<td>13.00</td>
<td>53.30</td>
</tr>
<tr>
<td>a</td>
<td>494</td>
<td>21</td>
<td>13.02</td>
<td>53.38</td>
</tr>
<tr>
<td>e</td>
<td>576</td>
<td>10</td>
<td>21.88</td>
<td>89.75</td>
</tr>
<tr>
<td>g</td>
<td>722</td>
<td>6</td>
<td>27.07</td>
<td>111.00</td>
</tr>
<tr>
<td>f</td>
<td>1080</td>
<td>12</td>
<td>32.56</td>
<td>133.52</td>
</tr>
<tr>
<td>h</td>
<td>1169</td>
<td>18</td>
<td>37.74</td>
<td>154.79</td>
</tr>
<tr>
<td>k</td>
<td>1262</td>
<td>12</td>
<td>39.02</td>
<td>160.03</td>
</tr>
<tr>
<td>d</td>
<td>1425</td>
<td>14</td>
<td>39.49</td>
<td>161.94</td>
</tr>
<tr>
<td>b</td>
<td>1568</td>
<td>11</td>
<td>37.20</td>
<td>152.56</td>
</tr>
<tr>
<td>l</td>
<td>1642</td>
<td>12</td>
<td>37.99</td>
<td>155.81</td>
</tr>
</tbody>
</table>
4.4. Results and Discussion

Since Nunatak Camera was located high enough above the glacier surface, it was possible to track points distributed across the width of the glacier. Tracked points show very low velocities very close to the north edge of the glacier (Table 4.4). Velocities increased with distance from the camera (southward across the glacier), except for the two furthest points (b and l), which were both within 150 m of the south margin of the glacier, which showed slightly lower flow speeds than those close to the centreline. The velocity time series for all tracked points showed a very similar pattern to the centreline points (Figure 4.11). All points show a decrease in flow velocity at the end of June, followed by a large peak in early July. Throughout the rest of the study period, velocity changes were minimal, with some points further from the camera (closer to the centreline and south margin) showing small increases in flow speed in late August. More generally, flow speeds were greater further southward, and faster points showed larger amplitude changes throughout the study period, while low velocity points were less variable.

The observed lateral velocity pattern (Figure 4.12, 4.13) is consistent with glacial flow models, where lateral drag reduces flow speeds close to the margins of the glacier, and the highest flows are found along the thalweg. In this case, the highest flow speeds are found further towards the south margin, most likely due to the fact that the glacier is
4.5. Error Analysis

coming out of a southeasterly bend, although it is also possible that this is a reflection of the longitudinal stress coupling between the floating terminus, and grounded up-valley ice. Extremely low flow speeds observed at the north (close) margin reflect the shallow, almost stagnant state of the ice, and the eddy created by the nunatak and flow bend.

Figure 4.12: Tracked points from the Nunatak Camera (40 m contour intervals).

4.5 Error Analysis

Although the measurements presented for Terminus and Nunatak camera conform relatively well with expected results, there are several sources of error that affect both the total and relative value of measured feature displacement, and are discussed below.
4.5. Error Analysis

Figure 4.13: Lateral profile from Nunatak Camera. In reality, points can be up to 200 m from centre of the frame in the \((x)\) direction, but have been lined up to estimate the lateral velocity distribution. Glacier margins are represented by the solid black lines at 337 m and 1698 m.

4.5.1 Human Measurement Error and Camera Shift

Manual tracking has benefits over automated matching solutions in that it can produce results when a statistical match is unlikely, benefitting from a human ability to discern changing feature shapes and shadows in each frame. While this suggests a superior method, this rational selection is at least partially offset by a loss in precision. Although Tracker software enables very high (sub-pixel) precision in manual measurements, in practice those measurements will be limited by an ability to select the ideal location. This selection error will be compounded by the fact that, given the finite nature of pixels, sub-pixel precision will be limited by having to manually interpolate within the pixel to select the centroid of the tracked feature.

The standard deviation was calculated for each time step from each camera, comparing the relative displacement (in pixels) of measured stationary Ground Control Points (Figure 4.14). The mean standard deviation for Nunatak Camera was 0.52 pixels/day, while Terminus Camera was 0.41 pixels/day. These values suggest that human selection error is roughly half a pixel per displacement step, and represents the relative uncertainty in correcting raw displacement values for camera shift. This uncertainty can be reduced
somewhat by filtering data using a threshold, and removing negative displacements in order to mitigate erroneous corrections.

Figure 4.14: Standard deviations for 3 tracked ground control points from each time lapse camera.

4.5.2 Photographic and Feature Quality

By assuming that the 8 points tracked within 100 m from Nunatak Camera and within 200 m on the floating margin of Terminus Camera have the same displacement, comparing their relative pixel displacements yields an estimate of the uncertainty due to human errors as well as error induced by photographic quality and the changing shape and orientation of tracked features. The quality of time lapse data is strongly dependent on the quality of images and camera used, both in absolute pixel density, and in sensor quality. The cameras used in this study are 12.1 megapixels, and are moderate quality and relatively inexpensive. Due to limited image storage capacity, there is a tradeoff between image quality and shooting frequency that must be balanced. In this study, standard deviations for both cameras are between 0.44 pixels/day (Terminus) and 0.46 pixels/day (Nunatak), suggesting this quality of image produces errors of roughly 0.12 m/d^{-1} at 1 km from the camera (Figure 4.15).

4.5.3 The Geometry of Control Points

Although the relative uncertainty of individual velocity time steps is relatively minor, over the season, error is introduced when attempting to derive absolute distances from time lapse data. In order to produce accurate absolute results, the distance to pixel ratio must be computed, which is itself dependent on the geometry of the tracked point relative to the
4.5. Error Analysis

Figure 4.15: Standard deviations for 8 tracked points from each time lapse camera. Points removed from the outlier mean calculated average are also omitted from this calculation. Distance estimates are derived from the mean distances of centreline tracked points.

camera. While the relationships for computing this geometry are elementary, substantial difficulties remain in selecting image control points and accurately surveying them.

In order to derive an initial distance to pixel ratio, two control points must be surveyed, and the distance between them calculated perpendicular to the camera (see Figure 4.5). In this study, Google Earth was used to calculate the distance between points, and the accuracy of the measurements are estimated to be ±5 m. This estimate is compounded by the accuracy of the surveyed location of the camera, which is given as 1 m. Uncertainty in surveying the elevation of the camera is assumed at ±10 m, producing an additional error of 0.7%. The total estimate from daily displacement measurements translates into an error of 2.7% on absolute displacement.

Another geometric uncertainty is in calculating the distance of a tracked feature from the camera. The first component in calculating this value is deriving accurate distances along the centre of the frame for two known control points. Since features used had to be along the glacier plane, both the close and far edge of the glacier were selected. The uncertainty involved in delineating the margin of the glacier and accurately surveying its position is estimated at ±10 m, translating to an error of 6% of the final distance to pixel translation factor. This error is likely a slight underestimate in the total error in this procedure, due to the fact that it is difficult to quantify the effect of a non-planar glacier surface, which will introduce further uncertainty in the calculated distance of a feature from the camera. The total error associated with geometric simplifications and surveying is estimated at 9% of the absolute measured displacement.
4.5.4 Comparison with Surveyed Results

Comparison between surveyed stake displacement, and time lapse camera measured displacement over the entire study period suggests that the time lapse camera method produces greater displacement values (Table 4.5). A comparison of surveyed stake ‘B’, which is within 100 m of Nunatak Camera centreline measurements, suggests absolute displacements are within the estimated error. Comparison between surveyed stake ‘A’ and Nunatak distributed point ‘h’, however, shows a significantly higher camera-derived displacement.

Table 4.5: Comparison of GPS surveyed stakes and time lapse camera derived study period displacements.

<table>
<thead>
<tr>
<th></th>
<th>Distance From Camera (m)</th>
<th>Displacement (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Nunatak Points</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Nunatak Averaged Centreline</td>
<td>942</td>
<td>33.6 ±3.1</td>
</tr>
<tr>
<td>Nunatak Distributed point ‘h’</td>
<td>1169</td>
<td>37.7 ±3.5</td>
</tr>
<tr>
<td><strong>Surveyed Points</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>830</td>
<td>27 ±5</td>
</tr>
<tr>
<td>A</td>
<td>1222</td>
<td>23 ±2</td>
</tr>
</tbody>
</table>

In this study, the GPS values are relatively unreliable for a full evaluation of the time lapse camera method due to the high errors involved. The quality of the GPS surveying was substantially reduced by high valley side walls, short occupation times, and the relative logistical challenges in bringing higher accuracy (D)GPS systems into the field. The measured stakes suggest; however, that tracked displacement values agreed with surveyed values within a relatively narrow range. Differences within a few metres suggest that the discrepancies may just as likely be due to the uncertainties associated with the GPS system as values measured from time lapse imagery manual feature tracking.

4.6 Conclusions

In this chapter, a method has been presented for producing usable glacier velocity measurements from a monoscopic time lapse camera using manual feature tracking. The method has proven reliable at daily timescales, although it has potential to provide sub-daily displacement measurements under ideal conditions.

The estimated error for this method is 0.45 pixels per displacement measurement due to human error, and 9% of the measured absolute displacement due to uncertainty in
4.6. Conclusions

the measurement of control points. This error, already acceptable for many glaciological applications, could be substantially reduced by using high quality GPS equipment to survey ground control points and the use of high resolution, up-to-date remotely sensed imagery to estimate glacier margins. This error could also be substantially reduced with the use of a high quality, up-to-date DEM, although this method is not discussed here (see Corripio (2004)).

Between June 17 and September 14, 2013, Bridge Glacier flowed at a speed of 139 ma$^{-1}$ in the terminus region, and up to 162 ma$^{-1}$ in the fastest section of the nunatak region. Averaged centreline speeds were 138 ma$^{-1}$. Flow speeds were relatively constant along the floating terminus, however, flow increased dramatically in early July all along the ablation area, before settling into significantly lower flow speeds from late July through the remainder of the study period.
Chapter 5

The Magnitude and Timing of Calving

5.1 Introduction

For glaciers that terminate in lacustrine or marine waters, calving can be an important component of mass balance, as it leads to greater ice loss than what would be possible through surface ablation alone (Van der Veen, 2002). The calving rate, $U_c$ (ma$^{-1}$), can be expressed as

$$U_c = U_T - \frac{dL}{dt} - U_{wm}$$  \hspace{1cm} (5.1)

and is a function of the average surface velocity at the terminus, $U_T$ (ma$^{-1}$), the change in glacier length with time, $\frac{dL}{dt}$ (ma$^{-1}$), and the terminus melt below the waterline, $U_{wm}$ (ma$^{-1}$).

Field measurements (or estimates) of the surface velocity near the terminus, and the change in terminus position, can be used to calculate the calving rate (Benn et al., 2007b). For continental lacustrine systems, where lakes freeze in the winter, it is assumed that calving events only occur in the summer months, while in low elevation or marine systems calving may be a source of ice loss throughout the year (Warren and Aniya, 1999; Skvarca et al., 2002; Nick et al., 2010; Van der Veen, 2002).

Multiple studies have emphasized the importance of water depth to the rate and magnitude of calving (Skvarca et al., 2002; Warren and Kirkbride, 2003; Post et al., 2011). Significant positive linear relationships have been found between water depth and calving rates (Pelto and Warren, 1991; Haresign, 2004). However, this relationship does not fully explain the complexity of calving responses in different environments, or the seasonal variation in calving rates (Van der Veen, 2002).

Water depth remains an important variable in determining calving rates due to the potential for deep water to promote terminus buoyancy. Terminus buoyancy is found to
5.1. Introduction

have, in many cases, coincided with disintegration and rapid retreat of the ice terminus (Boyce et al., 2007). Other calving systems have been found to withstand terminus flotation for several seasons, although they have become more sensitive to small perturbations in terminus conditions, and more likely to experience large calving losses (Benn et al., 2007b; Post et al., 2011).

Water depth is recognized as an important control on the multi-annual calving flux; however, calving loses are ultimately determined by a series of discrete, stochastic events. Current literature has made great strides in explaining the overarching controls on annual to decadal calving rates, but current literature lacks predictive systems to capture individual events. Attempts have been made to relate calving events to changes in water level (Haresign, 2004), water temperatures and subaqueous melt (Motyka et al., 2003; Rohl, 2006; Robertson et al., 2012), and strain rates (Boyce et al., 2007; Benn et al., 2007a). While these studies have elucidated several important controls on calving rates in lacustrine and marine environments, they have also served to emphasize the uniqueness and variability within and between each system. This uniqueness also extends into the relative importance of calving in each system. While some glaciers have shown massive recent ice losses due to calving (Van der Veen, 1996; Warren and Kirkbride, 2003), others have been shown to lose relatively little mass to calving processes (Boyce et al., 2007; Tsutaki et al., 2011).

This chapter aims to contribute to a better understanding of the roles and importance of calving at Bridge Glacier during the 2013 melt season, and has three main goals. First, it aims to identify the timing, nature, and relative size of individual calving events observed during the study period. Secondly, this chapter examines whether individual calving events can be related to meteorological, limnological or glaciological observations on a sub-seasonal timescale. Finally this chapter sets out to quantify the volume of ice lost through calving during the 2013 melt season and relate it to the volume of ice lost through surficial melt in order to assess the relative importance of calving on the glacier’s mass balance.
5.2 Methods

The volume of ice discharged through calving from the glacier terminus, $Q_{\text{calving}}$ ($\text{m}^3\text{a}^{-1}$), termed the calving flux (Figure 5.1), can be quantified as

$$Q_{\text{calving}} = \left( \frac{dA_T}{dt} + U \times x \right) H_I \tag{5.2}$$

where $\frac{dA_T}{dt}$ is the rate change in glacier area at the terminus ($\text{m}^2\text{a}^{-1}$), $U$ is the terminus flow velocity ($\text{m}a^{-1}$), $H_I$ is the height of ice at the terminus, and $x$ is the cross sectional area at the terminus ($\text{m}$). Ice front melt is assumed negligible compared with the magnitudes of flow velocity and terminus retreat, as well as the errors associated with estimating each variable.

![Figure 5.1: Bird’s eye view of a conceptual model of calving flux.](image_url)

5.2.1 Terminus Area Change

The rate of change in glacier area ($\frac{dA_T}{dt}$) during the study period was found by comparing Landsat images from June 23 and September 11, 2013. Historical terminus positions (1984 - 2012) were derived from Landsat images taken from mid-September to late October of each year (see Section 2.2). Landsat images were acquired using the LandsatLook Viewer (http://landsatlook.usgs.gov/). Shapefiles for end-of-season terminus positions were generated by manual deliniation from Google Earth imagery. The change in area was then calculated using routines developed in R, using the rgeos package.
5.2.2 Terminus Flow Velocity

The terminus flow velocity was measured by tracking features from two time lapse cameras pointed at the terminus and roughly 1 km above the floating terminus (Terminus and Nunatak TLC). Points were manually tracked using Tracker video analysis and modeling tool (https://www.cabrillo.edu/~dbrown/tracker/). Raw pixel displacement was converted into real-world displacement values using known camera and control point geometries (see Chapter 4).

Eight on-glacier points are tracked from each camera throughout the study period using daily noon-time images. All tracked points from the Terminus TLC are located on the floating terminus, and within 200 m of each other, while Nunatak TLC points were all selected near the centre flowline, within 300 m. Filtering routines discarded roughly 10% tracked data points due to negative displacement, or loss of target. Final terminus and nunatak velocity time series are generated by averaging the daily displacement for each tracked point, and an averaged summer velocity is found by averaging the total displacement for each tracked point during the study period.

5.2.3 Bathymetry and Ice Thickness Estimate

In order to convert changes in area into volumetric ice loss, a suitable estimate of glacier thickness must be derived. The terminus of Bridge Glacier sits in a large lake, and has a notable inflection point (see Figure 5.2) where it is assumed that the terminus transitions from grounded to floating. Large tabular calving events during the melt season have shown limited mobility immediately after calving, suggesting that the ice is right at the boundary criterion for flotation. Given these observations, the thickness of ice at the terminus can be approximated by assuming the glacier is at the threshold of flotation. Using the height above buoyancy criterion (Van der Veen, 1996; Benn et al., 2007b) where

\[ H_f = H_b + \frac{\rho_w}{\rho_i} D_W \]  

(5.3)

and \( H_b \) is the height of ice above the waterline (m), \( D_W \) is the depth of water (m), and \( \rho \) is the density of ice (\( \rho_i, 917 \text{kgm}^{-3} \)) and lake water (\( \rho_w, 1000 \text{kgm}^{-3} \)) respectively, the ice thickness can be calculated.
5.2. Methods

Figure 5.2: Photograph of the terminus inflection point, indicating flotation noted by red arrows. Yellow arrow indicates large crevasse that eventually leads to calving of the large tabular iceberg to the left of arrow.

Assuming the terminus is floating, the minimum ice thickness can be calculated as

\[ H_I = H_b \left( \frac{\rho_w}{\rho_i} - 1 \right)^{-1} \]  

which is preferable in many cases because it allows for supplemental *in situ* observations of the height of ice above the waterline, which can constrain our estimation of ice thickness.

To acquire water depth near the terminus, a bathymetric survey was conducted using a Lowrance HDS Gen2 fishfinder. The vertical range is given at 3000 ft while the horizontal GPS accuracy is estimated at 5 m. A total of 893 points were taken in an irregular pattern, while some areas in immediate proximity to the terminus and near Lake AWS, were choked full of icebergs and were unable to be surveyed. An additional 47 manual points were taken by linear interpolation between surveyed points to improve coverage in low spatial coverage areas. Bathymetric points were processed using R including routines from the *gstat* package (Pebesma, 2004). Irregular point data was interpolated onto a regular 10 m grid using inverse distance weighting (maximum 4 neighbours). The points
were overlain on a shapefile of Bridge Lake generated through Google Earth and Landsat imagery.

### 5.2.4 Limnology and Calving Events

Water level and temperature were recorded using a HOBO U20 Water Level Logger. The pressure transducer measured 10 minute mean pressure (which is converted into a depth) and water temperature, and is accurate to 0.5 cm and 0.44°C. The pressure transducer was housed inside a radiation shield, and attached to an anchor. The transducer was deployed June 18, 2013 into 1 m of water, roughly 5 m offshore, in the north end of the lake near the terminus.

Water level, $W_D$ (m), was calculated as

$$W_D = \frac{P_L - P}{g \rho_w}$$

(5.5)

where $P_L$ is the measured transducer pressure, $P$ is the atmospheric pressure obtained from the Lake AWS barometer (hPa), $g$ is the gravitational constant (9.81 m s$^{-2}$), and $\rho_w$ is the density of water (1000 kg m$^{-3}$).

Water levels were found to fluctuate due to waves, and also to rise improbably over very short time steps, most likely due to waves displacing the transducer or forcing it to slip and sink deeper into the lake. In order to remove these suspected artificial signals, the water level data was filtered. A routine was written to separate artificial from true water level fluctuations based on a 3 cm threshold. If the differences in water level between a time step ($t_i$) and the subsequent datapoint ($t_{i+1}$) were greater than the threshold, and two time steps past the data point of interest ($t_{i+2}$) and the time step preceding the data point of interest ($t_{i-1}$) were also greater than the threshold, the time step is flagged as an ‘artificial shift’. In order to correct for this shift, the magnitude of the shift is subtracted from all data points after the event. Several different thresholds were tested, and smaller thresholds were found to over-prescribe ‘artificial shifts’, while larger thresholds did not capture clearly visible shifts in the data. Given the current dataset and lake size, a 10 minute mean shift of over 3 cm is highly improbable. In total, 7 time steps were flagged as ‘artificial shifts’ (see Figure 5.3).

In order to capture the timing and relative size of individual calving events, the terminus was tracked by a time lapse camera located at Ridge AWS. The camera acquired pictures
5.3. Results

5.3.1 Calving Events

During the 2013 summer study period, the vast majority of the calving flux can be explained by three major calving events. On June 23 at 01:00 hrs the first major calving event is observed (Figure 5.4a and 5.4b). This event is characterized by a large tabular iceberg (roughly 400 m × 500 m) failing along a series of interconnected crevasses, removing the V-shaped protrusion from the centre of the terminus. The failure was preceded by several days of the slow propagation of the crevasses towards one another.

The second major calving event was observed August 2 at 13:00 hrs (Figure 5.5a and 5.5b). This calving event (roughly 100 m × 150 m) failed on the south side of the terminus. Similar to the previous major event, this event was preceded by up to a week of the slow propagation of a crevasse. During that period, the prow of the terminus in that region raised substantially above the surrounding ice, tilting backwards towards the terminus before ultimately failing.
5.3. Results

(a) 18:00 hrs, June 22, 2013.

(b) 06:00 hrs, June 23, 2013.

Figure 5.4: Large tabular calving event, June 23, 2013. Note large crack propagating along the full extent of the terminus protrusion.
5.3. Results

(a) 12:00 hrs, August 2, 2013.

(b) 13:00 hrs, August 2, 2013.

Figure 5.5: Moderate calving event, August 2, 2013. Note the uplifted prow of the terminus at far looker's left (south).
5.3. Results

(a) 06:00 hrs, August 21, 2013.

(b) 07:00hrs, August 21, 2013.

Figure 5.6: Large calving event on looker’s right sidewall, August 21, 2013. Note large crack propagating up-glacier along terminus (looker’s right).
The third and final major calving event during the summer was observed on August 21 at 07:00 hrs. This event was roughly 800 m × 100 m, and involved a failure along the full north facing (longitudinal) length of the terminus, starting near the identified inflection point between floating and grounded terminus (Figure 5.2). The ice appears to have failed along a series of interconnected crevasses. In the days leading up to the event, the terminus ice front rose several metres, and fell backwards as it calved. The event was characterized by multiple smaller events that occurred within several hours of each other (Figure 5.6a and 5.6b).

During the summer, there were also numerous small calving events observed where masses of ice less than 10 m in width failed along the ice-front. In most cases, this calving event exhibited different behaviour than the high magnitude, low frequency nature of the major calving events. These events appear to calve almost exclusively above the waterline, failing along a crevasse. Rough estimates of the size of calving events yields a volume of 0.032 km$^3$ assuming an average 110 m ice thickness.

5.3.2 Calving Correlations

Throughout the summer of 2013, there does not appear to be any clear relationship between major calving events, and pertinent glaciological, limnological or meteorological variables. Throughout the study period, the melt rate obtained from both a distributed energy balance model (DEBM) and temperature index model (TIM) is relatively consistent, oscillating between 5 and 10 cm/day (Figure 5.7a.). Melt peaks in early July, and again in early August. Several large storm events depress the melt rate for several days in late June, mid-August and early September. No large calving event is preceded by higher than normal melt days. In fact, the highest summer melt rates coincide with the longest non-calving period (July) during the study period.

The lake level has both diurnal and seasonal oscillations (Figure 5.7b.). Daily variations show a peak in lake level in the mid-to-late afternoon, roughly 17:00 hrs, with minimum levels occurring near sunrise (06:00 hrs). Seasonally, lake level rises through June and early July, before dropping almost to pre-summer levels in mid-July. The lake level rises again in August, before another drop in late August. Rises in lake level follow sustained periods of high melt rates, while drops follow storm events, which have low melt rates (even when significant rain occurs). Daily mean lake level is significantly positively correlated with mean daily air temperature and modelled glacier melt rates. Both the June and
5.3. Results

Figure 5.7: Measured melt rate, water level, water temperature and running mean (3 days) terminus velocities for two sites at Bridge Glacier. Calving events are shown as thick grey lines.

early August calving events coincide with an increase in daily mean water level; however, all three events also coincide with near-minimum diurnal (10-minute mean) water level variations.

During the study period, water temperature is found to decrease from 1.5°C to 0.9°C; however, this decrease is far less than the observed daily variability (±1.5°C). More in-
5.3. Results

depth lake temperature observations (L. Bird, pers. comm.) suggest that the lake is well mixed, and that temperatures are relatively uniform both spatially and with depth. It is likely that the variability observed here is slightly elevated due to the sensor’s proximity to the shore. Daily maximum water temperatures are observed in the late afternoon (roughly 16:00 hrs), while minima are observed close to sunrise (05:00 hrs) (Figure 5.7c.). Overall, there does not appear to be any discernible short-term relationship between calving events and water temperature, except for the final (late-August) calving event, where water temperature drops substantially following the event. This event is closest to the location of the water temperature probe, making it likely that newly discharged ice from the calving event may have flowed quite close to the sensor, depressing water temperatures in the vicinity. Seasonally, water temperature decrease coincides with higher melt rates and more calving. Both of these processes allow for water near the freezing point to be added to the lake. Minimal stratification and winds blowing icebergs several 100 m would enhance mixing and further depress temperatures.

Finally, there does not appear to be any relationship between calving events and tracked daily glacier flow velocities at either the terminus or up-glacier near Nunatak TLC (Figure 5.7d.). Terminus velocity remained relatively constant throughout the study period, oscillating between 0.35 and 0.50 m/day. There is no discernible seasonal trend, and observed velocities do not appear to respond to either changing melt rates or water level. However, nunatak measured velocities show a doubling of flow speed in early July, following a large spike in melt rate. The elevated flow speeds remain for over a week, before settling to a relatively consistent velocity, slightly lower on average than those observed along the floating terminus. Observed velocities show a slight decrease immediately following calving events; although this trend is minor, and the change in velocity is well within the measurement error for the method.

5.3.3 Calving Flux

The change in terminus area in 2013 is $2.97 \times 10^{-1}$ km$^2$ over the 89 day period. The average terminus velocity during the study period was measured as 139 ma$^{-1}$, while the terminus cross-sectional distance was 1055 m, giving an additional area change of $3.42 \times 10^{-2}$ km$^2$. Due to the paucity of water depth measurements close to the terminus, water depth was estimated from a cross-section adjacent to the June 2013 terminus position (see Figure 5.8). The median depth was 109 m, corresponding to a height above buoyancy of 9.9 m,
and an estimated ice thickness of 109 m. Combining these terms in Equation 5.2 yields an estimated calving flux of $3.62 \times 10^{-2}$ km$^3$ for the 85 day study period.

Comparing the volume of mass lost through calving with the volume of surface ice melt during the same period (Chapter 3) yields a total mass loss of $1.60 \times 10^{-1}$ km$^3$ of ice (Figure 5.9). For the study period, calving accounts for 23% of the mass loss, equivalent to an additional 1.3 m of surficial ice melt in the ablation area.
5.4. Discussion

5.4.1 Potential Calving Mechanisms

Calving events observed during the 2013 melt season share several commonalities in character and magnitude, and are consistent with observations from other lacustrine glaciers with floating termini (Dykes et al., 2011; Boyce et al., 2007). First, the observed calving events are relatively long processes, taking upwards of several days for crevasses to fully propagate to allow the ice to be discharged. Secondly, in the days and weeks leading up to the calving event, the ice margin tilted backwards against the terminus, exposing small
meltwater notches as the ice separates from the terminus (see Figure 5.10). Both of these observations suggest that buoyancy induced torque is the primary mechanism responsible for calving.

Figure 5.10: Ice front tilting backwards against the terminus prior to calving, exposing small meltwater notches, June 17, 2013, 5 days prior to calving event.

A backwards tilt of the ice margin, most clearly observed in the August 2 and August 21 events (see Figures 5.10, 5.11), suggest the presence of an ice-foot, similar to what has been observed at Tasman Glacier (Rohl, 2006; Dykes et al., 2011). Cold water near the terminus likely depresses melt rates below the waterline, save for a shallow near-surface layer that experiences daily warming. Given minimal expected subaqueous melt, it is likely that the ice front extends further into the lake below the waterline. This greater volume of ice below the waterline destabilizes the terminus due to an upwards buoyant force. This process is in contrast to warmer tidewater settings (Rignot et al., 2010; Motyka et al., 2003), where subaqueous melt is a significant source of ice loss, and can lead to terminus undercutting.

The observed tilt of the ice margin can also be explained by basal crevassing, created as the glacier terminus transitions from grounded to floating (Benn et al., 2007a). A clear inflection point (see Figure 5.2) along the terminus delineates the point at which the glacier begins to float in the lake, allowing for substantial torque and possibly basal crevassing. This mechanism also explains why all major calving events are observed in the buoyant region of the terminus.
5.4. Discussion

Ultimately, timing and the position of individual calving events appears to be most strongly controlled by the location, orientation, and size of crevasses in the buoyant terminus. One commonality with all three major calving events in 2013 was the slow calving of major tabular icebergs. This suggests that the gradual expansion, propagation, and joining of individual crevasses, due to shear and surface melt, form a ‘preferential line of weakness’ which allows a margin for calving (Benn et al., 2007a). This suggests that the nature of buoyant calving at Bridge Glacier is progressive and gradual, rather than a rapid, catastrophic failure brought about by abrupt perturbations in conditions.

5.4.2 Calving Forcings

An examination of daily and weekly (sub-seasonal) meteorological, glaciological and limnological variables yields no significant relationship with the timing or magnitude of calving events. This finding reflects the significant control of geometry and crevassing on the location and extent of calving. The relative insensitivity of calving events to changes in water level contrasts with observations in Patagonia and Iceland (Haresign, 2004), but is consistent with observations from Mendenhall Glacier, AK (Boyce et al., 2007). While changes in lake level should affect the balance of hydrostatic forces against the terminus and thus promote buoyant torque, it appears that these factors are muted when the ter-
5.4. Discussion

minus has already achieved flotation. A combination of the relatively low magnitude of diurnal and seasonal changes (< 5 cm, < 50 cm, respectively), combined with a floating terminus, creates conditions where lake level changes have a relatively small influence on the terminus force imbalance. This influence is likely further muted because the glacier is not close enough to being grounded to allow changes in lake level to affect the buoyancy of the terminus.

Figure 5.12: Photograph of terminus melt notches, July 19, 2014, which are exposed as the terminus flexes upwards due to buoyancy.

The relatively low water temperatures observed during the summer suggest that undercutting due to melt below the waterline is not a significant factor affecting calving at Bridge Glacier. Meltwater notches are observed in close proximity to the waterline; however, these notches are in the order of several centimetres (see Figure 5.12), and do not appear to measurably affect the terminus force imbalance. This hypothesis corresponds with very few observations of oversteepened ice-front failure, indicative of high subaqueous melt rates. Conversely, cool water promotes the growth of an ice foot along the terminus, due to lower subaqueous melt, relative to that above the waterline. The presence of an ice foot should serve to destabilize the terminus, and is consistent with observations of pre-calving terminus tilting (Figure 5.11). While the low lake temperatures should destabilize the terminus, and alter the character of calving events, its overall effect on the rate
of calving is muted since ice foot development is a slow process and also relies on velocity and surface melt to create the necessary force imbalance to promote failure.

Changes in glacier flow speed near the terminus did not coincide with observed major calving events. The observed differences in variability and average magnitude of the two velocity time series suggest that there is substantial strain between the floating and grounded terminal zone, and is supported by the presence of heavy crevassing as the glacier spills over a final rock bench into the lake. However, the majority of the glacier terminus in contact with the lake is not significantly crevassed, suggesting that strain is not significant at the surface in the floating portion of the terminus.

In general, crevassing patterns are complicated, although it is expected that strain should be highest at the transition from the grounded to floating ice margin, due to the reduction of basal drag. This signal is delayed, however, by the need for crevasses to propagate and connect before calving can occur. Furthermore, crevassing is also likely due to buoyant torque, something that is not adequately captured from surface velocity gradients. While glacier velocity, and more generally strain in the terminus region, will have an impact on the frequency, location and magnitude of individual calving events, the process has a much longer time scale than can be adequately captured by discrete changes in velocity patterns.

Finally, while the intensity and cumulative effects of surface melt certainly play a physical role in calving dynamics at the terminus, they do not show any relationship to individual calving events. Surface melt should reduce the volume of ice above the waterline, in turn increasing buoyant torque and the potential for calving induced failure, however, this forcing is cumulative rather than intensity based. Furthermore, melt is also directly related to other variables affecting forces acting on the terminus, such as lake level, and indirectly related with others, such as grounded ice velocity (due to the fact that more basal water supply reduces basal drag). Both of these factors complicate any meteorological control on major calving events.

While calving in the floating margins of Bridge Glacier is the product of multiple interrelated sub-seasonal variables, calving events are ultimately triggered by changes that are cumulative, rather than stochastic. Furthermore, the interrelation of relevant calving variables further complicates matters, and makes isolating individual factors and assigning causality difficult. Ultimately, although there is a physical basis for certain meteorological variables triggering individual calving events, these factors, at least at Bridge Glacier, act within the imposed bounds of the geometry and bathymetry of the terminus region. These
findings suggest that calving in floating lacustrine glacier termini does not significantly respond to sub-seasonal variability, and instead is the product of cumulative, multi-seasonal changes in crevassing, ice thickness, and buoyancy.

5.4.3 The Relative Importance of Calving

While calving still contributes significant ice loss that would not otherwise be possible, it is a comparatively small contributor of mass loss relative to climatic melt at Bridge Glacier during the 2013 melt season. Calving losses in this system are controlled by several glaciological, and topographical controls that ultimately limit the magnitude of the calving flux. While both high frequency, low magnitude and high magnitude, low frequency events were observed, rough calculations of the size of the three main calving events (Section 5.3.1), and a detailed calculation of the total summer calving flux (Section 5.3.3), suggest that roughly 90% of the calving losses can be explained by the three major events.

The calving flux is limited due to the fact that it only occurs over a very small percentage of the total glacier area. The cross-sectional area at the calving margin is just over 1 km wide, which strongly limits the volume of ice that can reach the floating terminus, in turn limiting the size of calving events. Conversely, the end-of-season ablation area was over 27.6 km$^2$ in 2013. The large contributing area for melt, relative to calving, allows for melt processes to contribute a substantially larger volume of ice loss than from calving processes.

The calving flux is also limited by relatively modest glacier flow speeds at the terminus. Velocity is limited at Bridge Glacier by a gentle gradient in the lower reaches of the glacier, and relatively narrow valley side-walls, both of which inhibit the speed at which ice can be delivered to deep lake waters at the terminus where it becomes buoyant and calves. Near-terminus flow speeds are 1 to 2 orders of magnitude smaller than observed at large tidewater calving glacier systems in Patagonia and Greenland (Rivera et al., 2012; Koppes et al., 2011; Meier and Post, 1987).

The calving flux is also limited by the bathymetry of Bridge Lake. While the terminus of Bridge Glacier is, at least partially, buoyant, it is dependent on the ice thickness of the terminus region. Any significant thickening, which would theoretically increase the potential volume of ice lost due to calving, would also serve to reduce buoyancy of the terminus. If the terminus were grounded, it would serve to stabilize, and significantly reduce calving losses. Therefore, any major increase in terminus thickness is more likely to reduce, rather than enhance, the calving flux.
5.5. Conclusions

These limitations on the magnitude of calving flux suggest that climate is the main driver of ice loss at Bridge Glacier. Although calving supplies a significant portion of ice loss at Bridge Glacier, three times as much ice is removed due to surface ablation. This has significant implications for the future health of Bridge Glacier as it retreats further up-valley, its terminus eventually grounds, and it no longer experiences major calving events. While it would be tempting to suggest that the glacier will begin to gain mass again, these findings show that calving losses, at least for the 2013 melt season, are within the range of expected variability in seasonal melt. While the future health of Bridge Glacier may depend in part on the changing magnitude and frequency of calving, ultimately it is more dependent on future climatic conditions.

5.5 Conclusions

During the 2013 melt season, Bridge Glacier experienced three major calving events, which were responsible for the vast majority of the seasonal calving flux. The events, in late June, early August, and late August, are characterized by the discharge of large tabular icebergs. The events were preceded by up to several weeks of crevasses propagating and joining, ultimately causing the ice front to tilt backwards against the terminus and calve. The character and magnitude of the three major calving events suggest that buoyancy is the primary driver of calving, while crevassing provides the margin where ice calves.

A high temporal resolution examination of meteorological, limnological and glaciological variables related to calving found no significant relationships with individual calving events. The surface melt rate, lake water level and temperature and terminus velocity ultimately have an effect on the nature and timing of calving events; however, their contributions are cumulative rather than intensity-based. The current environment, characterized by a floating terminus in cold, well mixed water, dampens any influence of individual variables. Calving events are more strongly controlled by seasonal and annual changes, rather than rapid perturbations in conditions.

During the 2013 melt season, Bridge Glacier lost \(3.62 \times 10^{-2} \text{ km}^3\) of ice through calving, accounting for 23% of total ice loss between June 20 and September 12, 2013. While calving is an important component of mass loss at Bridge Glacier, it contributes much less mass loss than surface melt in the much larger ablation area. Calving is limited by a modest terminus flow velocity, a relatively narrow cross-sectional area, and lake depth, all of which limit the amount of ice that can be supplied to the floating terminus.
This research suggests that calving is a significant source of ice loss at Bridge Glacier; however, the glacier is still substantially more influenced by climatic melt. While annual variations in calving can affect the mass balance of Bridge Glacier, the glacier is more sensitive to changes in the melt rate. As Bridge Glacier retreats to shallower waters (and eventually dry land), this research suggests that although future calving losses may be reduced, and ultimately cease completely, the mid to long-term mass budget of Bridge Glacier is at the mercy of the much larger influence of a changing climate.
Chapter 6

Calving, Retreat, and Surface Melt: Present and Historical Conditions

6.1 The Relationship Between Calving and Climate

In Chapter 2, the retreat of Bridge Glacier between 1971 and 2012 was reconstructed using satellite imagery. The climatic influence on this retreat was examined using proxy records from nearby stations, and a first order estimate of the impact of calving was made by modelling the expected retreat, had Bridge Glacier not been terminating in a lake. This study suggests that climatic warming was responsible for roughly 700 m (20%) of the observed 3.55 km retreat between 1971 and 2012; however, it does not consider how changes in retreat relate to volumetric changes.

Observed retreat was found to correspond quite well with the modelled climatic response until 1991. This transition coincided with the beginning of observations of tabular icebergs in Bridge Lake, suggesting the terminus had achieved flotation. From 1991 to present, the retreat rate diverges from the expected climatic response, suggesting that once a glacier achieves flotation, its retreat rate is driven primarily by calving rates. This finding is in agreement with studies in other lacustrine and marine environments (Post et al., 2011; Boyce et al., 2007; Warren and Kirkbride, 2003), which shows that once a glacier achieves flotation, its climatic signal is modified by the overarching control of calving dynamics.

That chapter also suggests that while the response of calving glaciers that achieve flotation can become desensitized to climatic forcing, the initial conditions to achieve this state are controlled by long-term climatic trends. In order to achieve flotation, the glacier must thin substantially in order to become buoyant. Given no likely increase in flow speeds and extensional flow to drive dynamic thinning before 1991, the most probable source of
thinning is due to a run of years with strongly negative mass balances.

Climate is responsible for the initial thinning required for the terminus to achieve flotation; however, once the terminus is buoyant, glacier velocity is expected to increase, due to a decrease in basal drag. This increase in velocity is enhanced by calving events at the terminus, creating a mass deficit near the terminus, and ‘drawing down’ more ice from the upper reaches of the glacier. In turn, the ‘draw-down’ promotes further thinning of the terminus, and further calving. This ‘runaway effect’ emphasizes the dominant role of buoyancy on terminus stability, and the degree to which buoyant calving glaciers can retreat more or less independently of climate.

6.2 Conditions at Bridge Glacier - 2013

6.2.1 Ablation

In Chapter 3, ice melt for the 2013 study period is reconstructed through a combination of in situ observations and modelling. Melt was reconstructed to be up to 5.9 m (w.e.) of ice loss in the lowest reaches of the glacier, translating to a volumetric ice loss of 0.124 km$^3$. Melt is shown to be strongly dependent on solar radiation, particularly over low albedo ice. The spatial distribution of temperature, humidity and wind speed is found to be strongly controlled by high magnitude katabatic winds that persist through 95% of the summer study period.

The geometry of Bridge Glacier plays a substantial role in both the magnitude of ice loss, and the spatial distribution of meteorological parameters. A large, low-elevation basin, combined with deep valley walls in the lowest reaches promote strong, persistent katabatic winds. The winds are funnelled by glacier side-walls, originating in a large up-glacier basin. Flow path lengths are an effective predictor with which to model the spatial variation of katabatic winds, and thus improve turbulent heat flux estimates in the ablation area.

6.2.2 Glacier Flow

In Chapter 4, a method is presented to measure glacier velocity using time lapse cameras and manual feature tracking. The method is computationally simple and relatively cost-effective, relying only on consumer grade hardware. The method produces velocity measurements that have an uncertainty of 0.45 pixels/day, corresponding to 0.12 md$^{-1}$ at 1 km from the target. This uncertainty allows for a reliable velocity time series at a
6.2. Conditions at Bridge Glacier - 2013

The temporal resolution of one measurement per day, with the potential for higher frequency measurements for closer targets or faster moving glaciers.

Bridge Glacier flow speed was measured at two sites: along the north flank of the floating terminus, and 1 km up-glacier, along a horizontal transect. Velocities were found to average 139 m\(^{-1}\) over the study period, while faster flow was observed closer to the centreline of the glacier, reaching average speeds of over 168 m\(^{-1}\). The time series shows a doubling of flow speed in early July, with highest magnitude changes closest to the glacier centerline, coinciding with high melt rates. This observation follows classical glacier flow theory (Nye, 1952), which suggests that the drainage of basal water, collected from early summer melt, temporarily reduces basal drag and allows for faster flow. Up-glacier velocities became more consistent later in the summer, and oscillated around 0.4 m/day.

Measured velocities in the terminus region were found to be, on average, 138 m\(^{-1}\). While the net displacement is found to be similar between both grounded and floating sites, the timing of those flow speeds varied. In contrast to the up-glacier measurements, velocities along the terminus were relatively consistent, reflecting a lack of drag. In all periods except early July, flow speeds were roughly 0.1 m/day greater at the terminus. The difference in flow speed is enough to promote the strain necessary to instigate crevassing in the lower reaches of the glacier. These surface crevasses were found primarily along a bench between the two camera sites, where the glacier spills into the lake. The relatively small velocity gradient between the two sites suggests that there is only minor spatial variability between in the lowest reaches of the glacier.

6.2.3 Calving at Bridge Glacier

In Chapter 5, the calving flux is reconstructed for the 2013 melt season. The volume of ice loss due to calving is found to be 0.036 km\(^3\), the majority of this loss occurring in three major calving events (June 20, August 2, and August 22, 2013). All three events failed along a series of interconnected crevasses, and occurred in the buoyant region of the terminus.

Observed Calving Mechanisms

Over the study period, calving was characterized by two distinct types of events. Low magnitude, moderate-high frequency events occurred throughout the study period, failing along the ice front. This event was found to involve a failure along fractures several metres
from the ice front, which was most likely due to a force imbalance along the terminal ice cliff (see Figure 6.1). Slightly warmer surface water undercut the ice front due to thermal erosion, and ultimately created an overhanging ice cliff that failed along a point of weakness, most often a near-terminus crevasse.

Figure 6.1: Photograph of a terminus force imbalance. The potential locations of future failure are suggested by red arrows. For scale, the ice cliff is approximately 10 m tall.

The second type of calving observed at Bridge Glacier in 2013 was characterized by a high magnitude, low frequency event that occurred along a series of interconnected crevasses. The size of icebergs discharged from this mechanism, relative to the calculated calving flux (roughly 90%, see Section 5.3.1 and 5.3.3), suggest that this mechanism was responsible for the majority of the calving flux during the 2013 study period. All three events had similar character: slow (several days) processes where the ice cliff failed along a series of crevasses. In all three cases, the terminal ice cliff tilted backwards in the days and weeks preceding the event, exposing thermal notches below the waterline. These observations suggest that buoyant torque was the primary force initiating calving events.

It is likely that in order for buoyant torque to trigger calving events, crevasses are required along the base of the glacier as well as at the surface, similar to what has been proposed elsewhere Benn et al. (2007a). During the glacier’s transition from grounded to floating, torque at the base of the glacier is most likely great enough to create crevasses along the inflection point. As the glacier advances further into the deep lake waters, basal crevasses can eventually connect with surface crevasses, providing a preferential line of weakness on which calving can occur.
Primary Controls on Calving

In Chapter 5, the timing of the three major calving events is compared with meteorological, limnological and glaciological variables. Over the study period, short, rapid perturbations in water level and temperature, surface melt and glacier flow speed have only a minor association with individual calving events. Although there is a physical basis for threshold conditions to promote calving, those conditions are achieved cumulatively rather than through daily (or sub-daily) perturbations.

The cumulative conditions required to trigger calving events are likely dampened by terminus flotation. Since the glacier is not grounded, changes in water level and ice thickness, within the range observed during the study period, do not appear to alter the force balance at the terminus. These findings suggest that buoyancy is the main control on terminus stability in lacustrine environments. In environments where the terminus is buoyant, calving events, and ultimately the flux of ice discharged, is determined by the cumulative effects of glacier flow, water depth, and surface melt creating the conditions required to provide the necessary buoyant torque for individual calving events.

6.3 Relative Importance and Long Term Implications

6.3.1 Relative Importance of Calving at Bridge Glacier 2013

By combining surface melt calculations from Chapter 3 with the calving flux calculations from Chapter 5, the summer balance \( b_s \) can be determined. For the 2013 study period, calving is responsible for just under 23% of the ice loss from Bridge Glacier. Calving losses are limited by a modest flow speed, small cross-sectional area at the terminus, and the bathymetry of the lake. These factors serve to limit both the rate at which ice can be supplied to deep water, and the thickness of ice that will remain buoyant. Absent significant inter-annual changes in the flow speed or ice thickness, there is a relatively narrow range of potential calving losses. These constraints, coupled with a large ablation area, ensure that surface melt is a consistently greater supplier of ice loss, and more substantial impact on the overall mass balance of Bridge Glacier.
6.3.2 Long-Term Mass Loss Fluctuations

In order to constrain the representativeness of the 2013 melt season, and better understand the long-term variability of calving at Bridge Glacier, estimates of historical calving fluxes and surface melt are derived. Historical surface melt is estimated using ELA observations and a fitted linear mass balance gradient (Shea et al., 2013). Below the snowline, the net balance (where $b_n = b_s$) is estimated using the 2013 glacier hypsometry, where

$$b_n(z) = b_1(ELA - z) \quad (6.1)$$

and is calculated for the elevation of every point, $z$ (m a.s.l.), below the ELA.

The coefficient value ($b_1 = 6.62$ m(w.e.)/m) taken from Shea et al. (2013) underestimates the volume of ice loss during the 2013 melt season calculated using distributed energy balance modelling (Chapter 3). The coefficient $b_1$ is tuned using the mass balance gradient estimated from the DEBM ($b_1 = 9.07$ m(w.e.)/m, Figure 6.2), and is used for all years.

The glacier area is determined from the end-of-season calving margin. All points that have calved off are given an elevation of 1400 m (a.s.l.) and are considered in Equation 6.1. Historical ELAs are measured from end-of-summer (between mid-September to mid-October) Landsat images from 1984 to 2013.

Errors in ELA-derived mass balance calculations are estimated by assuming a 75 m uncertainty in measuring the ELA, due to timing of available Landsat images, or 22% according to Shea et al. (2013), whichever is greater. The large error in ELA estimate is to account for errors that cannot be adequately quantified without additional historical data, such as the linearity of the mass balance gradient.

Historical terminus positions, lake bathymetry, and estimates of ice thickness are made using field data from the 2013 field campaign, and methodology discussed in Section 5.2. Historical velocities were assumed to be equal to the 2013 summer flow speed (140 m a$^{-1}$), and calving calculations are run with 70 m a$^{-1}$ (50%) potential annual variability. A 60 m uncertainty in measuring the terminus cross-section ($x$) (equal to 2 Landsat pixels) is applied. The rate of change in terminus area ($\frac{dA}{dt}$) uncertainty is estimated as 7200 m$^2$a$^{-1}$ (2 × 60 m × 60 m). The ice thickness uncertainty is estimated as 5.6% (from Section 5.3.4) plus an additional 10 m to account for changes in sedimentation, and ice thickness relative to water depth. Before 1991, the terminus was not floating; therefore, an ice thickness uncertainty of 60 m is estimated to account for a range of grounded terminus geometries.
6.3. Relative Importance and Long Term Implications

Between 1991 and 2004, bathymetry has poor data coverage, and a water depth uncertainty of 30 m is estimated (giving an additional ice thickness uncertainty of 33 m).

During the study period, the volume of ice lost from surface melt is variable and shows only a minor decrease over time. Between 1984 and 2013, the ELA varied from 1926 m to 2202 m; however, most years the ELA fell between 2050 m and 2150 m, leading to a volumetric ice loss standard deviation of 0.018 km$^3$. The 2013 study season is above average, but within one standard deviation of the mean ($\bar{x} = 0.107$ km$^3$) for the 30 year period. Surface melt shows a minor decrease over time, which can be attributed to the loss of area in the lowest reaches of the glacier ($z = 1400$ m) due to calving.

Calving returns are minimal before 1991, due to the relative stability of the grounded terminus. From 1992 to 1994, the calving flux increases to 0.020 - 0.029 km$^3$ (19 - 27% of the total annual ice loss) before a two year period of low flux (< 0.015 km$^3$). From 1997 to 2000, calving losses increase (0.023 - 0.052 km$^3$), before settling into another period of relative stability in 2001 and 2002. The highest calving fluxes occur between 2003 to 2006.
6.3. Relative Importance and Long Term Implications

Figure 6.3: Historical ice loss from calving and surface melt, 1984-2013. Dark vertical line corresponds with terminus flotation.

(0.030 - 0.084 km$^3$) and again from 2008 to 2011 (0.036 - 0.100 km$^3$) with a period of stability in 2006 and 2007.

Calving losses are characterized by several years of high flux, and periods of relative stability. The magnitude of the calving losses increased once the glacier achieved flotation in 1991. The calving flux increased again from 2005 to 2010, and the magnitude of surface melt losses decreased, making the calving flux a much larger contributor of ice loss. The volume of ice loss due to calving is roughly equal to the volume lost through surface melt in 2005, 2008 and 2010 (44 - 49% of total volumetric ice losses).

The pattern of several high magnitude calving flux years followed by several low-flux years suggests that calving has the potential to contribute large volumetric ice losses over single seasons. However, high magnitude calving losses do not persist over more than a few consecutive seasons, suggesting that large calving fluxes are part of a transient stage in the glacier’s retreat. It is also possible that this pattern reflects the glacier’s response to large calving events. Large calving fluxes could potentially restore stability to the terminus, as a result of a change in geometry, and a redistribution of buoyant forces acting on the...
6.3. Relative Importance and Long Term Implications

6.3.3 Bridge Glacier Relative to Other Lacustrine Calving Systems

This study above sheds light on the main controls and drivers of mass loss at Bridge Glacier. In order to draw broader conclusions about the nature and variability of mass loss in lacustrine glacier systems, Bridge Glacier is compared to other similar systems across the globe (see Table 6.1, revisited from Table 1.2). Bridge Glacier falls in the middle of a continuum of magnitude and frequency in lacustrine calving systems. The calving rate for Bridge Glacier (281 ma\(^{-1}\) in 2013) is larger than that for smaller New Zealand glaciers, such as Maug, Grey and Hooker (Warren and Kirkbride, 2003), as well as Mendenhall Glacier in Alaska at the end of the 1990s (Motyka et al., 2003; Boyce et al., 2007). Conversely, calving rates at Patagonian glaciers Leon and Ameghino are up to an order of magnitude greater (Warren and Aniya, 1999).

Bridge Glacier’s calving rate is also tempered by moderate water depth and flow speeds. Higher calving rates are associated with greater water depths and significantly larger terminus velocities. Large Patagonian and Icelandic glaciers have terminus velocities between 258 and 1810 ma\(^{-1}\) (Haresign, 2004), up to an order of magnitude greater than what has been measured at Bridge Glacier (140 ma\(^{-1}\)). Conversely, smaller calving glaciers in New Zealand terminate in shallow lakes (< 50 m) and many have low flow speeds (< 70 ma\(^{-1}\)).

Lake temperatures also appear to play a role in the calving rate. Many Patagonian icefields terminate in large lakes where water temperatures can be up to 7.6°C (Warren and Aniya, 1999), significantly warmer than the well-mixed 1°C water observed at Bridge Lake. This discrepancy is most likely related to the size of the lakes. Although Bridge Glacier is relatively large (6.3 km\(^2\) in 2013), it is dwarfed by the much longer lakes of Southern Patagonia, where depths of over 300 m and large areas limit the cooling influence of glacier melt.

Bridge Glacier shares similar calving characteristics with both Tasman and Mendenhall Glaciers, which have undergone significant retreat as they transitioned from grounded to floating termini (Boyce et al., 2007; Dykes et al., 2011). During this transition, terminus velocities increased at Tasman from 69 ma\(^{-1}\) to 218 ma\(^{-1}\) (Dykes and Brook, 2010; Dykes et al., 2011), while the calving rates for both sites increased from 50 ma\(^{-1}\) to between 227 and 431 ma\(^{-1}\) (Boyce et al., 2007; Dykes et al., 2011). These rates are consistent with
6.3. **Relative Importance and Long Term Implications**

Table 6.1: Characteristics of selected major lacustrine calving glacier studies with added Bridge Glacier variables. $D_w$ is the mean water depth, $T_w$ is the mean water (depth averaged or range) temperature, $U_c$ is the calving rate, and $U_T$ is the terminus averaged flow speed. Citations: a: Boyce et al. (2007), b: Motyka et al. (2003), c: Warren and Kirkbride (2003), d: Dykes et al. (2011), e: Warren and Aniya (1999), f: Stuefer et al. (2007), g: Haresign (2004), h: this study.

<table>
<thead>
<tr>
<th>Location</th>
<th>Year</th>
<th>$D_w$ (m)</th>
<th>$T_w$ ($^\circ$C)</th>
<th>$U_c$ (ma$^{-1}$)</th>
<th>$U_T$ (ma$^{-1}$)</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alaska</td>
<td>Mendenhall</td>
<td>1997-2004</td>
<td>45 - 52</td>
<td>1 - 3</td>
<td>12 - 431</td>
<td>45 - 55</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>New Zealand</td>
<td></td>
<td>1995</td>
<td>4 - 20</td>
<td>1.7 - 4.3</td>
<td>14 - 88</td>
<td>5 - 151</td>
</tr>
<tr>
<td></td>
<td>Maud, Grey, Godley, Ruth, Hooker</td>
<td>2000 - 2006</td>
<td>50</td>
<td>1 - 10</td>
<td>78</td>
<td>69</td>
</tr>
<tr>
<td></td>
<td>Tasman</td>
<td>1995</td>
<td>10</td>
<td>0.5</td>
<td>28</td>
<td>11</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2006 - 2008</td>
<td>153</td>
<td>1 - 10</td>
<td>227</td>
<td>218</td>
</tr>
<tr>
<td>Patagonia</td>
<td>Upsala West, Ameghino, Grey, Nef</td>
<td>1995</td>
<td>153 - 325</td>
<td>2 - 7</td>
<td>355 - 2020</td>
<td>370 - 1620</td>
</tr>
<tr>
<td></td>
<td>Perito Mereno</td>
<td>2001</td>
<td>65</td>
<td>4.5 - 7.0</td>
<td>520 - 1770</td>
<td>520 - 1810</td>
</tr>
<tr>
<td></td>
<td>Leon</td>
<td>2001</td>
<td>65</td>
<td>4.5 - 7.0</td>
<td>520 - 1770</td>
<td>520 - 1810</td>
</tr>
<tr>
<td>Iceiland</td>
<td>Fjallsjokull</td>
<td>2003</td>
<td>75</td>
<td>1.5 - 3.0</td>
<td>582</td>
<td>258</td>
</tr>
<tr>
<td>Canada</td>
<td>Bridge</td>
<td>2013</td>
<td>109</td>
<td>1.1 - 1.5</td>
<td>281</td>
<td>140</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1984 - 1990</td>
<td>61</td>
<td>30</td>
<td>70 - 210</td>
<td>h</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1991 - 2003</td>
<td>90</td>
<td>82 (0 - 351)</td>
<td>70 - 210</td>
<td>h</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2004 - 2012</td>
<td>102</td>
<td>237</td>
<td>70 - 210</td>
<td>h</td>
</tr>
</tbody>
</table>

Bridge Glacier.

At both Tasman and Mendenhall Glaciers, water depth and buoyancy, control the magnitude of calving (Boyce et al., 2007; Dykes et al., 2011; Dykes, 2013). These findings are corroborated by observations at Bridge Glacier, which suggests that the vast majority of the ice discharged from the terminus is triggered by buoyant forces. While the rate and magnitude of calving is greater at Bridge Glacier than in small, grounded glaciers, it is relatively modest in relation to large floating lacustrine glaciers of Patagonia and Iceland.
6.3.4 Future Implications

Since 1991, Bridge Glacier has been experiencing large tabular calving events that have resulted in an irregular retreat, defined by years of large retreat (and high calving flux), and years of relative stability. This retreat is a product of buoyant terminus conditions, which allow for large volumes of ice to be discharged in singular events as terminus crevasses connect and ultimately fail due to buoyant torque. However, as Bridge Glacier retreats further up-valley, it will terminate in shallower waters and will eventually cease to float (Figure 6.4).

Figure 6.4: Bridge Glacier from above, September 13, 2013, showing projected future grounding line.

Once the terminus retreats to shallower water and becomes grounded, large tabular calving events will no longer occur. Instead, the calving regime will transition from high magnitude, irregular events, to low magnitude, higher frequency events. Observations from this study indicate that small, high frequency ‘terminus force imbalance’ events are a relatively small source of mass loss. This transition in the character of calving suggests that future calving should become an increasingly minor portion of mass loss at Bridge Glacier.

Even when Bridge Glacier no longer experiences large calving losses characteristic of a floating terminus, it will still be responding to a legacy calving. High rates of calving, which remove large volumes of ice from the terminus, also serve to reduce glacier backstress, and ‘draw’ ice down from up-valley, leading to dynamic thinning (Naruse and Skvarca, 2000; Benn et al., 2007a). Although not yet quantified, this thinning is apparent in historical satellite images (see Figure 2.1). Dynamic thinning at Bridge Glacier would serve to
enhance and/or prolong future calving due to a smaller ice thickness, reducing the depth of water required to maintain flotation. This mechanism would serve to enhance future dynamic thinning, and also enhance surface melt induced retreat once the glacier terminus retreats to dry land, due to a thinner terminus.
Chapter 7

Conclusions

7.1 Major Findings

The primary findings from this study are summarized below:

- During the 2013 melt season, surface ablation accounts for 77% of the ice loss. Topographic and glaciological factors constrain the potential variability in calving flux, suggesting that surface melt is consistently the main driver of mass loss at Bridge Glacier.

- Historical calving and surface melt rates show that calving does not significantly contribute to mass loss before 1991. From 1991 to 2003 calving was more pronounced, while the calving flux was roughly equal to the volumetric ice loss from surface melt for 2005, 2008 and 2010. Calving contributes significant mass loss in multi-annual periods, alternating with years of low calving output.

- Surface ablation at Bridge Glacier is driven primarily by solar radiation, particularly over low albedo ice. Turbulent flux parameters are strongly affected, both spatially and in magnitude, by strong, persistent katabatic flow, and have a moderate effect on the volume, and spatial distribution of ice melt.

- Calving events over a single season are found to be dissociated with major meteorological, limnological and glaciological variables. Calving events are triggered primarily by buoyant torque, suggesting that variables that affect this force balance, such as water level and ice thickness (melt rates) affect calving more through a cumulative effect, rather than short timescale perturbations.

- As the retreat of Bridge Glacier continues, calving is expected to become an even less important contributor of mass loss, particularly once the terminus retreats into shallow water, and becomes grounded. However, the effect of calving on Bridge
Glacier’s mass balance is not limited to the calving flux. Future retreat is expected at Bridge Glacier due to a legacy of dynamic thinning brought about by its transient calving phase.

7.2 Future Work

This study has explored the mechanics and relative importance of calving and ablation at Bridge Glacier and highlights further questions about mass loss in lacustrine calving glaciers.

One of the major limitations of this study is that high resolution mass balance data is only taken from one melt season, and historical surface melt rates rely on coarse estimates. This broad investigation of historical melt rates limits the scope of the conclusions that can be reached. In order to better constrain the relative importance of calving in a floating terminus lacustrine system, more work is needed to refine estimates of long-term surface melt rates and relate them to calving losses.

Further work is also needed to better understand the transition from grounded to floating terminus. A more robust understanding, either through modelling or observation, of the drivers of calving, and how they change over these transitions will improve our ability to explain retreat rates of lacustrine calving glaciers.

An examination and quantification of the dynamic thinning at Bridge Glacier would also provide valuable insight into lacustrine glacier dynamics. The timing and rates of thinning would allow for a better separation of the role of surface melt and calving, and the interaction between the two main drivers of ice loss at Bridge Glacier.

Finally, more work is needed to compare and contrast the history of Bridge Glacier with other lacustrine systems. While this study attempts to relate observations from the study period with historical rates from other lacustrine glaciers (Section 6.3.3), these studies only represent snapshots in the life-cycle of a calving glacier. A more robust look at the commonalities between lacustrine glaciers will elucidate the main drivers of calving, and allow for a clearer understanding of the relationship between climate and calving in glacier mass balance.
Bibliography


Bibliography


Appendix A

Distributed Energy Balance Modelling Data Processing

A.1 Pressure Transducer Post-Processing

In order to obtain a high temporal resolution melt rate dataset, a pressure transducer was installed in a water-filled borehole at Glacier AWS throughout the melt season. At the time of installation, water was free-flowing in the vicinity, and the borehole was consistently refilling to the surface. However, when the borehole was re-drilled July 19, water was only filling to 25 cm from the surface.

Raw water depths taken from the pressure transducer show a marked diurnal trend of emptying and filling. The magnitude of this diurnal trend is variable, with spreads of several centimetres, up to 80 cm (see Figure A.1). This diurnal trend does not appear to be related to either air temperatures (at Glacier AWS), or modelled melt rates. The spread does decrease, however, as the pressure transducer approaches the surface. This is due most likely to the fact that there is simply less daily filling and emptying that can occur in a shallower hole.

Although there is significant observed diurnal variation in water depth, taking the maximum water depth for each day produced no physically impossible results (i.e. a negative melt rate). There was, however, significantly higher variability in the magnitude of daily melt rates relative to modelled results from both a distributed and point Energy Balance model (see Figure A.2). While the DEBM consistently predicts melt between 4 - 12 cm/day, measured results from the pressure transducer predict melt rates for some days close to 20 cm/day, substantially higher than would have otherwise been expected.

The variability observed in the pressure transducer melt rate suggests that it is difficult to discern how the glacial hydrology refills and empties the borehole. While in aggregate it produces meaningful results, higher frequency (sub-daily or daily) measurements become...
A.2. Solar Geometry Calculations

Figure A.1: Difference between maximum and minimum depth readings for each day for the pressure transducer. The two grey lines correspond to the days when the borehole was re-drilled and melted to the surface respectively.

increasingly less reliable, and seem to produce daily melt rates that are up to 10 cm different than what would have been predicted from energy balance modelling. Conversely, the pressure transducer method seems to be a reliable method for measuring lower frequency (weekly or bi-weekly) melt rates, allowing for additional valuable low-cost information that could be used to inform and constrain modelled higher resolution melt rates.

A.2 Solar Geometry Calculations

All solar geometry calculations follow classical astronomical theory, and are outlined in detail by Oke (1988). The declination for every timestep is calculated as

\[
\delta = 0.006918 - 0.399912 \cos(\gamma) + 0.070257 \sin(\gamma) - 0.006758 \cos(2\gamma) \\
+ 0.000907 \sin(2\gamma) - 0.002697 \cos(3\gamma) + 0.00148 \sin(3\gamma)
\]  

(A.1a)

where \( \gamma \) is the fractional year. The fractional year is calculated as follows:

\[
\gamma = 2\pi(DOY - 1)/365
\]  

(A.1b)
A.2. Solar Geometry Calculations

Figure A.2: Pressure transducer observed melt rate plotted against Distributed Energy Balance Model melt rate, extracted for Glacier AWS location.

and is a function of the Julian day of year (DOY).

The hour angle \( h \) is calculated as

\[
h = 15(12 - t) \tag{A.2a}
\]

where \( t \) is the fractional hour. The fractional hour is converted to local apparent time \( (LAT) \), which is calculated as

\[
LAT = LMST - 60 \times 229.18(0.000075 + 0.001863 \cos(\gamma) - 0.032077 \sin(\gamma)
\]

\[
- 0.014615 \cos(2\gamma) - 0.040849 \sin(2\gamma)) \tag{A.2b}
\]

where \( LMST \) is the local mean standard time.

The elevation angle is calculated as

\[
\cos Z = \cos \varphi \cos \delta \cos h + \sin \varphi \sin \delta \tag{A.3}
\]

and is a function of latitude \( (\varphi) \), declination \( (\delta) \), and the hour angle \( (h) \). The azimuth
angle \( (\Omega) \) is calculated as follows:

\[
\cos \Omega = (\sin \delta \cos \varphi - \cos h \cos \delta \sin \varphi)(\sin Z)^{-1}.
\]  

(A.4)

The mean and instantaneous Sun-Earth distance ratio \( \frac{R_m}{R} \) is calculated as

\[
\left( \frac{R_m}{R} \right)^2 = 1.00011 + 0.034221 \cos(\gamma) + 0.001280 \sin(\gamma) \\
+ 0.000719 \cos(2\gamma) + 0.000077 \sin(2\gamma)
\]  

(A.5)

and is a function of the fractional year.

### A.3 Modelled Data Gaps

During the summer there were a couple instances when data was not collected from one of the AWS sensors. To have a complete and continuous data record for energy balance modelling, the data gaps were infilled using several statistical routines and assumptions outlined below.

#### A.3.1 Incoming Longwave Radiation Gaps

Between June 19 and 20, as well as August 17 for 4 hours, the pyrgeometer measuring incoming longwave radiation at Lake AWS malfunctioned, most likely due to a lack of battery power, and did not collect data for several timesteps. In order to infill the data, the Stefan-Bozmann equation was used, which was adapted following Oke (1988).

\[
L_{\text{in}} = \varepsilon \sigma (T + 273.15)^4 (1 - 0.22n^2)
\]  

(A.6)

was assumed, and air temperatures were taken from the Ridge AWS. \( n \) is the fraction of cloud cover during the period. Over the 12 hours before pyrgeometer failure, the incoming shortwave radiation was approximately 50% of the potential incoming radiation, leading to an assumed 50% cloud cover during the time period.

Emissivity for the period is calculated following Idso et al. (1969) where

\[
\varepsilon_{a(0)} = 1 - 0.261[-7.704(273.15 - T)^2]
\]  

(A.7)
and $T$ is the air temperature.

### A.3.2 Incoming Wind Speed and Direction Gaps

Between July 14 and 19, Glacier AWS was tipped over due to a suspected combination of uneven melt, high winds, and a crevasse opening up under one of the legs of the tripod. During this period, wind speed and direction data were unable to be measured. Other data measured from the Glacier AWS, such as temperature and humidity, were not found to have varied in their correlations with both Ridge and Lake AWSs, suggesting those data were not measurably affected.

In order to fill wind speed data gaps, wind speed was correlated against ambient air temperature, taken from Ridge AWS. The highest correlation was found with a 1 hour lag time (6 time steps), showing an adjusted $r^2$ of 0.65, which is strongly statistically significant.

Wind direction was modelled based on a threshold difference between ambient and on-glacier air temperatures. If the difference was greater that 5°C, the wind direction was classified as downslope (255°), while if the temperature difference was smaller, wind direction was classified as upslope (90°).
Appendix B

Time Lapse Camera Methodology and Walkthrough

B.1 Getting Started

Tracker is a free Open Source modelling tool built in Java using Open Source Physics (OSP). It has a nice GUI, and allows an easy way to physically track features as they move across successive pictures. It can be downloaded on Mac OS X, Windows, and Linux from http://www.cabrillo.edu/~dbrown/tracker/. I have not used Linux, but both Mac and Windows options come with a nice installer, so all you need to do is follow their directions. Xuggle is the preferred video engine, and is automatically installed with your Tracker install (this can be disabled). Install FAQ and further documentation available at http://www.cabrillo.edu/~dbrown/tracker/installers/installer_help.html.

B.2 Importing Pictures

Tracker recognizes both videos (.mov format) and pictures. Images will have a visibly higher resolution than a video in all likelihood, due to video compression routines and are recommended. In order to import time-lapse images, they need to be named successively, without gaps. An added wrinkle in this is that the numbering must all have the same character length. For instance: nunatak_8.jpg, nunatak_9.jpg, nunatak_10.jpg will not work because 9 and 10 are not the same character length; instead, they need to be named nunatak_08.jpg, nunatak_09.jpg, nunatak_10.jpg. This naming scheme should happen automatically out of the camera, but if you are selecting only certain pictures and have ‘jumps’ in numbering, that will need to be fixed before Tracker recognizes they are successive frames.

In all likelihood, the volume and resolution of your images will require added memory.
B.2. Importing Pictures

Figure B.1: Memory in use

to properly display your images. When you get close to full, the banner in the top right corner will blink red and show you how close you are to full (see Figure B.1). When you are full, the program will have trouble displaying frames, and will display some as blank, leaving holes in your data. You can fix this by clicking on the Memory in use button (Figure B.1) and then navigating to the Runtime tab where you can manually increase the memory size (see Figure B.2).

Figure B.2: Manually increase the memory size

Choosing pictures may also end up being one of the most critical steps to getting good data:
B.2. Importing Pictures

- Pick pictures with similar lighting. This will make tracking features easier, and more accurate.

- Shoot roughly three times as many pictures as you may need. Inclement weather, such as fog, low clouds or rain/water on the lens can make pictures unusable.