Why Do Glaciers Surge? Understanding the Last Eight Surges of Donjek Glacier, Yukon, Canada

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WHY DO GLACIERS SURGE? UNDERSTANDING THE LAST EIGHT SURGES OF DONJEK GLACIER, YUKON, CANADA

By

William Kochtitzky

B.S. Dickinson College, 2016

A THESIS

Submitted in Partial Fulfillment of the
Requirements for the Degree of
Master of Science
In Earth and Climate Sciences

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Glacier surges are short periodic episodes of rapid glacier flow that are driven by internal instabilities and bracketed by longer periods of slow flow. These glaciers are important to understand because they are vital to predicting future sea level rise, mitigating glacier hazards, and understanding basal glacial processes. Donjek Glacier, located in the Yukon, Canada has an unusually short and regular surge cycle, with eight surges identified since 1935 from aerial photographs and satellite imagery with a ~12 year repeat interval and ~2 year active phase. Recent surges occurred during a period of long-term negative mass balance and cumulative terminus retreat of 2.5 km since 1874. In contrast to previous work, we find that the constriction where the valley narrows and bedrock lithology changes, 21 km up-glacier of the terminus, represents the upper limit of surging, with negligible surface speed or elevation change up-glacier from this location. This positions the entire surge-type portion of the glacier in the ablation zone. The constriction geometry does not act as the dynamic balance line, which we consistently find at 8 km up-glacier from the glacier terminus. During the 2012–2014 surge, the average lowering rate in the lowest 21 km of the glacier was 9.6 m a$^{-1}$, while during quiescence it was 1.0 m a$^{-1}$. Due to reservoir zone refilling, the ablation zone has a positive geodetic balance in
years immediately following a surge event. An active surge phase can result in a strong negative geodetic mass balance over the surge-type portion of the glacier.

Potential links between climate and glacier surges are not well understood, but are required to enable prediction of glacier surges and mitigation of associated hazards. This thesis investigates the role of snow accumulation and atmospheric temperature on surge periodicity, glacier area changes, and surge initiation since the 1930s for Donjek Glacier. Three ice cores from Eclipse Icefield, at the head of the glacier, indicate that a total accumulation of 13.1 to 17.7 m w.e. of snow occurred in the 10-12 years between each of its last eight surges. This suggests that a threshold must be passed before the initiation of a surge event, although it remains unclear whether the relationship between cumulative snowfall and surging is due to the consistency in repeat surge interval and decadal average precipitation, or if it is indeed related to surging. The 1968 to 2017 climate record from Burwash Landing tests if there is a relationship between surge periodicity and an increase of 2.5°C in mean annual air temperature over this period. No such relationship was found, although each of the past 8 surge events has been less extensive than the previous, with the maximum terminus extent approximately 8 km² smaller in the most recent 2012-2014 surge event than the ~1947 surge event.
ACKNOWLEDGEMENTS

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CHAPTER 1

INTRODUCTION

Sea level has already risen by ~22 cm since 1900 (Church et al., 2013) with devastating consequences for low lying communities around the world. Sea level is predicted to further rise between 26 cm and 98 cm by 2100, causing even more costly and deadly storm surges, flooding, and coastal erosion (Church et al., 2013). The large uncertainty in this projection primarily rests in how glacier velocities will respond to climate change. At present, it is unclear whether glaciers in a warming climate will increase their flow velocity, or even collapse, due to increased basal lubrication (leading to greater sea level rise), or slow down due to reduced ice thickness (leading to a reduced sea level rise contribution; AMAP, 2017). We cannot reliably constrain sea level rise estimates until this question is answered. This uncertainty is problematic because it reduces our ability to ensure that adaptation in communities around the world is effective and economically efficient.

In recent decades, mountain glaciers, particularly those in Alaska and Arctic Canada, have contributed more to global sea level rise than the Greenland and Antarctic ice sheets (Gardner et al., 2013; Harig and Simons, 2016; AMAP, 2017). However, future sea level rise will be dictated by the stability of outlet glaciers and ice streams in Greenland and Antarctica (Alley and Bindschadler, 2001; Bennett, 2003). If these ice streams collapse, they could lead to tens of meters of sea level rise (Alley et al., 2005). Ice streams are extremely difficult to study due to their remoteness and large size, but the same processes that govern the stability of ice streams also influences surging behavior exhibited by smaller glaciers outside the ice sheets. These surge-type glaciers, which are glaciers that have a cyclical flow instability on decadal timescales, offer a unique analogue to ice streams and have much more abundant data spanning periods of both stability and instability (Meier and Post, 1969; Clarke, 1987). Alaska and the Yukon are
home to one of the highest densities of surge-type glaciers in the world (Clarke et al., 1986; Sevestre and Benn, 2015), and offer an ideal natural laboratory for exploring the range of ice dynamic responses to climate change. Thus, understanding glacier dynamics in this region is vital for near and long-term sea level rise projections, and for the ability of these glaciers to provide an analogue for the evolution of large ice streams.

Surge-type glaciers in the Yukon and Alaska typically alternate between short-lived active phases and long-lived quiescent phases (Meier and Post, 1969). Active phases typically last 1-2 years when the ice velocity is high (typically >1000 m/yr) and the terminus rapidly advances. Quiescent phases typically last 10-40 years and are characterized by slower or stagnant ice velocity (typically <10 m/yr) and a retreating terminus (Meier and Post, 1969; Fig. 1). Active phases of surge type glaciers can be quite hazardous for surrounding communities. Donjek Glacier, located in Yukon, Canada has a long history of surging in First Nations’ oral history (Cruikshank, 1981) and scientific studies (Abe et al., 2015). When Donjek previously advanced it blocked the Donjek River at its terminus, creating a glacier-dammed lake (Collins and Clarke, 1977). When the glacier dam forms, a reservoir fills as the river is halted, causing flooding of potentially large swaths of land upstream. When the dam breaks, it can devastate communities downstream. While recent damming by Donjek Glacier has not substantially damaged infrastructure, it remains a possible hazard to the Alaska Highway, located ~50 km downstream. This thesis focuses on Donjek Glacier in Yukon, Canada, and is divided into two results chapters. The first chapter discusses the surging of Donjek Glacier since the 1930s. Through remote sensing work, I describe the eight most recent surge events that have occurred on Donjek Glacier. This includes changes in the glacier surface elevation and velocity from the quiescent to the active phase. The second chapter focuses on the connections between climate and surging at
Donjek Glacier. This chapter uses ice core analysis, weather observations, and remote sensing to quantify the impact of cumulative accumulation and temperature on surging. Ultimately, the objective of these papers is to 1) understand the recent history of Donjek Glacier surging, 2) better predict associated glacier hazards, and 3) increase our understanding of glacier instabilities and the connections to climate.

The two chapters of this thesis are currently in review (as of April, 2019) in peer reviewed scientific journals and result from the work of numerous people. The first chapter incorporates suggestions from all the co-authors including Hester Jiskoot, Luke Copland, Ellyn Enderlin, Robert McNabb, Karl Kreutz, and Brittany Main as well as two anonymous reviewers. This chapter received extra help from Hester Jiskoot including her help in writing pieces of the introduction and discussion sections of this chapter. Robert McNabb provided ASTER DEMs and did the DEM coregistration. Brittany Main provided SAR images for velocity mapping.

Chapter two represents the work of a large number of people over many years. This chapter could not have been written without the enormous efforts of several scientific teams who collected three ice cores used in this chapter. Portions of the methods section about ice core data processing were written by Dominic Winski and Erin McConnel, who also helped analyze the ice core data. The manuscript benefited from additional comments from the co-authors including Seth Campbell, Ellyn Enderlin, Luke Copland, Brittany Main, Christine Dow, and Hester Jiskoot. Hester Jiskoot also wrote portions of the literature-review based content in the introduction and discussion sections in chapter 2.
CHAPTER 2
TERMINUS ADVANCE, KINEMATICS, AND MASS REDISTRIBUTION DURING EIGHT SURGES OF DONJEK GLACIER, ST. ELIAS RANGE, CANADA, 1935 TO 2016

2.1. Abstract

Donjek Glacier has an unusually short and regular surge cycle, with eight surges identified since 1935 from aerial photographs and satellite imagery with a ~12 year repeat interval and ~2 year active phase. Recent surges occurred during a period of long-term negative mass balance and cumulative terminus retreat of 2.5 km since 1874. In contrast to previous work, we find that the constriction where the valley narrows and bedrock lithology changes, 21 km from the terminus, represents the upper limit of surging, with negligible surface speed or elevation change up-glacier from this location. This positions the entire surge-type portion of the glacier in the ablation zone. The constriction geometry does not act as the dynamic balance line, which we consistently find at 8 km from the glacier terminus. During the 2012–2014 surge, the average lowering rate in the lowest 21 km of the glacier was 9.6 m a⁻¹, while during quiescence it was 1.0 m a⁻¹. Due to reservoir zone refilling, the ablation zone has a positive geodetic balance in years immediately following a surge event. An active surge phase can result in a strongly negative geodetic mass balance over the surge-type portion of the glacier.

2.2. Introduction

Glacier surges are short periodic episodes of rapid glacier flow that are driven by internal instabilities and bracketed by longer periods of slow flow (Meier and Post, 1969). While much research has been focused on understanding surging mechanisms (Meier and Post, 1969; Raymond, 1987; Harrison and Post, 2003; Qiu, 2017), surge dynamics are not yet fully understood, partially due to a lack of repeat observations of multiple surge events for the same
glacier. In addition, surging behavior varies widely between glaciers, and even for repeat surges of the same glacier (Harrison and others, 1994; Björnsson and others, 2003).

Although less than 1% of glaciers worldwide exhibit surge behavior (Sevestre and Benn, 2015), Alaska and western Canada are home to 113 confirmed surge-type glaciers, the third highest number after Svalbard and the Pamirs (Post, 1969; Clarke and others, 1986; Kotlyakov and others, 2010; Sevestre and Benn, 2015). In the Canadian portion of the St. Elias Mountains, approximately 6.4% of 2356 glaciers are of the surge-type (Clarke and others, 1986). The Yukon hosts surge-type glaciers with a wide variety of dynamic characteristics, including surge phases of up to several years and 12–50+ year repeat intervals (e.g., Donjek and Lowell Glaciers: Abe et al., 2016; Bevington and Copland, 2014; Steele Glacier: Clarke and others, 1986) and slow surges with surge phases that last several decades and surge speeds $< 50$ m a$^{-1}$ (e.g., Trapridge Glacier: Clarke and Blake, 1991; Unnamed Glacier: De Paoli and Flowers, 2009). Many of the glaciers in this region are classified as polythermal (Jarvis and Clarke, 1975) and overlay soft basal sediments (Clarke and others, 1986; Harrison and Post, 2003; Crompton and Flowers, 2016; Crompton and others, 2018). Both these aspects have been suggested to be conducive to surge behavior (Hamilton and Dowdeswell, 1996; Jiskoot and others, 2000; Truffer and others, 2000). Glaciers with polythermal regimes typically exhibit slower surge development with lower peak velocities and longer surge intervals than their temperate counterparts (Clarke and Collins, 1984; Murray and others, 2003; Frappé and Clarke, 2007). Therefore, fewer repeat surge events have been observed for polythermal glaciers, limiting the understanding of surge mechanisms and initiation triggers (Murray and others, 2003; Bevington and Copland, 2014), structural development, speed-up magnitudes and advance patterns (Jiskoot and others, 2001; King and others, 2015; Quincey and others, 2015; Herreid and Truffer, 2016), glacial land system
development (Schomacker and others, 2014) and climate controls on surging (Eisen and others, 2001; Hewitt, 2007; Jiskoot and Juhlin, 2009; Flowers and others, 2011).

Surges are thought to initiate when a buildup of ice in a reservoir zone steepens the local surface slope at the dynamic balance line, or the location through which mass moves to the receiving zone but experiences no net elevation change (Dolgoushin and Osipova, 1975), until it reaches a critical basal shear stress. When the slope-steepening increases the gravitational driving stress above the critical basal shear stress, the subglacial hydrologic system rapidly evolves and surging occurs (Meier and Post, 1969; Raymond, 1987). In the classic surge cycle a surge will propagate down glacier from a reservoir zone where mass has built-up (Meier and Post, 1969). This type of surging has been observed at several surge-type glaciers in Yukon-Alaska, including Bering (Roush and others, 2003), Trapridge (Clarke and others, 1984; Frappé and Clarke, 2007) and Variegated (Kamb and others, 1985) Glaciers, as well various glaciers in Svalbard (Murray and others, 1998; Murray and others, 2000; Dowdeswell and Benham, 2003; Sund and others, 2009; Mansell and others, 2012), East Greenland (Jiskoot and Juhlin, 2009) and the Karakoram (Quincey and others, 2015). However, some tidewater glacier surges in Svalbard have been observed to propagate up-glacier (Rolstad and others, 1997; Luckman and others, 2002; Murray and others, 2003; Dowdeswell and Benham, 2003; Murray and others, 2012; Dunse and others, 2015; Flink and others, 2015; Sevestre and others, 2018). Up-glacier surge propagation has been observed elsewhere as well, including at Sabche Glacier, Nepal (Lovell and others, 2018). Both up- and down- glacier surge propagation have also been observed, such as at Sortebræ in Greenland (Murray and others, 2002).

Mass redistribution through increased ice flow is a key characteristic of surging (Meier and Post, 1969). Surge events cause a short-term redistribution of mass from the reservoir zone,
across the dynamic balance line (Dolgoushin and Osipova, 1975), into the receiving zone (Meier and Post, 1969). Long-term changes in climate and associated glacier mass balance can cause glaciers to either become or cease to be of surge-type, or alter the number of surges within a region (Dowdeswell and others, 1995; Copland and others, 2011) or the individual surge recurrence interval (Eisen and others, 2001). As such, it is critical that the mass balance of glaciers prior to and during surge events, as well as the location of the dynamic balance line with respect to the equilibrium line altitude (ELA), are well understood.

To better understand mechanisms of and controls on glacier surging, we reconstruct all surge events of Donjek Glacier from 1935 to present using aerial photography and satellite image archives. The primary goal of this analysis is to test the hypothesis by Abe and others (2016) that a valley constriction ~20 km from the terminus controls the surging of the glacier by causing ice to locally thicken. For this purpose, we measure long-term changes in terminus position, surface velocity, ice elevation and surface slope, and temporally constrain the velocity patterns before, during, and after surge events in 2000–2002 and 2012–2014. These measurements provide information concerning the drivers of surge initiation and termination and enable the quantification of mass movement during surge events. Finally, we compare the surge kinematics, including the dynamic balance line location, reservoir and receiving zone length, and elevation change, to other glaciers around the world.
2.3. Study Region

Donjek Glacier (61°11'14" N, 139°31'30" W; Fig. 1) is located in the St. Elias Mountains of southwest Yukon. In 2010, the glacier was 65 km long and had an area of 448 km² (RGI Consortium, 2017). Using a WorldView digital elevation model (DEM) from 2013 and the RGI outline, we find that the glacier surface elevations range from ~1000 m a.s.l. at the terminus to 4507 m a.s.l at the peak of Mount Walsh. In the 21st century the late-summer snowline has gradually increased in elevation from ~2430 m a.s.l in 2004 to ~2550 m a.s.l. in 2017, as measured in repeat Landsat imagery. Larsen and others (2015) found that Donjek Glacier had an area-averaged negative mass balance of -0.29 m w.e. a⁻¹, or -0.13 Gt a⁻¹, between May 2000 and May 2012.
Denton and Stuiver (1966) used C\textsuperscript{14} age dating to determine that Donjek Glacier receded from Kluane Lake (35 km to the northeast) ~9800 years ago and its youngest major neoglacial advance occurred pre-1874 (Fig. 2). Previous research on Donjek Glacier has recorded six surge events since 1935, namely in 1935, 1978, 1969, 1989, 2001, and 2013 (Denton and Stuiver, 1966; Johnson and others, 1972a; Clarke and Holdsworth, 2002; Abe and others, 2016). Since at least 1874 Donjek Glacier has periodically dammed the Donjek River during surge events, each time culminating in a post-surge outburst flood that endangers downstream infrastructure along the Alaska Highway (Clarke and Mathews, 1981). The most recent surge of Donjek, in 2013,
caused the terminus to increase in area by nearly 2 km² and move at a rate of ~3 m d⁻¹ over the lowest 5 km (Abe and others, 2016).

Table 1. Aerial photographs and images used for terminus delineation in this study

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*UAF = University of Alaska Fairbanks Elmer E. Rasmuson Library; Yukon EMR = Yukon Energy, Mines, and Resources Library, Whitehorse; NAPL = National Air Photo Library, Ottawa; ASF = Alaska Satellite Facility; EE = United States Geological Survey Earth Explorer; PGC = Polar Geospatial Center at University of Minnesota

2.4. Methods

2.4.1. Maximum surge terminus position mapping

We used aerial photographs and satellite imagery to reconstruct the timing of surge events since 1937 and to digitize past terminus extents (Table 1). Terminus positions were manually mapped from georeferenced images in a geographic information system. Vertical aerial photographs from 1947 were originally acquired by the Royal Canadian Air Force and obtained from the Canadian
National Air Photo Library (Ottawa) and Yukon Energy, Mines, and Resources Library (Whitehorse). Vertical and oblique photos from the collections of Bradford Washburn and Austin Post provided coverage from the 1930s and 1960s, respectively (Table 1). We used two WorldView-2 images (Table 1) to manually georectify vertical aerial photographs with at least eight tie points each. The root mean square error of the georectified images ranged from 17.2–114 m. We used 99 Landsat 2-8 images from Earth Explorer (https://earthexplorer.usgs.gov) to map terminus position change during surge events from 1977–1979, 1988–1990, 2000–2002, and 2012–2014 surge events. Four scenes of the entire Landsat record were used to delinteate maximum terminus position after the four most recent surge events (Table 1). We conservatively estimate our maximum uncertainty to be 2 pixel lengths or 120 m for Landsat 2, 60 m for Landsat 5, and 30 m for Landsat 7 and 8.

Table 2. Elevation data sources for ice surface change

<table>
<thead>
<tr>
<th>Source</th>
<th>Date</th>
<th>Vertical uncertainty</th>
</tr>
</thead>
<tbody>
<tr>
<td>Operation IceBridge, airborne lidar</td>
<td>30/05/2000</td>
<td>&lt;10 cm</td>
</tr>
<tr>
<td></td>
<td>22/05/2012</td>
<td></td>
</tr>
<tr>
<td></td>
<td>15/05/2016</td>
<td></td>
</tr>
<tr>
<td>ASTER (satellite)</td>
<td>28/09/2001</td>
<td>10 m</td>
</tr>
<tr>
<td></td>
<td>26/05/2002</td>
<td></td>
</tr>
<tr>
<td></td>
<td>01/08/2003</td>
<td></td>
</tr>
<tr>
<td>PGC/WorldView (satellite)</td>
<td>10/08/2013</td>
<td>~0.2 m</td>
</tr>
<tr>
<td></td>
<td>27/09/2013</td>
<td></td>
</tr>
<tr>
<td>SPOT (satellite)</td>
<td>13/09/2007</td>
<td>6 m</td>
</tr>
</tbody>
</table>

2.4.2. Elevation extraction and elevation change mapping

We created or obtained digital elevation models (DEMs) for 2000–2017 from Operation IceBridge (OIB) LiDAR swaths, Satellite Pour l’Observation de la Terre 5 (SPOT-5), WorldView, and Advanced Spaceborne Thermal Emissions and Reflection Radiometer (ASTER; Table 2). The LiDAR tracks from 2000, 2012 and 2016 were downloaded from the National
Snow and Ice Data Center (https://nsidc.org/icebridge/portal), and down-sampled to 8 m resolution for comparison with other datasets. The 2000 OIB fixed LiDAR flight line deviated slightly from the glacier centerline, so we used the location of this line for comparisons with other swath LiDAR datasets (Fig. 1). The 2000 OIB LiDAR flight made elevation measurements every ~1.75 m beneath the flight path. The 2012 and 2016 OIB datasets cover a swath width of ~500 m perpendicular to the flight path with a spatial resolution of one point per ~0.4 m². All LiDAR swaths were down-sampled to 8 m resolution for comparison with other datasets. We downloaded one SPOT 5 DEM from the SPIRIT Project (https://theia-landsat.cnes.fr) with an uncertainty of ± 6 m (Korona and others, 2009) from 13 September 2007 at 40 m spatial resolution. We obtained 2 m-resolution DEMs derived from WorldView stereo pairs from the University of Minnesota Polar Geospatial Center (PGC), with an estimated 0.2 m vertical accuracy (Shean and others, 2016). We mosaicked the individual WorldView DEMs from 10 August and 27 September 2013 (hereby referred to as the August/September 2013 DEM, down-sampled to 8 m spatial resolution), captured during the 2012–2014 surge event, to create a more spatially extensive DEM of the glacier. These DEM strips do not overlap; thus, we are unable to quantify the potential aliasing of melt and/or glacier flow on the accuracy of these measurements. Finally, we made DEMs from ASTER imagery using the software package MMASTER at 10 m vertical uncertainty and 30 m horizontal resolution for 2001, 2002, and 2003 (Girod and others, 2017). We vertically co-registered SPOT and ASTER DEMs to the WorldView DEMs using overlapping bedrock elevations. All DEMs were then smoothed using a 300 m moving window to better visualize the data without removing large-scale patterns. We assumed that pixels uncertainties are independent (i.e., random error) such that area-averaged elevation uncertainties were calculated as $1.96\sigma\sqrt{\frac{A}{n}}$, where $\sigma$ is the standard deviation for on ice
measurements, A is the area of coverage, and n is the number of pixels (Howat and others, 2008). We calculated volume change by multiplying the profile average elevation change by the glacier zone area.

2.4.3 Ice velocity mapping

We used Landsat 7 and 8 panchromatic images (Band 8) from Earth Explorer and Radarsat-2 amplitude images to measure glacier velocity (Table 3). After manually inspecting optical images for cloud cover, we used the Ames Stereo Pipeline to cross correlate panchromatic scenes to measure ice displacement (Shean and others, 2016) using the Extreme Science and Engineering Discovery Environment computing resources (Towns et al., 2014). We used a custom Python script (Derek Mueller, pers. comm.) to convert RADARSAT-2 ultrafine wide (~2.2 m resolution) level-0 raw data to amplitude images. We then processed image pairs acquired in subsequent 24-day orbits using the AMES stereo pipeline (Shean et al., 2016). Landsat 1, 2, and 5 scenes were visually assessed using manual feature tracking (e.g. on looped moraines) to observe glacier velocity increase (decrease) at the start (end) of a surge event as automated velocity mapping yield results with high uncertainties. We correlated every image without major cloud cover for each year, and smoothed the resulting velocity profiles using a moving window of 300 m (Fig. 1). To assess the precision of our measurements, we measured the apparent motion over non-glacierized terrain (i.e. static objects) for each velocity map. First, we discarded values above 1800 m a\(^{-1}\), which we considered indicative of false matches (e.g. due to variations in snow cover), as these exceeded the highest glacier motion measured during the study period. We then computed the mean apparent velocity of the remaining area surrounding Donjek Glacier that is not currently glacierized to obtain uncertainty estimates ranging from 17.0–155.1 m a\(^{-1}\) (Table 3).
Table 3. Landsat (L) 5, 7, 8, and Radarsat-2 (R2) scenes used to determine glacier velocities with associated uncertainty measured by movement of surrounding non-glacierized terrain. Date format is dd/mm/yyyy.

<table>
<thead>
<tr>
<th>Satellite</th>
<th>First scene</th>
<th>Second scene</th>
<th>Mean/STD velocity over non-glacierized terrain (m a(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>L5</td>
<td>04/05/1988</td>
<td>21/06/1988</td>
<td>81.7/200.0</td>
</tr>
<tr>
<td>L5</td>
<td>21/06/1988</td>
<td>23/07/1988</td>
<td>65.8/137.1</td>
</tr>
<tr>
<td>L5</td>
<td>08/06/1989</td>
<td>11/08/1989</td>
<td>76.3/230.7</td>
</tr>
<tr>
<td>L5</td>
<td>08/11/1989</td>
<td>28/09/1989</td>
<td>92.2/262.2</td>
</tr>
<tr>
<td>L5</td>
<td>13/07/1990</td>
<td>14/08/1990</td>
<td>45.7/150.3</td>
</tr>
<tr>
<td>L5</td>
<td>23/07/1991</td>
<td>25/09/1991</td>
<td>42.0/176.4</td>
</tr>
<tr>
<td>L7</td>
<td>05/07/1999</td>
<td>29/08/1999</td>
<td>43.9/89.3</td>
</tr>
<tr>
<td>L7</td>
<td>17/08/2000</td>
<td>18/09/2000</td>
<td>67.3/129.0</td>
</tr>
<tr>
<td>L7</td>
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<td>28/04/2001</td>
<td>73.1/134.6</td>
</tr>
<tr>
<td>L7</td>
<td>16/05/2001</td>
<td>19/07/2001</td>
<td>53.7/116.2</td>
</tr>
<tr>
<td>L7</td>
<td>19/07/2001</td>
<td>20/08/2001</td>
<td>68.0/105.5</td>
</tr>
<tr>
<td>L7</td>
<td>16/03/2002</td>
<td>03/05/2002</td>
<td>62.1/106.7</td>
</tr>
<tr>
<td>L7</td>
<td>03/05/2002</td>
<td>26/05/2002</td>
<td>155.1/232.4</td>
</tr>
<tr>
<td>L7</td>
<td>26/05/2002</td>
<td>05/08/2002</td>
<td>49.0/90.9</td>
</tr>
<tr>
<td>L7</td>
<td>05/08/2002</td>
<td>22/09/2002</td>
<td>48.5/103.0</td>
</tr>
<tr>
<td>R2</td>
<td>24/02/2012</td>
<td>19/03/2012</td>
<td>62.4/51.7</td>
</tr>
<tr>
<td>L7</td>
<td>18/03/2012</td>
<td>19/04/2012</td>
<td>90.1/146.1</td>
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<td>06/06/2012</td>
<td>67.8/163.0</td>
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<td>L7</td>
<td>06/06/2012</td>
<td>08/07/2012</td>
<td>63.8/134.3</td>
</tr>
</tbody>
</table>
Table 3 continued

L7  21/03/2013  22/04/2013  124.7/170.2
L8  02/04/2013  23/05/2013  44.9/102.0
L8  23/05/2013  24/06/2013  48.2/126.0
L8  24/06/2013  11/08/2013  21.9/65.9
L8  11/08/2013  27/08/2013  59.2/156.6
R2  20/02/2014  16/03/2014  28.4/59.7
L8  25/03/2014  10/04/2014  61.7/130.6
L8  10/04/2014  26/04/2014  125.8/200.1
L8  26/04/2014  04/06/2014  54.5/108.8
L8  04/06/2014  22/07/2014  31.7/89.1
L8  22/07/2014  23/08/2014  17.0/60.9
L8  23/08/2014  15/09/2014  64.7/127.1

2.5. Results

2.5.1. Surge initiation, termination, and timing

Surge initiation can be measured by terminus advance, velocity increase, crevasse formation and/or surface elevation change (Meier and Post, 1969; Raymond, 1987; Sund and others, 2009) each with a different date and uncertainty range. We quantified the surge initiation of Donjek Glacier by the first record of terminus advance or up-glacier velocity increase. Similarly, surge termination was defined by the first terminus retreat or velocity decrease. We consider the terminus to have advanced (retreated) when the glacier area increased (decreased) outside the terminus delineation uncertainty. We report uncertainties in surge timing based on velocity and terminus position change where the uncertainty in active or quiescent phase length
is due to satellite image availability. Based on our analysis of historical aerial photography and satellite imagery, we independently confirmed that surge events took place during the years ~1935 (Denton and Stuiver, 1966), ~late-1950s, ~1969, 1978 (Clarke and Holdsworth, 2002), 1989, 2001, and 2013 (Abe and others, 2016).

The 1937 oblique photograph from Bradford Washburn shows glacial morphology consistent with a recent surge, independently confirming an observation of a circa-1935 surge event by Johnson and others (1972a). We additionally observed geomorphology consistent with an active surge phase in 1947 aerial photographs from the Royal Canadian Air Force (e.g. debris-free, advancing margin and small push moraines), suggesting that a surge initiated sometime within 2 years prior to 1947. While Johnson (1972a) states there was a surge in 1961, examination of glacial geomorphic features (e.g., push moraines) in Austin Post’s oblique photographs from 1961 suggest that a surge event had recently occurred but that the glacier was not actively surging at that time.

The 1977–1979 surge event was characterized by an increase in glacier velocity, followed by an active phase of ~2–3 years before the velocity decreased and terminus retreated. The surge initiated when the glacier up-glacier velocity increased between 11 June and 28 July 1977 in the lower 21 km of the glacier (Fig. 3, Table 2). The terminus then began to advance between 28 July and 17 August 1977 (Fig. 3). The terminus began to retreat and the ice velocity returned to quiescent values between 28 August 1979 and 12 May 1980 (Fig. 3; Table 2). The active phase of the 1977–1979 surge therefore lasted 2.08–2.92 years.
The 1988–1990 surge event started with terminus advance followed by velocity increase up-glacier and an active phase of ~2 years before the velocity decreased and finally the terminus retreated. The event initiated when the terminus began to advance between 6 and 23 August 1988 (Fig. 3; Table 4). The up-glacier velocity began to increase following terminus advance, between 9 September 1988 and 8 June 1989 (Fig. 3). We measured a peak velocity of $850 \pm 76$ m a$^{-1}$ between 8 June 1989 and 11 August 1989; however, Landsat images are not available to provide more temporally constrained velocity observations, so we are not confident this was the maximum velocity for the surge event. Velocity measurements from 1988–1990 show that the surge event was limited to a velocity increase in the lower 21 km of the glacier. The velocity data suggest the glacier had returned to its quiescent phase between 29 July and 14 August 1990 and the terminus began to retreat between 6 August and 1 October 1990 (Fig. 3; Table 4). The active phase of the 1988–1990 surge therefore lasted 1.97–2.15 years.
Figure 4. Surge velocity. (a-c) Active surge phase velocity for the east, middle, and west flowlines (locations on Fig. 1) during the 2000-2002 surge event. Dates for a-c are below c. (d-f) Active surge phase velocity for the east, middle, and west flowline for the 2012-2014 surge event. Dates for d-f are below f. Velocity profiles are shown from blue (beginning of surge) to red (end of surge). Extent of the reservoir zone indicated by black dashed lines at 8 and 21 km from the terminus. Dates presented in YYYYMMDD format.

Table 4. Date of transition between active (A) and quiescent (Q) phases for Donjek Glacier, based on changes in terminus position and velocity. Date format is dd/mm/yyyy. Uncertainties in earlier years due to image availability.

<table>
<thead>
<tr>
<th>Date of phase transition</th>
<th>Active/quiescent phase and duration</th>
</tr>
</thead>
<tbody>
<tr>
<td>Q₀</td>
<td>~1935</td>
</tr>
<tr>
<td></td>
<td>A₁ (~2 years)</td>
</tr>
<tr>
<td>~1935</td>
<td></td>
</tr>
<tr>
<td>Q₁ (~10 years)</td>
<td>~1947</td>
</tr>
<tr>
<td>Table 4 continued</td>
<td>A₂ (~2 years)</td>
</tr>
<tr>
<td>-------------------</td>
<td>----------------</td>
</tr>
<tr>
<td>~1947</td>
<td>Q₂ (~10 years)</td>
</tr>
<tr>
<td>Late-1950s</td>
<td>A₃ (~2 years)</td>
</tr>
<tr>
<td>Late-1950s</td>
<td>Q₃ (~10 years)</td>
</tr>
<tr>
<td>~1969</td>
<td>A₄ (~2 years)</td>
</tr>
<tr>
<td>~1969</td>
<td>Q₄ (~7 years)</td>
</tr>
<tr>
<td>11/06/1977-28/07/1977</td>
<td>A₅ (2.08-2.92 years)</td>
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<tr>
<td>28/08/1979-12/05/1980</td>
<td>Q₅ (8.24-8.99 years)</td>
</tr>
<tr>
<td>06/08/1988-23/08/1988</td>
<td>A₆ (1.97-2.15 years)</td>
</tr>
<tr>
<td>14/08/1990-01/10/1990</td>
<td>Q₆ (9.86-10.11 years)</td>
</tr>
<tr>
<td>01/08/2000-18/09/2000</td>
<td>A₇ (1.87-2.14 years)</td>
</tr>
<tr>
<td>05/08/2002-22/09/2002</td>
<td>Q₇ (9.71-10.17 years)</td>
</tr>
<tr>
<td>06/06/2012-08/07/2012</td>
<td>A₈ (1.28-1.69 years)</td>
</tr>
</tbody>
</table>
The 2000–2002 surge event began with terminus advance followed by an increase in up-glacier velocities and an active phase of ~2 years. The surge ended when the terminus began to retreat then the glacier velocities returned to quiescence. The active phase started with a small advance of the glacier terminus along the middle and eastern segments between 1 August and 18 September 2000 (Fig. 3; Table 4). We first observed up-glacier velocity increase between 18 September 2000 and 28 April 2001 (Figs. 3 and 4), when the velocity approximately doubled over the entire ~15 km up-glacier from the terminus (Fig. 4). Velocities then increased by an order of magnitude in the lowest 18 km from 16 May to 19 July 2001 compared to quiescence (Figs. 3 and 4). However, velocities did not peak until 19 July to 20 August 2001 (Figs. 3 and 4). The 2001 surge termination started with a decrease in ice motion between 21–28 September 2001 (Fig. 3). The decrease in velocity was particularly pronounced 8–10 km from the terminus, where velocities decreased by as much as 50% along the middle flowline. We then observed terminus retreat between 14 October 2001 and 23 March 2002 (Fig. 3). However, velocities did not return to quiescent values (<100 m a$^{-1}$ at the terminus) until 5 August to 22 September 2002 (Figs. 3 and 4; Table 4). The active phase of the 2000–2002 surge event therefore lasted 1.87–2.14 years (Table 4).

The 2012–2014 surge event initiated when up-glacier glacier velocities increased followed by terminus advance and an active phase of ~1.5 years. This was followed by a decrease in glacier velocities and finally terminus retreat to mark full surge termination. The first indicator of the 2012–2014 surge event occurred between 6 June and 8 July 2012 (Figs. 3, 4, and 5; Table 4) as velocity increased by as much as ~200% within the entire zone up to 15 km from
the terminus. The terminus began to advance between 8 July 2012 to 21 March 2013 (Figs. 3 and 4). However, we did not measure velocities an order of magnitude greater than quiescence until 21 March to 22 April 2013 (Fig. 3). Velocities peaked during the 2012–2014 surge event between 23 May and 24 June 2013 (Figs. 3, 4, and 5). We first observed a decrease in velocity between 24 June and 11 August 2013 in the lowest 21 km of Donjek (Fig. 3). The last period of observed velocities at least an order of magnitude above quiescent velocities was 11 August to 27 August 2013 (Fig. 3). Over the next year, velocity decreased until it returned to a quiescent rate between 22 July to 23 August 2014 (Figs. 3–5). We first observed terminus retreat between 16 October 2013 and 12 February 2014 (Fig. 3; Table 4). The active phase of the 2012–2014 surge therefore lasted 1.28–1.69 years (Table 4).

In contrast with previous studies (Clarke and Holdsworth, 2002; Abe and others, 2016) of Donjek Glacier we find that the active surge phase duration is between 1.28 and 2.15 years, and the quiescent phase duration between 9.0 and 10.17 years, for the well-defined events since the 1970s (Table 4). The surge recurrence interval was ~12 years between the first four observed surges (~1935, ~1947, late-1950s, and ~1969), ~10 years between the following three (~1969, 1977–1979, and 1988–1990) and again ~12 years between each of the last three surge events (1988–1990, 2000–2002, and 2012–2014).
Figure 5. 2012-2014 surge transects. (a-f) Velocity transects before, during, and after surge event are shown, position of each transect shown in figure 1, all velocities flow into page. Line color transitions from dark blue to dark red as the surge progresses.

2.5.2. Changes in maximum surge extent

The terminus of Donjek Glacier retreated by ~2.5 km between the pre-1874 Little Ice Age terminal moraines and its most recently advanced surge extent in May 2014. The rate of retreat has varied across the glacier terminus, with the fastest retreat rate of 18 m a\(^{-1}\) over 140 years occurring along the northwest terminus (Fig. 2). Between the maximum extent of the 1947 and 2014 surges, Donjek retreated by 1.5 km at the northwestern terminus, at a mean rate of 22 m a\(^{-1}\). The eastern terminus of Donjek has changed less through time, retreating at most 490 m between 1947 and 2014 (Fig. 2a). Between the maximum extents of two most recent surges, in 2002 and 2014, the entire terminus retreated by an average of 32 m a\(^{-1}\). In general, the magnitude of terminus advance during surges has progressively decreased over time. An exception to this
pattern is the 1978 surge, during which the terminus advanced up to 200 m further than the 1969 surge.

The only tributary of Donjek Glacier that is known to surge enters the eastern side of the main valley at a distance ~23 km upstream of the terminus (Fig. 2b). We confirm two surges of the tributary in 1974 and 2010 from looped moraines in the main trunk of Donjek Glacier in Landsat imagery (Clarke and Holdsworth, 2002; Abe and others, 2016). From 1973 to 1975 the tributary advanced 640 m into the trunk of Donjek Glacier (Fig. 2b), in 2010 it had advanced 220 m into the trunk and by 2011 it had advanced an additional 270 m into the trunk. The ice then flowed towards the terminus at an average rate of 50 m a\(^{-1}\) (Fig. 2b).

2.5.3. **Velocity patterns**

The highest surge velocities always occurred in the lower 10 km of Donjek Glacier, with increases up to two orders of magnitude from quiescence, while upstream of 15 km only a doubling or tripling of surge velocity relative to quiescence was observed (Figs. 4 and 5). In 2001 a maximum velocity of 1700 m a\(^{-1}\) occurred within 3 km of the terminus and in 2013 a maximum of 1150 m a\(^{-1}\) occurred 8 km upstream of the terminus (Fig. 4). The 2000–2002 surge velocities were generally higher than the 2012–2014 surge velocities, and near the western end of the terminus they were almost double those in 2012–2014 (Fig. 4).

Following surge initiation, velocities begin to rapidly increase below 16 km from the terminus (Fig. 4a-f). When a surge terminated and the velocity in the lower 21 km of the glacier returned to that of quiescence (average ~130 m a\(^{-1}\)), we observed velocities between 21–30 km from the terminus to remain constant or slightly increase when compared to the surge phase (Fig. 4). We were unable to reliably measure velocities >~30 km from the terminus due to transient snow cover, particularly in the accumulation zone >40 km from the terminus. Velocities were
elevated across the width of the glacier during surges (Fig. 5). Some profiles in the lower 13 km of the glacier appeared to have a slight parabolic shape with faster flow in the middle, but above 16 km velocity cross-sections showed nearly constant ice motion across the glacier width (Fig. 5). At 21 km from the terminus, velocities across the glacier width remained fairly constant throughout the surge event at ~300 m a\(^{-1}\) (ranging from ~200 to ~600 m a\(^{-1}\); Fig. 5f).

2.5.4. **Elevation and slope changes**

During the 2000–2002 and 2012–2014 surges of Donjek Glacier, changes in surface elevation suggest a net movement of ice mass from a 13 km-long reservoir zone (between 8 and 21 km from the terminus, representing an area of 28.6 km\(^2\)) to an 8 km-long receiving zone (including the advanced terminus; area of 16.0 km\(^2\); Fig. 6). During the second half of the 2000–2002 surge, we observe a redistributed of ice towards the terminus between 28 September 2001 and 26 May 2002. During this period, the surface elevation of the reservoir zone (11 and 16 km from the terminus) lowered by an average of 1.33 ± 2.1 m and locally as much as 37 ± 14 m, while the surface elevation of the receiving zone rose by an average of 1.8 ± 0.64 m and locally as much as 52 ± 14 m (Fig. 6). This does not represent the total mass displaced by the surge event, as the surge initiated before our first elevation observation in 2001 (Fig. 6).
During the 2002–2007 quiescent phase, we observed a thickening of $14.8 \pm 1.0$ m ($0.423 \pm 0.0014$ km$^3$) in the reservoir zone and a thinning of $20 \pm 1.7$ m ($0.32 \pm 0.0013$ km$^3$) in the receiving zone (Figs. 7a-c and 8). From 2007 to 2013, which encompasses ~5 years of quiescence and most of a surge event, the reservoir zone lowered by an average of $21.3 \pm 0.44$ m ($0.609 \pm 0.00060$ km$^3$) and the receiving zone thickened by an average of $5.8 \pm 0.72$ m ($0.092 \pm 0.00057$ km$^3$; Figs. 7a-c and 8). Although the receiving zone thickening does include some terminus retreat from 2007 to 2013 along the western flowline (Figs. 7 and 8). Surface elevation changes were more subtle up-glacier from the reservoir zone: from 21–30 km, the ice surface
lowered by an average of $3.5 \pm 0.37$ m from 2007 to 2013 (Fig 7). Drainage of the reservoir zone and filling of the receiving zone were not uniform across the glacier during this time period. From 2007 to 2013, we observed a maximum ice elevation increase of $44 \pm 6$ m at the terminus along the middle flowline and a maximum lowering of $61 \pm 6$ m at 12.2 km up-glacier along the west flowline.

Figure 7. 2013 surge slope and elevation change. (a-c) Elevation change for each flowline from 30 May 2002 to 24 May 2012 (red) and 24 May 2012 to August/September 2013 mosaic (blue). (d) Absolute slope of the middle profile for 30 May 2002 (dark blue), 24 May 2012 (medium blue), and the August/September 2013 mosaic (light blue). (e) Slope difference for middle flowline on 30 May 2000 to 24 May 2012 (red) and 24 May 2012 to the August/September 2013 mosaic (blue). Extent of the reservoir zone indicated by black dashed lines at 8 (dynamic balance line) and 21 km from the terminus. Note that elevation and slope change have been smoothed with a 0.3 km moving window.

As the mass moved during the surge event, the glacier surface slope also changed. Slope changes during quiescence were mirrored by a nearly equal and opposite slope change during the surge phase in a wave like pattern (Fig. 7e). The wavelength is 1.5 to 2 km with a maximum
amplitude of 2° (Fig. 7 d-e). In the lower 1 km of the reservoir zone (8 to 9 km from the terminus) the glacier surface slope increased from 1.1° to 2.2° between 2002 and 2007 (Fig. 7d-e). From 2007 to the end of the surge event in 2013, the slope over this area decreased by 1.2° on average. Overall, between 2007–2013 the glacier absolute slope decreased from 2.30° to 2.15° in the receiving zone and 2.05° to 1.8° in the reservoir zone (Fig. 7). The largest glacier surface slope changes occurred at 8.5 km from the glacier terminus on the eastern flowline, where the slope increased by 2.5° from 2002 to 2007 and decreased by 2.7° from 2007 to 2013. The observed change in surface slope along the eastern flowline corresponds with a maximum elevation gain of 51 ±12 m during quiescence along the same flowline (Fig. 7c).

Figure 8. Absolute elevation change. (a) Elevation change from 13 September 2007 to August/September 2013 mosaic from red (negative) to blue (positive). Image is Landsat 8 scene...

We measured the long-term geodetic mass balance of the lower 32 km of Donjek Glacier and the impact of the surge events using Operation IceBridge lidar swaths along the approximate centerline and WorldView DEMs. From 30 May 2000 to 22 May 2012, during which time a complete surge cycle occurred, the lower 32 km of the glacier surface lowered by an average of
1.0 m a\(^{-1}\) (Fig. 9). During the surge, from 22 May 2012 to August/September 2013 the lower 21 km of the glacier surface lowered by an average of 9.6 m a\(^{-1}\) (Fig. 9), divided between a thickening of 6.11 m (~0.098 ± 0.0000056 km\(^3\)) in the receiving zone (0–8 km from terminus) and a thinning of -21.32 m (~0.61 ± 0.0000059 km\(^3\)) in the reservoir zone (8–21 km from terminus), for a net volume change of -0.512 ± 0.0000082 km\(^3\) (calculated from profile average elevation change multiplied by zone area). From August/September 2013 to 15 May 2016, the first three years of the current quiescent phase, the lower 32 km of the glacier surface rose by an average of 1.2 ± 0.0062 m a\(^{-1}\) (Fig. 9).

From these observations, we infer the dynamic balance line to be 8 km upstream of the terminus. There is little change in surface elevation over time at this hinge, but there is pronounced mass movement across this line from the reservoir zone to the receiving zone during both the 2000–2002 and 2012–2014 surge events. Consistent with an inferred dynamic balance line in this location, there are significant changes in slope during surges in this area, especially along the eastern flowline (Fig. 7).

2.6. Discussion

2.6.1. Mass redistribution

Abe and others (2016) hypothesized that the narrowing of the valley at 20 km up-glacier of the Donjek terminus acts as a constriction to outflow, allowing ice to accumulate in an upstream reservoir zone. The thickening and steepening at this location would eventually lead to the onset of a surge. Therefore, according to Abe and others (2016), the location of this constriction may be the location of the dynamic balance line. However, our velocity and elevation observations indicate that glacier velocity only increases significantly in the lowermost 21 km of the glacier during a surge, suggesting that both the reservoir and receiving zones are within this region.
Indeed, we observe mass movement from a reservoir zone at 8–21 km from the terminus, across the dynamic balance line at ~8 km, and into the receiving zone at 0-8 km from the terminus. This suggests that the constriction at 21 km limits the up-glacier extent of the surges, where only the portion of the glacier downstream of the valley constriction exhibits true surge-type behavior. However, the role of the valley constriction is not fully understood. While we and Abe and others (2016) both suggest it plays a crucial role in controlling how surges, we lack observations to determine how it controls the surges. New bed elevation and glacier surface velocity observations are needed to understand the role of the Donjek constriction in surging.

Although the region up-glacier of the constriction is not actively involved in surging, it plays a key role in the refilling the reservoir zone. We observe rapid refilling of the reservoir zone following a surge event, leading to a positive geodetic mass balance over the lower 32 km of the glacier from 2013 to 2016 (Fig. 9). We currently lack sufficient elevation data in the accumulation zone to determine the precise source of this mass. This suggests that the constriction may still exert a control on the surge behavior of Donjek Glacier, but as the boundary between the lower ‘surge-type’ and upper ‘normal’ portions of the glacier.

Consistent with other findings (Adalgeirsdóttir and others, 2005), our results suggest that surge-type glaciers can have a strong negative mass balance during a surge by moving large amounts of mass to lower, warmer elevations (Fig. 9). Immediately following a surge, the ablation area of a glacier can gain mass as it is dynamically refilled from an area above the reservoir zone. This suggests that the mass loss during quiescence does not reflect the climatic mass balance of the glacier (cf. Gardelle and others, 2013). Instead, our results indicate that a climatic mass balance can be derived from surge-type glaciers by comparing the glacier surface elevation at equivalent points in a surge cycle (e.g., just before a surge event; Fig. 9). However, it
should be noted that this study is limited in the spatial extent of elevation measurements by narrow lidar swaths collected in 2000, 2012, and 2016.

The relationship between mass balance and surging at Donjek may be complicated by the presence of the surge-type tributary glacier 23 km upstream of the terminus. Abe and others (2016) suggest that surges of this tributary occur independently of surges of the main trunk, but because the tributary adds mass to the trunk it could cause the trunk to surge sooner than it would do otherwise. Although the 1974 and 2010 tributary surges appear to have added mass to the trunk (Fig. 2b), it seems that they had little impact on the recurrence interval of the surges of the main glacier. This suggests that the surge-type tributary plays a minor role, if any, in the surging of Donjek Glacier.

2.6.2 Surge mechanisms and comparison between events

Robust observations of repeat surges in a glacier are rare due to decadal to multi-decadal surge recurrence intervals, the relatively short time period over which high resolution satellite observations exist, and the scarcity of repeat elevation data. Variegated (Eisen and others, 2005), Lowell (Bevington and Copland, 2014), Bering (Fatland and Lingle, 1998; Burgess and others, 2012) and now Donjek Glaciers, are among the only glaciers to have both elevation and velocity measurements for multiple surge events. The surges of Variegated Glacier were all similar in timing and scope, except the 1995 surge, which did not reach full maturity (Eisen and others, 2005). It is still difficult to assess variability in recent surges of Lowell Glacier due to a lack of elevation data and spatially extensive velocity maps (Bevington and Copland, 2015). The two recent surges of Bering Glacier have showed different initiation patterns and timing, as the 2008-2011 surge initiated over 90 km of Bering Glacier (Burgess and others, 2012) whereas the 1993-1995 surge had an isolated initiation zone (Fatland and Lingle, 1998; Burgess and others, 2012).
The surge time series presented here suggests that Donjek Glacier may have multiple surge initiation mechanisms. During some surge events the terminus begins to advance first, while the entire glacier starts moving at surge initiation in other events (Fig. 3). The 1977-1979 and 2012-2014 surge events both exhibited rapid motion of upstream ice (1-21 km from terminus) followed by terminus advance, while the 1988-1990 and 2000-2002 events exhibited terminus advance followed by upstream speedup (Fig. 3). Terminus advance appears to consistently occur at approximately the same time of year (August-September) for each surge event, regardless of velocity further up glacier (Fig. 3). We hypothesize that a surge can be triggered in two different ways, observed as up-glacier velocity increase or near-terminus advance, at Donjek Glacier, potentially related to how the subglacial drainage system is routed. We do not observe any kinematic waves propagating up or down glacier, suggesting that meltwater at the bed of the glacier acts over the lowest 21 km of the glacier nearly equally and/or rapidly propagates (in a matter of hours or days). However, the specific surge mechanisms at Donjek remain an open question that requires further research to address.

While we observe different patterns in surge initiation and termination, the location of the dynamic balance line appears to be controlled by the bed or valley curvature or both. We observe the dynamic balance line to be in the same location for the two most recent surge events with elevation measurements, for which we also observe the upper limit of surge activity to be at ~21 km upstream of the terminus. The geology underlying Donjek Glacier is composed of Silurian limestone, marble, argillite and phyllite in the accumulation area to Pennsylvanian quartz monzodiorite and diorite, and Devonian quartz-rich, micaceous, calcareous siltstone to sandstone in the lowest 22 km (Yukon Geological Survey, 2018). The terminus of Donjek sits on a strike-slip fault running east-west (Yukon Geological Survey, 2018). The lithological transition
suggests that bedrock geology may have a control on the segment of Donjek Glacier that exhibits surge-type behavior. However, it is unclear if the valley constriction or the transition in bedrock lithology determines the upper limit (Fig. 1). Observations of other glaciers suggest that geology can favor surge-type over ‘normal’ glaciers (Jiskoot and others, 2000; 2003), but there appears to be no spatial pattern amongst the geology underlying surge-type glaciers in the Yukon (Clarke and others, 1986; Crompton and others, 2018). Observations of the bed profile, glacier thickness and substrate (bedrock type; till properties; etc.) are needed to better understand the causes and spatial extent of surging on Donjek Glacier.

2.6.3. **Surge frequency and extent**

Long-term changes in mass balance can cause changes in glacier surge frequency within a region (Dowdeswell and others, 1995; Copland and others, 2011). Although Donjek Glacier has experienced a negative mass balance in recent decades (Larsen and others, 2015) the recurrence interval of 12 years since the 1988–1990 surge event has not changed. While the quiescent interval appears to be 2 years shorter from 1980 to 1988 than it has been since (Table 4), we are unable to confidently measure quiescent phase length in early time periods. Interestingly, despite the largely constant repeat interval, the most recent active phase from 2012 to 2014 was the shortest on record at 1.28–1.69 years (Table 4). Despite the improved characterization of surging presented here, there is still insufficient evidence to state whether the duration of the recurrence interval, the quiescent phase, or the active phase is changing significantly through time. The long-term negative mass balance in the study region has likely caused each of the successive surge events to be less extensive than the last (Fig. 2), consistent with the findings of Abe and others (2016). The next surge of Donjek Glacier is projected to occur in the mid-2020s, based on
a consistent ~12 repeat interval, and will likely be less extensive than the 2014 maximum surge extent.

2.6.5. **Comparison to other surge-type glaciers**

Although other authors have found that glacier surges in Alaska tend to initiate in winter (Raymond, 1987; Harrison and Post, 2003; Abe and Furuya, 2015), we show that the 1977–1979, 1988–1990, 2000–2002, and 2012–2014 surges of Donjek Glacier initiated in summer, similar to West Fork Glacier, Alaska (Harrison and others, 1994). Furthermore, Abe and Furuya (2015) found that surge-type glaciers in their quiescent phase frequently have higher winter than summer velocities. However, at Donjek Glacier, inter-annual variability during quiescence is greater than the seasonal difference between summer (Abe and others, 2016) and winter velocity (Van Wychen and others, 2018).

Velocity cross sections show that sliding is likely the dominant mechanism of increased motion during the surge phase of Donjek Glacier as glacier velocity is elevated nearly uniformly across the width of the glacier (Fig. 5). This is consistent with the commonly observed “plug flow” during surges (Kamb and others, 1985; Harrison and others, 1994; Murray and others, 2003; Pritchard and others, 2005). It is difficult to discern patterns in velocity across the width of glaciers in their surge phase due to a lack of observations.

Steele (~35 km long), Lowell (~65 km long), and Trapridge Glaciers (<3 km) are the three geographically closest surging glaciers to Donjek that have received considerable attention. The 1966–1968 surge of Steele Glacier caused the ice surface to uplift 260 m (Stanley, 1969). Altena and others (in review) found that Steele surged again between 2014–2016, although it has not yet been thoroughly documented. Lowell Glacier has experienced five surges since 1948, with recurrence intervals of 12 to 20 years (Bevington and Copland, 2014). Over the course of
the 2009–2010 surge, Bevington and Copland (2014) measured a 60 m maximum increase in elevation of the glacier surface. Trapridge Glacier underwent a slow surge from ~1980–2000 leading to little surface uplift, however, the glacier advanced 450 m (Frappé and Clarke, 2007). Of these four well documented surge-type glaciers in the Yukon, Donjek has the shortest recurrence interval and displaces the second least amount of mass during a surge event after Trapridge Glacier.

Other surge-type glaciers in the Alaska region with surface elevation change observations include Muldrow (Post, 1960), West Fork (Harrison and others, 1994), Variegated (Eisen and others, 2005), and Bering glaciers (Burgess and others, 2012; Table 5). Of these, Donjek Glacier experiences the least amount of elevation change in either the reservoir or receiving zone and has the least extensive reservoir zone. This suggests a connection between reservoir zone size and surge volume. While a great deal of work has been done on Trapridge Glacier (Clarke and others 1984; Clarke and Blake, 1991; Frappé and Clarke, 2007), little work has been done to understand mass transfer from the reservoir to receiving zones for surge-type glaciers in general.

With the exception of Bering Glacier (Burgess and others, 2012), the maximum elevation gain is always larger than the minimum elevation loss for all surge-type glaciers in Alaska and the Yukon. Peak elevation loss exceeds elevation gain at the tidewater glacier Sortebrae in Greenland (Jiskoot and others, 2001), while tidewater glacier surges in Svalbard have almost equal peak elevation increase/decrease (Table 5). Variegated Glacier is the only known Alaskan glacier whose surge involves the entire length of the glacier (Eisen and others, 2005). Variegated is also one of the only two glaciers in the Alaska region which has its annual snowline within the reservoir zone (West Fork Glacier being the other; Table 5). Although it is possible that other surge-type glaciers exhibit similar elevation change characteristics as those mentioned here, it is
difficult to compare surge phase elevation change to other glaciers around the world due to a lack of glacier wide elevation data.

Table 5. Surge elevation change, zone length, and glacier details from around the world.

<table>
<thead>
<tr>
<th>Glacier name</th>
<th>Location</th>
<th>Surge year(s)</th>
<th>Maximum measured elevation increase/decrease (m)*</th>
<th>Dynamic balance line distance from terminus (km)**</th>
<th>Reservoir zone length (km)</th>
<th>Glacier length (km)</th>
<th>Snow line relative to reservoir zone?</th>
<th>Years since previous surge (years)</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Muldrow</td>
<td>Alaska</td>
<td>1956 to 1957</td>
<td><del>+200/</del>-100</td>
<td>17</td>
<td>20</td>
<td>~63</td>
<td>Up-glacier</td>
<td>&gt;50</td>
<td>Post, 1960</td>
</tr>
<tr>
<td>West Fork</td>
<td>Alaska</td>
<td>1987 to 1988</td>
<td>+120/-70</td>
<td>16</td>
<td>21</td>
<td>~40</td>
<td>Within</td>
<td>52</td>
<td>Harrison and others, 1994</td>
</tr>
<tr>
<td>Variegated</td>
<td>Alaska</td>
<td>1995</td>
<td>+110/-40</td>
<td>4</td>
<td>16</td>
<td>~20</td>
<td>Within</td>
<td>12</td>
<td>Eisen and others, 2005</td>
</tr>
<tr>
<td>Bering</td>
<td>Alaska</td>
<td>2008 to 2011</td>
<td>+20/-110</td>
<td>120-130</td>
<td>&gt;30</td>
<td>~16/5</td>
<td>Unknown</td>
<td>13</td>
<td>Burgess and others, 2012</td>
</tr>
<tr>
<td>Sabche</td>
<td>Nepal</td>
<td>2012-2017</td>
<td>+90/-60</td>
<td>1.5</td>
<td>4.5</td>
<td>6.5</td>
<td>Unknown</td>
<td>10</td>
<td>Lovell and others, 2018</td>
</tr>
<tr>
<td>Khurdop in</td>
<td>Pakistan</td>
<td>2017</td>
<td>+160/-80</td>
<td>12</td>
<td>13</td>
<td>~41</td>
<td>Up-glacier</td>
<td>18</td>
<td>Steiner and others, 2018</td>
</tr>
<tr>
<td>Glacier</td>
<td>Svalbard</td>
<td>Start Year - End Year</td>
<td>Elevation Change</td>
<td>Max Elevation</td>
<td>Mass Balance</td>
<td>Frequency of Surges</td>
<td>Duration</td>
<td>Authors</td>
<td></td>
</tr>
<tr>
<td>---------------</td>
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<td></td>
</tr>
<tr>
<td>Usherbre<strong>en</strong></td>
<td>Svalbard</td>
<td>1978 to 1985</td>
<td>+150/&gt;=-70</td>
<td>4.8</td>
<td>&gt;2</td>
<td>12 (pre-surge)</td>
<td>Unknown</td>
<td>Hagen, 1987</td>
<td></td>
</tr>
<tr>
<td>Comfortlessbreen</td>
<td>Svalbard</td>
<td>2006 to 2010</td>
<td>~100/-80 to -100</td>
<td>9</td>
<td>&gt;3</td>
<td>~15</td>
<td>Within</td>
<td>&gt;70</td>
<td>King and others, 2016</td>
</tr>
<tr>
<td>Osbornebreen</td>
<td>Svalbard</td>
<td>1986 to 1988</td>
<td>+100 /-100</td>
<td>Unknown</td>
<td>Unknown</td>
<td>20</td>
<td>Unknown</td>
<td>Unknown</td>
<td>Rolstad and others, 1997</td>
</tr>
<tr>
<td>Zawadzkibreen</td>
<td>Svalbard</td>
<td>2000s</td>
<td>+70/-40</td>
<td>9</td>
<td>8.5</td>
<td>17.5</td>
<td>Within</td>
<td>~70</td>
<td>Sund and others, 2014</td>
</tr>
<tr>
<td>Bivachny</td>
<td>Tajikistan</td>
<td>2012 to 2015</td>
<td>+85/-68</td>
<td>14.5</td>
<td>~22.5</td>
<td>~37</td>
<td>Within</td>
<td>21</td>
<td>Wendt and others, 2017</td>
</tr>
<tr>
<td>Medvezhiy</td>
<td>Tajikistan</td>
<td>1988 to 1989</td>
<td>~+100/-20</td>
<td>3</td>
<td>5</td>
<td>~15</td>
<td>Up-glacier</td>
<td>15</td>
<td>Osipova and Tsvetkov, 1991</td>
</tr>
<tr>
<td>Lowell</td>
<td>Yukon</td>
<td>2009 to 2010</td>
<td>+60/Unknown</td>
<td>&gt;25</td>
<td>Unknown</td>
<td>~65</td>
<td>Unknown</td>
<td>12</td>
<td>Bevington and Copland, 2014</td>
</tr>
<tr>
<td>Donjek</td>
<td>Yukon</td>
<td>2012 to 2014</td>
<td>+74/-66</td>
<td>8</td>
<td>13</td>
<td>65</td>
<td>Up-glacier</td>
<td>12</td>
<td>This study</td>
</tr>
<tr>
<td>Steele</td>
<td>Yukon</td>
<td>1966 to 1968</td>
<td>+260/-160</td>
<td>8 to 13</td>
<td>&gt;7</td>
<td>~35</td>
<td>Unknown</td>
<td>Unknown</td>
<td>Stanley, 1969</td>
</tr>
</tbody>
</table>

*Elevation change is dependent on timing of available DEMs

** Terminus position defined by each study author

2.7. Conclusions

Donjek Glacier has retreated ~2.5 km since its pre-1874 Little Ice Age maximum extent to its most recent advanced post-surge position in 2014. Despite this retreat, and an overall negative mass balance, Donjek has surged regularly, with 8 events since 1935 (~1935, ~1947, late-1950s,
~1969, 1977–1979, 1988–1990, 2000–2002, 2012–2013) and a repeat interval of 9–12 years. During each of the last three surges Donjek increased in area by 3 to 7 km$^2$ (0.7–2% of total area) due to terminus advance. We found that the 2000–2002 and 2012–2014 surge events initiated in summer months, but did not reach their full active phase until the following spring (2001 and 2013, respectively). The surge is limited to the lower 21 km of Donjek Glacier as mass is redistributed from the reservoir zone (8–21 km from the terminus) to the receiving zone (0–8 km from the terminus). Ice velocity is fastest in the receiving zone, reaching speeds as high as 1700 m a$^{-1}$. While the location of the dynamic balance line remains constant between recent surge events, observations suggest different ways in which surges can initiate for unique surge events. The entire portion of the glacier affected by the surge begins to move rapidly before we observed terminus advance for the 1977–1979 and 2012–2014 surge events, while the opposite is true for the 1988–1990 and 2000–2002 events. This suggests that Donjek surge events can be triggered at different locations, and perhaps by different mechanisms.

Abe and others (2016) hypothesized that significant narrowing of the valley at 21 km from the terminus controlled the surging of Donjek Glacier, but we show here that the constriction actually defines the upper end of the active surge zone. This surge zone corresponds to the part of Donjek Glacier underlain by metasedimentary rock, and the constriction corresponds to the area were the underlying lithology changes from metasedimentary up-glacier to igneous down-glacier (Fig. 1). Thus, the role of the constriction in surging is complicated as it represents both a change in geometry and lithology.

We find elevated rates of mass loss during the surge phase, while comparatively little mass is lost during quiescence. During the 2012–2013 surge event, Donjek Glacier had a net ice volume loss of $\sim 0.51 \pm 0.0000082$ km$^3$ in the lower 21 km of the glacier, the active surge zone.
In the few years following a surge event we observe the ablation area to have a positive geodetic mass balance, suggesting rapid refilling from up-glacier regions. Thus, it is vital that surge-type glacier mass balance measurements are undertaken at the same time in the surge cycle (e.g., immediately before or after a surge event) for inclusion in glacier response to climate studies (cf. Yde and Paasche, 2010) and mass balance inventories (cf. Larsen and others, 2015).
CHAPTER 3
THE IMPACT OF CLIMATE ON SURGING AT DONJEK GLACIER, YUKON, CANADA

3.1. Abstract
Potential links between climate and glacier surges are not well understood, but are required to enable prediction of glacier surges and mitigation of associated hazards. Here, we investigate the role of snow accumulation and temperature on surge periodicity, glacier area changes, and surge initiation season since the 1930s for Donjek Glacier. Three ice cores from Eclipse Icefield, at the head of the glacier, indicate that a total accumulation of 13.1 to 17.7 m w.e. of snow occurred in the 10-12 years between each of its last eight surges. This suggests that a threshold must be passed before the initiation of a surge event, although it remains unclear whether the relationship between cumulative snowfall and surging is due to the consistency in repeat surge interval and decadal average precipitation, or if it is indeed related to surging. We also examined the 1968 to 2017 climate record from Burwash Landing to determine whether a relationship exists between surge periodicity and an increase of 2.5°C in mean annual air temperature over this period. No such relationship was found, although each of the past 8 surge events has been less extensive than the previous, with the maximum terminus extent approximately 7.96 km2 smaller in the most recent 2012-2014 surge event than the ~1947 surge event.

3.2. Introduction
Surge-type glaciers account for about 1% of glaciers globally (Sevestre and Benn, 2015), but can be the dominant glacier type in some regions (e.g., Clarke et al., 1986; Jiskoot et al., 2003), and are important for understanding ice flow instabilities and anomalous glacier response to climate change (Yde and Paasche, 2010). Surge-type glaciers have long periods of flow below their balance velocity (quiescent phase), typically on the order of decades, which are interrupted by
short-lived phases of glacier flow at rates much higher than the balance velocity (active phase or surge phase), typically on the order of months to years, driven by internal instabilities, and sometimes leading to a marked frontal advance (Meier and Post, 1969; Clarke, 1987). When a glacier surges, its reservoir zone at higher elevations loses mass and its receiving zone at lower elevations gains mass, with the line of zero net mass change defined as the dynamic balance line (DBL: Dolgoushin and Osipova, 1975). When mass gain in the receiving zone leads to a significant advance of the terminus, an increased calving flux or other proglacial hazards can occur.

Surges of mid-latitude glaciers are typically hypothesized to initiate when a critical basal shear stress is reached in a surge initiation region, causing the subglacial hydrologic system to reorganize and the glacier to rapidly redistribute its accumulated mass downglacier (Meier and Post, 1969; Raymond, 1987; Eisen et al., 2005). While this hydrologic mechanism dominates Yukon-Alaska type surging, a thermal triggering mechanism (i.e., surging controlled by basal ice temperature), or combined hydro-thermodynamic mechanism, has been documented in surges of polar and polythermal glaciers, such as those in Svalbard and smaller glaciers in Yukon-Alaska (Murray et al., 2003; Frappé and Clarke, 2007; De Paoli and Flowers, 2009; Dunse et al., 2015). Finally, overarching theories related to balance flux (Budd, 1975) and enthalpy (Sevestre et al., 2015) have been proposed as well.

The length of a surge cycle (i.e., combined quiescent and active phases) is typically quite consistent for a glacier, and is proportional to the length of the surge phase (Meier and Post, 1969; Dowdeswell and others, 1991). In turn, quiescence duration is controlled by mass balance conditions (Robin and Weertman, 1973), meaning that surge periodicity is inversely related to accumulation rates (Dyurgerov et al., 1985; Osipova and Tsvetkov, 1991; Dowdeswell et al.,
Prolonged quiescent phases typical of the Svalbard region have been ascribed to low accumulation rates, often only on the order of 0.3-0.6 m a\(^{-1}\) (Dowdeswell et al., 1995), while short repeat intervals on Variegated Glacier, AK, correspond to accumulation rates on the order of 1.4 m a\(^{-1}\) (Eisen et al., 2001; Van Geffen and Oerlemans, 2017). However, there can be large variations in surge periodicity between glaciers in the same region, perhaps related to whether their surges are driven by a hydrologic or thermal triggering mechanism. For example, Icelandic glaciers have irregular quiescent intervals; 5-30 years for some glaciers and up to 100-140 for others (Björnsson et al., 2003; Sigurdsson, 2005).

Changes in surge recurrence interval have been linked to changing cumulative mass balance (Dowdeswell et al., 1995; Copland et al., 2011; Eisen et al., 2001; Striberger et al., 2011). Dowdeswell et al. (1995) found a persistent negative mass balance to reduce the glacier surge activity in Svalbard. In contrast, Copland et al. (2011) found an increase in precipitation and positive glacier mass balance on Karakoram glaciers to drive an increase in the number of surge events. Similarly, Eisen et al. (2001) reported a variable surge recurrence interval that is consistent with changing amounts of precipitation on Variegated Glacier, Alaska. Striberger et al. (2011) find variable rates of surging of Eyjabakkajökull, Iceland over the last several hundred years linked to changes in climatically driven mass balance.

Previous efforts to examine connections between cumulative snow accumulation and length of the quiescent phase have used mass balance models, off-ice meteorological measurements, and only a very limited record of in situ mass balance measurements (Eisen et al., 2001; Tangborn, 2013; Dyurgerov et al., 1985). Although these studies found that a snow accumulation threshold had to be reached before each surge started, this potential linkage has not yet been tested with observations of glacier surface mass balance. Here, we use the well-
documented history of surge events at Donjek Glacier (Abe et al., 2016; Kochtitzky et al., In Review), and ice cores extracted from Eclipse Icefield at the head of the glacier (Wake et al., 2002; Yalcin et al., 2006; Kelsey et al., 2012), to explore linkages between snow accumulation and surging since the 1930s. We combine these observations with weather station records, digital elevation models, and remote sensing analysis to examine the impacts of climate and ice kinematics on surge behavior. The combination of a high surge recurrence interval, documentation of eight surge events, and three independent ice core records in the accumulation zone, make Donjek Glacier the most ideal site to test the influence of climate on surge behavior. The prediction of surge occurrence from snow accumulation observations would allow for improved hazard forecasting and for the deployment of field instruments to observe surging kinematics in detail, advancing the present understanding of controls on glacier instabilities.

Figure 10. (a) Donjek Glacier (blue outline; RGI Consortium, 2017), with Eclipse Icefield marked with yellow star and Donjek River in light blue. Black line indicates the separation between the downglacier surge-type and upglacier non-surge-type portions of the glacier. Green box indicates extent of figure 16a. (b) Location of Donjek Glacier in southwestern Yukon; red box indicates extent of a. Base image from Landsat 8, 23 September, 2017.
3.3. Study Site

Donjek Glacier (61°11’N, 139°31’ W; Figure 10) is a surge-type glacier located in southwest Yukon in the St. Elias Mountains. In 2010, Donjek Glacier was 65 km long with a surface area of 448 km² (RGI Consortium, 2017). While the Tlingit indigenous peoples of the Yukon were the first to observe Donjek Glacier surge (Cruikshank, 1981), the first scientific records in the form of Bradford Washburn’s air photos are from 1937. Subsequent scientific work focused on the moraines and geomorphology (Denton and Stuvier, 1966; Johnson, 1972a and b), meteorological measurements at Eclipse Icefield as part of the Icefield Ranges Research Project (Ragle, 1972), and Donjek’s surge-related outburst floods in the Donjek River (Figure 10; Clarke and Mathews, 1981). Ice coring campaigns have occurred at least four times at Eclipse Icefield since the 1990s, and provide a wealth of accumulation and atmospheric information (Wake et al., 2002; Yalcin et al., 2006; Kelsey et al., 2012).

![Figure 11. Surge event timing. Grey bars indicate uncertainty among surges before the satellite era. Black bars indicate duration of active surge phase for the last four surge events, constrained by satellite imagery.](image)

The area-averaged mass balance was -0.29 m water equivalent (w.e.) yr⁻¹, or -0.13 Gt yr⁻¹, between May 2000 and May 2012 (Larsen et al., 2015). Despite this negative mass balance, the glacier has continued its history of frequent surging, which has occurred approximately every 10-12 years since the 1930s (Abe et al., 2016; Kochtitzky et al., In Review; Figure 11). Air photo records, satellite imagery and previous reports indicate that the glacier surged in ~1935, ~1947, late-1950s, ~1969, 1977-1980, 1988-1990, 2000-2002, and 2012-2014, with progressively less
extensive terminus advances up to the present day. Ice flow velocities are only available for the two most recent surges (Abe et al., 2016; Kochtitzky et al., In Review). Only the lower 21 km of the glacier was involved in these surge events, coinciding with the portion of the glacier below a valley constriction (Kochtitzky et al., In Review; Figure 10). This extensive record of surge behavior is here complemented by ice cores from the Eclipse Icefield at the head of Donjek Glacier.

3.4. Methods

3.4.1. Ice cores and snow accumulation record

Ice cores were collected at Eclipse Icefield (Fig. 10) in 1996 (160 m absolute length; Yalcin and Wake, 2001), 2002 (350 m absolute length; Fisher et al., 2004), and 2016 (59 m absolute length; unpublished), to develop an understanding of past climate in the St. Elias Range. Cores were collected during late spring, but preceding the melt season, in 1 m segments using the 8 cm diameter Eclipse Drill. The accumulation record from the 1996 ice core was originally report by Yalcin and Wake (2001) and we use their original, un-altered, record in this study. We extracted accumulation records from the 2002 and 2016 ice cores via annual layer counting of cyclic glacio-chemical signals. The 2016 core was primarily dated using oxygen and deuterium isotope ratios, and deuterium excess, with additional constraints from sodium, sulfate, and magnesium. We do not apply any thinning corrections to the 2016 core, as it only covers the top 59 m of the firn zone where thinning is negligible. Due to a lack of oxygen isotope measurements, deuterium excess data are unavailable for the 2002 core, so this core was dated using hydrogen isotope ratios with sodium, magnesium, calcium, sulfate, and cesium. The 2002 core was additionally constrained by known volcanic eruption markers indicated by a spike in sulfate concentrations (Yalcin et al., 2007) and the Cs-137 peak in 1963 from nuclear bombs. The seasonal timing of
each of these peaks is well characterized from previous studies in the North Pacific region (Yalcin et al. 2001, Wake et al. 2002, Yasunari et al. 2007, Osterberg et al. 2014, Winski et al. 2017).

Five individuals independently picked the approximate position of the 1 January marker throughout the last 500 years for the 2002 ice core. These individual annual pick positions were reconciled using the methods described in Winski et al. (2017). With the resulting annually-dated timescale, annual layer thicknesses were calculated as the distance between successive years, and water equivalent annual layer thicknesses were calculated as the annual layer thickness multiplied by the density at the corresponding depth in the ice core. The density for each layer was extrapolated from the 1 m-increment field density observations.

In the 2002 ice core we accounted for thinning due to glacier flow using three widely used 1-dimensional glacier flow models, which we refer to as the Nye (Nye, 1963), Hooke (Kaspari et al. 2008), and Dansgaard-Johnsen (Dansgaard and Johnsen, 1969) models. Following Winski et al. (2017), we tested all reasonable combinations of free parameters in each model to assess which model run most closely matches our observed depth-age scale (Fig. 12). In each model, we generated a suite of different age scales using long-term average accumulation rates ranging from 20 to 300 cm in 10 cm increments. In the Hooke model, we also permitted the flow parameter \(m\) in Kaspari et al. 2008 to vary between 1 and 2. In the Dansgaard-Johnsen model we permitted a flow regime change occurring between 10 and 250 m above the bed. These activities resulted in a total of 1363 separate model runs (29 Nye, 609 Hooke and 725 Dansgaard-Johnsen), each producing a unique depth age scale.
Figure 12. Ice core accumulation and depth-age scale. The 1995 ice core (purple), 2002 ice core (green), and 2016 ice core (blue) accumulation records are shown with the linear fit (red) of the mean of the combined core record since 1770. The linear fit has a slope of 1.6 mm a$^{-1}$. The observed depth-age scale from the 2002 ice core is shown in black.

For each modeled depth age scale, we calculated the sum of root-mean squared errors (RMSE) between the layer-counted and modeled depth-age scale positions at each year. We found the optimized version of the Dansgaard-Johnsen Model to produce the closest match to our observed depth-age scale. In our depth-age scale modeling, we used 337 m w.e. (approximately 376 m of absolute thickness) which yielded the lowest error between the optimized Dansgaard-Johnsen model (the closest fit) and the layer counted timescale. The accumulation rate used herein is equal to the ratio of the observed layer thickness (from the annual layer counting) over the modeled layer thickness (from the optimized Dansgaard-Johnsen model) multiplied by 1.4, which is the optimized value of long-term accumulation that produces the best fit to the
timescale. Based on the range of results among the three flow models, the accumulation uncertainty was estimated as ±15% in the 1930s, with lower uncertainties near the top of the record.

We define our cumulative accumulation interval for each quiescent phase to stretch from the year following surge initiation to the initiation year of the next surge (Eisen et al., 2001), which equates to 1935-1944, 1945-1955, 1956-1966, 1967-1977, 1978-1988, 1989-2000, and 2001-2012. The surge initiation dates we use are from Kochtitzky et al. (In Review), which are well constrained in the satellite era, but within the uncertainty boundaries determined by Kochtitzky et al. (In Review) from advanced terminus positions and/or push moraines.

3.4.2. Glacier surface elevation mapping

Digital Elevation models (DEMs) for 2002, 2007, 2012, and 2016 were created, or obtained, from Operation IceBridge (OIB) LiDAR measurements, Satellite Pour l’Observation de la Terre 5 (SPOT-5), WorldView, and Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER; Table 6). OIB LiDAR tracks from 2012 and 2016 were downloaded from the National Snow and Ice Data Center (https://nsidc.org/icebridge/portal) and down sampled to 8 m spatial resolution for comparison with the DEMs. We obtained one SPOT-5 DEM (40 m spatial resolution) from the SPIRIT Project (https://theia-landsat.cnes.fr) with an uncertainty of ±6 m (Korona and others, 2009), collected on 13 September 2007. We received DEMs at 8 m spatial resolution derived from WorldView imagery from the University of Minnesota Polar Geospatial Center (PGC), with ~0.2 m vertical accuracy (Shean and others, 2016). We mosaicked the individual WorldView DEMs from 10 August and 27 September 2013 (hereby referred to as the August/September 2013 DEM) to create a more spatially extensive DEM of the glacier. These 2013 DEM strips do not overlap or intersect, so we are unable to quantify the
potential aliasing of glacier flow and/or melt on the accuracy of these measurements. Finally, we created one 2002 DEM from ASTER imagery using the software package MMASTER with 30 m spatial resolution and 10 m vertical uncertainty (Girod and others, 2017). We co-registered all DEMs to the WorldView DEMs follow methods from Nuth and Kääb (2011) and smoothed extracted centerline elevation values using a 300 m moving window to visualize the data.

Table 6. Elevation data sources for ice surface change

<table>
<thead>
<tr>
<th>Source</th>
<th>Date</th>
<th>Vertical uncertainty</th>
</tr>
</thead>
<tbody>
<tr>
<td>Operation IceBridge</td>
<td>22/05/2012</td>
<td>&lt;10 cm</td>
</tr>
<tr>
<td>(airborne LiDAR)</td>
<td>15/05/2016</td>
<td></td>
</tr>
<tr>
<td>ASTER (satellite)</td>
<td>26/05/2002</td>
<td>10 m</td>
</tr>
<tr>
<td>PGC/WorldView (satellite)</td>
<td>10/08/2013</td>
<td>~0.2 m</td>
</tr>
<tr>
<td></td>
<td>27/09/2013</td>
<td></td>
</tr>
<tr>
<td>SPOT-5 (satellite)</td>
<td>13/09/2007</td>
<td>6 m</td>
</tr>
</tbody>
</table>

3.4.3. Snowline measurements

To infer the position of the equilibrium line altitude, we digitex the position of the snowline using the Landsat archive. All available cloud-free Landsat images of Donjek Glacier were downloaded from Earth Explorer (https://earthexplorer.usgs.gov), and the last available image of the ablation season (July, August, or September) of each year was selected to determine the snowline for most years from 1972-2017. We additionally used one air photo from 8 July 1951, which we georeferenced with 8 tie points to produce an estimated horizontal uncertainty of 72.4 m.
We estimated the mean elevation of the snowline for each year using a WorldView digital elevation model (DEM) from 2013 (see section 3.4). We are unable to account for glacier surface elevation change over time due to a lack of high quality surface DEMs prior to 2002, but a lack of change in exposed rock along the glacier margins since the 1970s suggests that elevation changes have not been large.

3.4.4 Ice thickness measurements

We used a ground penetrating radar from Blue System Integration Ltd. (http://www.radar.bluesystem.ca/) with 5 and 10 MHz antennas to measure ice thickness over the lower ablation area of Donjek Glacier in July 2018. Data were processed using IceRadarAnalyzer 4.2.5, assuming a radio-wave velocity of 0.300 m ns\(^{-1}\) in air and 0.170 m ns\(^{-1}\) in ice (Mingo and Flowers, 2010).

3.4.5. Climate and weather observations

To infer climate conditions at Donjek Glacier, we use temperature and precipitation data from the Environment and Climate Change Canada weather station at Burwash Landing (61°22’14” N, 139°2’24” W, 806 m a.s.l.), 30 km northeast of the current glacier terminus (~1000 m a.s.l.). Data was downloaded from http://climate.weather.gc.ca using the Canadian Climate Data Scraping Tool (Bonifacio et al., 2015). The Burwash Landing weather station has been operational since 1968 and has a nearly continuous hourly and daily record.

We constructed a continuous annual mean temperature record from monthly average temperatures recorded at the weather station to examine long-term temperature change. We also reconstructed a record of annual positive degree days (PDD) from the daily temperature data as a means to infer surface ablation (e.g., Ohmura, 2001). Of the 18,263 day record from 1 January, 1968 to 31 December, 2017, 1038 days did not have mean daily temperature data. To fill these
gaps, we linearly interpolated missing data using the daily mean temperature observation nearest in time. We then calculated the number of annual positive degree days by summing the daily mean temperature for all days that exceeded 0°C for each calendar year.

We summed daily rainfall data from Burwash Landing to calculate annual liquid precipitation. Some of these daily data are missing, but we do not attempt to fill these, so annual estimates should be considered as minima. The precipitation data cover October 1966 to January 2013. These data allow us to examine the impacts of extreme and cumulative rain events. No elevational correction was applied to any weather data from Burwash Landing.

During summer months, from June to September for all years 307 days do not have weather readings. Of these no gap is more than 5 days except during the second half of August and all of September in 1987, September 2001, and June and September 2002.

3.5. Results

3.5.1. Cumulative accumulation

Using the cumulative annual snow accumulation from the three ice cores, we find that between 13.1 and 17.7 m w.e. (mean of 15.5 ± 1.46 m w.e.) accumulated at Eclipse Icefield between each of the eight recent surges of Donjek Glacier (Figure 13a). While the three ice cores do not record the same amount of accumulation each year, they do not show a pattern of persistent spatial bias of snow accumulation across Eclipse Icefield when compared to each other (Fig. 12).
Figure 13. Cumulative accumulation between surge events. (a) Cumulative annual accumulation from 1995 (black), 2002 (blue), and 2016 (red) ice cores between each surge event. Green circles indicate cumulative positive degree days between surge events on the right y-axis. (b) The cumulative accumulation from the 2002 ice core offset by 160 years, the time need for surface snow/firn/ice to travel from Eclipse Icefield to the constriction at 21 km from the terminus, where the surge-type portion of the glacier begins. Solid and dashed black lines show the mean and one standard deviation cumulative accumulation average between surge events.

Because surging is limited to the lower 21 km of Donjek Glacier (Kochtitzky et al., In Review; Figure 10), snow accumulation 32.3 km upstream of the valley constriction coinciding with the surge onset region may not have a strong influence on surge behavior. We therefore calculated the time that it would take snow/firn/ice to advect from Eclipse Icefield to this topographic constriction from the surface flow speed, neglecting ablation and any submergence or emergence velocity. In 2007, the average surface flow speed was 201.4 m yr\(^{-1}\) along the center flowline, with a spatial variability of 11.4 – 398 m yr\(^{-1}\) over the 32.3 km trajectory (Van Wychen et al., 2018). Thus, snow accumulated on Eclipse Icefield takes ~160 years to reach the...
constriction, assuming that present-day velocities are similar to those of the past. We therefore offset the accumulation record derived from the 2002 ice core by 160 years to reconstruct the accumulation history preceding the surges (e.g. precipitation that reached the constriction in 2002 fell in 1842). Using this offset record, the cumulative accumulation between the eight surge events ranges between 14.2 and 20.4 m w.e. (mean of 16.6 ± 2.0 m w.e; Figure 13b). Although this results in only a marginally wider range than the accumulation history that is not offset, the average cumulative accumulation is 6% lower.

3.5.2. Changes in the reservoir zone surface height

Donjek Glacier can be divided into two parts: surge-type and non-surge-type (Kochtitzky et al., In Review). The surge-type portion can further be divided into a reservoir zone (8-21 km upstream of terminus) and a receiving zone (lower 8 km). The area separating the reservoir and receiving zones is known as the dynamic balance line (DBL: Dolgoushin and Osipova, 1975), or the area that shows no net mass change during a surge event. Our surface DEM analysis demonstrates that surface elevation increases in the reservoir zone following a surge event, even though the entire reservoir zone is located in Donjek Glacier’s ablation area. Between 2002 and 2007, after the 2000-2002 surge, we measured a glacier surface height increase of up to 41.6 ±11.6 m in the 8-21 km reservoir zone, with an average of 12.5 m (Figure 14). From 2007-2012, covering the beginning of the 2012-2014 surge, the reservoir zone had an average surface elevation increase of 1.00 m (Figure 14). From 2013-2016, a period which includes the end of the 2012-2014 surge event, we measured an average surface elevation increase of 10.7 m in the reservoir zone (Figure 14).
Figure 14. Surface elevation change in the reservoir zone. Surface elevation change from 2002 to 2007 (dark blue), 2007 to 2012 (light blue), and 2013 to 2016 (light green). Extent of the reservoir zone indicated by black dashed lines at 8 km (dynamic balance line) and 21 km (constriction) from the terminus.

3.5.3. Snowline and regional temperature change

Our remote sensing data illustrate that the Donjek Glacier snowline has migrated upglacier by 55 m yr\(^{-1}\) horizontally and risen by \(~1.0\) m yr\(^{-1}\) in elevation over the period 1951 to 2017 (Figures 15 and 16a). Over the study period the snowline was lowest in 1977 (Figure 16a), with an accumulation area of 337.3 km\(^2\) and an Accumulation Area Ratio (AAR) of 75.3%. It reached its highest level in 2017 at \(~2550\) m, with an AAR of 68.4%. We do not find our snowline measurements to be biased by timing of the observation as months late in the melt season are not consistently different from months early in the melt season (Figure 16a).
Figure 15. Donjek Glacier snowline. (a) Green box indicates extent of b, black outline shows extent of Donjek Glacier on top of SPOT-5 DEM from 13 September, 2007. (b) Snowline from 1951 (blue) to 2017 (red). Satellite image from Landsat 8, 15 August, 2017.
Figure 16. Donjek Glacier climate. (a) Snowline measurements from the last available satellite image of each year in July (light blue), August (medium blue), and September (dark blue). Red line shows linear trend for study period with black error bars indicating one standard deviation. (b) Burwash Landing annual average temperature record (blue) with linear trend (red). (c) Mean accumulation record from 1995, 2002, and 2016 ice cores from Eclipse Icefield (green) with linear trend (red). (d) Rain from Burwash landing with annual (cyan) and monthly (black) totals from 1967 to 2012. Blue bars indicate a period when Donjek Glacier was known to surge, time periods found by Kochtitzky et al. (In Review).
3.5.4. Glacier geometry

Based on ground penetrating radar measurements below the dynamic balance line (8 km from glacier front; Fig. 10), we measured a bedrock rise towards the terminus (Figure 17). The bedrock elevation rises from 810 to 890 m over 0.70 km in the downstream direction, causing a 6.5° reverse bedrock slope (Figure 17), although the full spatial extent of this reverse slope is unclear. The ice thickness in this region ranges from 360 to 470 m, with deeper ice located closer to the dynamic balance line.

Figure 17. Donjek Glacier bed mapping (a) Donjek Glacier bed elevations in the reservoir zone, which indicate a reverse slope towards the terminus. Base image: Landsat 8, 23 September, 2017. Extent of figure indicated by green box in figure 10a. (b) Profile for along flow GPR transect of surface ice (blue) and bedrock (red). Extent of profile indicated by black line in 16a.
2.5.5. Temperature and precipitation patterns

An increase in mean annual temperature of ~0.05°C yr\(^{-1}\) occurred at Burwash Landing between 1968 and 2017, equivalent to ~2.5°C over the 50-year study period (Figure 16b), which is consistent with a rising snowline. The mean annual temperature at Burwash Landing reached a minimum of -6.86°C in 1973 and a maximum of 1.73°C in 2003. In addition to mean annual temperature rise, the cumulative positive degree days increased during each of the past four quiescent phases, from 15095 for the 1967-1977 quiescent period to 18899 for the 2001-2012 period (Figure 16a).

Annual snow accumulation derived from the Eclipse Icefield cores has not shown any significant trends over the study period, with values ranging from 0.62 to 1.91 m w.e. yr\(^{-1}\) (Fig. 16c). A linear fit to the annual average accumulation from 1948 to present has a non-significant positive slope of 0.6 mm w.e. yr\(^{-1}\) (95% confidence). However, the accumulation variance has increased from 0.0393 m\(^2\) (1948 to 1982) to 0.0677 m\(^2\) (1982 to 2016) in recent decades.

Precipitation records from Burwash Landing indicate that the initiation of the 1988, 2000, and 2012 surges have coincided with several of the rainiest years on record (Figure 16d). The top five annual rainfall totals on record from 1967 to 2012 for Burwash Landing were 2000 (293.5 mm), 2012 (284.0 mm), 1983 (274.9 mm), 1988 (273.9 mm), and 2005 (260.0 mm). However, the 1977 surge initiation coincided with relatively dry conditions (27th highest annual total rainfall year in the study period) (Figure 15d).

Three of the top ten rainiest months appear to coincide with surge onsets (Figure 16d). The rainiest month on record was July 1988 (131.8 mm) and Donjek started surging the next month (Kochtitzky et al., In Review). The third rainiest month on record occurred in August 2000 (114.7 mm) and Donjek started to surge that month or the next (Kochtitzky et al., In
Donjek surged at the end of the 1960s (Kochtitzky et al., In Review) and the tenth rainiest month on record occurred in July 1967.

3.6. Discussion

3.6.1. Snow and mass accumulation on surge-type glaciers

The time it takes for a glacier to build up to its pre-surge geometry depends on the initial ice volume displacement in the reservoir zone, the subsequent reservoir zone cumulative mass balance, and the flux imbalance between actual and balance flux during quiescence (c.f. Clarke, 1987). Eisen et al. (2001) found that Variegated Glacier’s cumulative mass balance consistently reached a threshold of 43.5 m ice equivalent (39.9 m w.e.) before the glacier surged, while Dyugerov et al. (1985) similarly found that a total of 360 ±70 million tons of mass accumulated between each of four surge events of Medvezhiy Glacier, Tajikistan. On Donjek Glacier, 15.5 ± 1.46 m w.e. or 16.6 ± 2.0 m w.e. accumulates at Eclipse Icefield between surge events, dependent on whether we account for an offset in redistribution to the surge initiation region, ~32 km downstream of the accumulation zone. For some glaciers, however, it is known that during surges not all mass accumulated in the reservoir zone is emptied during a subsequent surge: in Dyngjujökull, Iceland, for example, 13 km$^3$ of the 20 km$^3$ of ice accumulated in the reservoir zone during its 20-year quiescence was transported to the receiving zone in the 2 years of active surging (Björnsson et al., 2003). In addition, it is possible that the consistent net accumulation observed at Eclipse Icefield between surge events simply reflects consistent average accumulation (Wake et al., 2002; Kelsey et al., 2012) over each of the ~12-year surge intervals (Abe et al., 2015; Kochtitzky et al., In Review).

Surges of glacier systems with surge-type tributaries, or mass advection to or from adjacent basins (e.g., outlets from ice caps), can be irregular, and in some cases it can be difficult
to relate a surge interval to climatic conditions and accumulation rates, even under quasi-stable climatic conditions (Glazovskiy, 1996; Björnsson et al., 2003). One of Donjek Glacier’s tributaries surged in 2004 and 2010 (~23 km from terminus on east side of main trunk of Donjek shown in Fig. 10), adding mass to Donjek’s main trunk ~2 km upstream of the top of the trunk’s reservoir zone (Kochtitzky et al., in review). However, even though both these surges occurred in the quiescent phase of the main glacier, the tributary surges do not seem to have shortened the duration of these quiescent phases.

3.6.2. Climate and surge behavior

Surge-type glaciers occur preferentially, but not exclusively, in specific climate zones that are bounded by temperature and precipitation thresholds (Sevestre and Benn, 2015). Temporal changes in surge controls, and thus in surge propensity, can occur due to climate change or climate-forced changes in glacier size, elevation, hypsometry, thermal regime and/or subglacial drainage system. Glaciers have been observed to change their surge behavior to being less vigorous or complete cessation in some regions (Hoinkes, 1969; Frappé and Clarke, 2007; Hansen, 2003; Christoffersen et al., 2005; Clarke, 2014), while widespread renewed surge activity has recently occurred in the high Karakoram (Hewitt, 2007; Copland et al., 2011; Quincey et al., 2011). This suggests that the potential surge dependence on mass balance, melt conditions, thermal regime and related supra-, en- and subglacial hydrology, and changes therein, require better understanding (e.g. Dowdeswell et al., 1995; Eisen et al., 2005; Sund et al., 2009). Copland et al. (2011) report increased occurrence of surge-type glaciers in the Karakoram after a period of increased precipitation and positive mass balance. Conversely, Dowdeswell et al. (1995) report a reduced number of surge-type glaciers in Svalbard due to a negative shift in
glacier mass balance. Others have observed individual surge-type glaciers to alter their surge recurrence interval with climate, such as Eyjabakkajökull in Iceland (Stribeger et al., 2011).

Although temperature is increasing by 0.05°C per year at Burwash Landing, and Donjek Glacier has a negative mass balance, we do not observe an altered surge recurrence interval. Ice cores from Eclipse Icefield also show no significant solid precipitation trends (Wake et al., 2002; Kelsey et al., 2012). However, Kochtitzky et al. (In Review) report that each of the past 8 surges has been less areally extensive than the previous one, similar to other glaciers in the St. Elias, such as Lowell Glacier (Bevington and Copland, 2014). Less extensive surge events are likely caused by a persistent negative mass balance (Larsen et al., 2015), rising snowline, and increasing number of positive degree days. Similar observations of less extensive surge events during a period of negative mass balance have occurred in Iceland (Sigurdsson and Jónsson, 1995). This suggests that glacier wide mass balance controls the intensity of each surge event, while other mechanisms control the surge recurrence interval.

Rapid mass redistribution, and related surface lowering and frontal advance, during surges are important for short- and long-term glacier surface mass balance. Post-surge accelerated ablation, thinning and retreat rates have been measured and modeled for surge-type glaciers in Iceland (Adalgeirsdottir et al., 2005), West Greenland (Yde and Knudsen, 2007), Alaska (Muskett et al., 2008), and Svalbard (Nuttall et al., 1997; Moholdt et al., 2010). For Donjek Glacier, Yukon, surges lead to glacier-wide negative mass balance (Kochtitzky et al., In Review). While many of these glaciers are already experiencing a negative mass balance (Larsen et al., 2015), surge-type behavior is important to monitor in calculating and predicting future sea level rise in the face of current climate change.
3.6.3. Surge onset and weather

Weather has been suggested to affect surge initiation and termination (Harrison and Post, 2003), and in particular strong melt, heavy rainfall, and large annual accumulation rates. Here, we focus on surge initiation, as our results show that three of the top ten rainiest months at Burwash Station coincided with surge onsets of Donjek Glacier.

Lingle and Fatland (2003) postulated that a temperate glacier will not surge until it has built-up critical thickness (basal shear stress), and surface meteorological conditions occur that store a large volume of water englacially. For Alaska this is suggested to result in a common late-winter to spring surge onset (Raymond, 1987; Harrison and Post, 2003). A suite of anecdotal evidence supports this hypothesis (Kamb et al., 1985; Muskett et al., 2008; Pritchard et al., 2005), but there are also examples of temperate glaciers with surge initiation in seasons other than winter (Harrison et al., 1994; Björnsson et al., 2003). Surge initiation in polythermal glaciers may not be as dependent on the influx of surface meltwater, but rather on reaching a critical thickness combined with water trapped at the bed, and may therefore still involve enhanced snow or rainfall, but the start is potentially in any season (Quincey et al., 2011), although a spring start is also common for polythermal glaciers (Hodgkins, 1997; Jiskoot & Juhlin, 2009). Surge trigger zones in polythermal glaciers have also been correlated with ponding of water and extensive slush flows associated with heavy late-spring (wet) snowfalls alternated with short-term episodes of exceptionally high temperatures (Hewitt, 2007).

Although some of the above evidence and intuitive reasoning suggest that the seasonality of surges could indeed be different for temperate glaciers than for polythermal glaciers, no comprehensive analysis of seasonality of surge initiation and termination in combination with thermal regime and surge development exists to date. This is also hampered by the fact that,
traditionally, the onset of a glacier surge was considered when the fast motion could be measured, when crevasses became visible, or when a terminus advance was observed, but with current remote sensing capabilities it is possible to detect surge onset at an earlier stage (e.g., Sund et al., 2009).

### 3.6.4. Donjek surge mechanisms

Abe et al (2015) suggested that the constriction at 21 km from the terminus plays a crucial role in causing Donjek Glacier to surge. Kochtitzky et al. (In Review) suggested that the constriction was rather an upper extent of surge-type behavior, coincident with a change in bedrock lithology. We find no one conclusive factor that causes Donjek Glacier to surge, although we can conclude that positive degree days are not a significant control on surge recurrence interval. While Donjek Glacier reaches a consistent 13.1 to 17.1 m w.e. accumulation before a surge event, this number cannot be confidently linked with the surge recurrence interval given that it could also be an indicator of consistent decadal averaged accumulation. Even though we show refilling of the reservoir zone on Donjek Glacier, limited elevation measurements during recent surge events are inconclusive to use the reservoir zone as a predictor for future surge events without more data. Assuming that past accumulation is an indicator of future surge events, as displayed in Figure 13b, then the next surge is likely to occur between 2022 and 2026.

More observations of Donjek Glacier surge kinematics, bedrock, and valley geometry are needed to understand the surging dynamics. While we show a bedrock rise beneath the dynamic balance line, the relationship between the rise and surging is presently unclear. Flowers et al. (2011) suggest for a glacier in the nearby Donjek Range that its bedrock rise facilitates surging, because the reverse slope resists ice flow and enhances buildup of the surge reservoir zone.
during quiescence. Björnsson et al. (2003) conversely suggest from modeling results that overdeepenings and adverse bed slopes enhance hydraulically inefficient subglacial drainage on two surge-type glaciers in Iceland. The role of the bedrock rise in Donjek surging is presently unknown, although it almost certainly plays a role in controlling near-terminus ice dynamics, and thus is likely involved in surge dynamics.

3.7. Conclusion

We use three ice cores to reconstruct the accumulation record for Donjek Glacier leading up to seven documented surge events since the 1930s. We find that Eclipse Icefield received between 13.1 and 17.7 m w.e. (mean of 15.5 ± 1.46 m w.e.) total accumulation between surge events. While mean annual air temperatures increased by 2.5°C from 1968 to 2017 at Burwash Landing, 30 km from Donjek Glacier terminus, we observe no change in the surge recurrence interval over this time period, although each recent surge advance has become less extensive than the previous. While we find that cumulative accumulation is the most consistent climate variable between surge events of Donjek Glacier, our results remain inconclusive as to the role of accumulation in driving surge behavior. We suggest that yet unknown subglacial processes, possibly including changes in till deformation rates, are the primary driver of surging at Donjek Glacier, but mass accumulation remains a necessary precondition for a surge to initiate.

Satellite glacier surface elevation measurements reveal rapid refilling of the surge reservoir zone 8-21 km from the terminus of Donjek Glacier within the first 2 years following a surge event. We find almost no reservoir zone refilling occurring in the 5 years leading up to a surge event. This suggests that reservoir zone thickening is not the only cause of surge initiation, and therefore that a critical basal shear stress may need to be coincident with a hydrological switch. The highest rainfall amounts typically occur during the summer month preceding a surge
initiation. While not every observed surge initiates with a high rainfall amount, the three most recent surges (1988-1990, 2000-2002, 2012-2014) all coincide with one of the top five years on record for precipitation.

While we observe a bedrock rise in the receiving zone of Donjek Glacier, beneath the dynamic balance line, the role of overdeepening and a reverse bedrock slope in surging on Donjek Glacier remains a crucial question. Further observations of bedrock and bed elevation are necessary to understand surge mechanisms of Donjek Glacier. Monitoring surface elevation changes on Donjek Glacier as it prepares for its next surge event by the mid-2020s can yield valuable knowledge about how the subglacial hydrology beneath Donjek Glacier changes as a surge initiates. This will ultimately lead to more knowledge of surge initiation mechanisms, which can lead to better forecasting of surge events and magnitudes and therefore mitigate glacier hazards.
CHAPTER 4

SUMMARY AND CONCLUSIONS

Donjek Glacier is advancing our knowledge of glacier instabilities in the St. Elias mountains and world because we have been able to quantify eight instability events with detailed observations from in situ and remote sensing. The collections of unique records from Donjek Glacier, particularly the ice coring efforts over the last three decades at Eclipse Icefield, make Donjek a unique glacier to answer fundamental questions about glacier instabilities and glacier surging. Donjek Glacier is unique in its behavior amongst surge-type glaciers as surging is limited entirely within the ablation area, surges initiate in summer months and continue for nearly two years, and the glacier potentially has different mechanisms of initiating a surge event. This makes Donjek Glacier an important glacier to continue to monitor to better understand the glacier dynamics in the St. Elias mountains.

Donjek Glacier has retreated ~2.5 km since its pre-1874 Little Ice Age maximum extent to its most recent advanced post-surge position in 2014. Despite this retreat, and an overall negative mass balance, Donjek has surged regularly, with 8 events since 1935 (~1935, ~1947, late-1950s, ~1969, 1977–1979, 1988–1990, 2000–2002, 2012–2013) and a repeat interval of 9–12 years. The surge behavior is limited to the lower 21 km of Donjek Glacier as mass is redistributed from the reservoir zone (8–21 km from the terminus) to the receiving zone (0–8 km from the terminus). Observations of velocity increase of varying portions of the glacier suggest that Donjek surge events can be triggered at different locations, and perhaps by different mechanisms.

The surging portion of Donjek Glacier corresponds to the part of the glacier underlain by metasedimentary rock, and the constriction corresponds to the area were the underlying lithology changes from igneous up-glacier to metasedimentary down-glacier. Future work on
understanding the basal properties of Donjek Glacier is vital to understand what causes the glacier to undergo surge events.

While elevated rates of mass loss occur during the surge phase, comparatively little mass is lost during quiescence. During the 2012–2013 surge event, Donjek Glacier had a net ice volume loss in the active surge zone; however, in the few years following a surge event the ablation area had a positive geodetic mass balance, suggesting rapid refilling from up-glacier regions. This shows that surge-type glacier mass balance measurements need to be undertaken at the same time in the surge cycle (e.g., immediately before or after a surge event) for inclusion in glacier response to climate studies (cf. Yde and Paasche, 2010) and mass balance inventories (cf. Larsen and others, 2015). This also suggests that we need to carefully observe elevation changes along the glacier in the years during the quiescent phase to understand why and how Donjek Glacier surges.

Eclipse Icefield does receive a consistent amount of precipitation between surge events. During the last 8 surge events Eclipse Icefield received between 13.1 and 17.7 m w.e. (mean of 15.5 ± 1.46 m w.e.) total accumulation. While mean annual air temperatures increased by 2.5°C from 1968 to 2017 at Burwash Landing, 30 km from Donjek Glacier terminus, there was no change in the surge recurrence interval over this time period, although each recent surge advance has become less extensive than the previous. Thus, the impact of mass accumulation on the Donjek surge cycle remains unclear. It is likely that yet unknown subglacial processes, possibly including changes in till deformation rates, are the primary driver of surging at Donjek Glacier, but mass accumulation remains a necessary precondition for a surge to initiate.

It is highly likely that additional climate variables impact the surging of Donjek Glacier. For instance, the highest rainfall amounts typically occur during the summer month preceding a
surge initiation. While not every observed surge initiates with a high rainfall amount, the three most recent surges (1988-1990, 2000-2002, 2012-2014) all coincide with one of the top five years on record for precipitation.

While the chapters that comprise this thesis have substantially advanced knowledge of Donjek Glacier surge behavior, the mysteries of Donjek Glacier surging and surge-type glaciers more broadly abound. It remains unclear what is causing Donjek Glacier to surge. While it is highly likely that changes in the subglacial system are causing changes in glacier velocity, we currently do not understand the mechanism(s) by which this occurs. This necessitates future observations and modeling of Donjek Glacier to understand why and how Donjek Glacier surges.

Monitoring changes in the surface of Donjek Glacier are important as Donjek may enter into another active phase in the coming years. We particularly need improved surface elevation and velocity measurements with improved spatial and temporal resolution. While some of these observations can and will come from space, we need in situ observations as well. Installing time lapse cameras and GPS stations on and near the glacier would be useful to capture short term change in glacier surface height, velocity, and frontal position. These are important monitoring steps to forecast potentially blockages of the Donjek River and associated hazards for the Alaska Highway. We also need better weather data on and around Donjek Glacier. The closest temperature record comes from Burwash Landing, some 30 km from the terminus of Donjek Glacier. Monitoring temperature change over just one year could help to provide a better understanding of differences between weather on Donjek Glacier and at Burwash landing.

Furthermore, it is presently unclear what role temperature inversions may play in the role of changing ice temperature at Donjek Glacier. Establishing several temperature stations at various elevations will illuminate seasonal and annual patterns.
Further observations of bedrock and bed elevation are necessary to understand surge mechanisms of Donjek Glacier. Monitoring surface elevation changes on Donjek Glacier as it prepares for its next surge event by the mid-2020s can yield valuable knowledge about how the subglacial hydrology beneath Donjek Glacier changes as a surge initiates. This will ultimately lead to more knowledge of surge initiation mechanisms, which can lead to better forecasting of surge events and magnitudes and therefore mitigate glacier hazards. Collecting observations of bed lithology and temperature change through time would provide insights into the surge mechanism.

Donjek Glacier provides unique insights to understand how and why glaciers surge. Due to the high repeat interval of surges and the wealth of climate and geophysical data for Donjek Glacier, it is one of the best sites in the world to better understand the connections between climate and glacier instabilities. This work identified 8 surge events since the 1930s, showed that surging is limited to the lower 21 km, and describes the connections between surging and climate. Continued monitoring of Donjek Glacier is vital to both better understand glacier instabilities and reduce the hazards of surge type glaciers around the world, mainly in the Kluane region.
BIBLIOGRAPHY


http://www.geology.gov.yk.ca/update_yukon_bedrock_geology_map.html
BIOGRAPHY OF THE AUTHOR

William was born in Nashville, TN on August 24, 1993. He was raised in Nashville, TN and graduated from the University School of Nashville in 2012. He attended Dickinson College and graduated in 2016 with a Bachelor’s of Science degree in Earth Sciences. He came to Maine and entered the Earth and Climate Sciences & Climate Change Institute graduate programs at The University of Maine in the fall of 2016. William is a candidate for the Master of Science degree in Earth and Climate Sciences from the University of Maine in August 2019.