

Reconstructing the surge history and dynamics of Fisher Glacier, Yukon, 1948-2022

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Abstract

Glacier surges are periods of dynamic instabilities which result in semi-regular alternating periods of slow flow, termed the quiescent phase, and fast flow, termed the active phase. This study uses remotely sensed imagery, digital elevation models, glacier velocity datasets, and *in situ* oblique photographs to reconstruct the surge history and dynamics of Fisher Glacier to better characterize surging in the southwest Yukon and assess the risk posed by this glacier's surges on surrounding regions. Fisher Glacier has previously been identified as a surge-type glacier but, until now, it had not been the focus of any detailed studies.

We find evidence that Fisher Glacier underwent two surges during the study period from 1948 to 2022. Visual analysis of characteristic surge features on the glacier surface show that the glacier was in quiescence from <1948 to at least 1963. In 1972, an advanced terminus position, intense surface crevassing, and high point velocities suggest that a surge had recently terminated, corroborating a previous report of a surge occurring around 1970. This was followed by a 40-year quiescent phase from ~1973-2013 during which the terminus underwent consistent retreat, totaling a terminus-wide average of 2058 ± 8 m (up to 3567 ± 8 m in certain sections). Velocities during the quiescent phase were low (generally <50 m yr⁻¹), but underwent a slow multidecadal increase starting around 1985, spreading from the center of the glacier towards the head and the terminus. A pre-surge buildup phase beginning in ~2008-2010 resulted in velocities of up to ~200 m yr⁻¹. The active phase of the surge initiated in winter 2013/14 and was characterised by a velocity increase to ~1500 m yr⁻¹ that propagated both up- and down-glacier from the surge nucleus in the mid-region (~22 km upglacier from the terminus). Velocities peaked at >2100 m yr⁻¹ in the winter/early spring of 2016 at ~12 km from the terminus. The surge resulted in a mean terminus-wide advance of 868 ± 8 m, intense surface crevassing and a downglacier transfer of mass from the reservoir zone to the receiving zone. The terminus area increased in elevation by a mean of ~80 m. In July 2016, the surge rapidly terminated within a period of ~1 month, although velocities at the head and the terminus took a few more months to slow to quiescent values. Since then, average annual velocities along the centerline have been lower than pre-surge velocities, the crevasses have closed up, and the rate of ice surface elevation change has been negative across the entire glacier.

Fisher Glacier's surge dynamics suggest predominantly hydrologically controlled surging, but with some aspects more representative of thermally controlled surging. Thus, we propose that

more than one mechanism might be at play in controlling its surges, although further research is required to confirm this. Under current climate conditions, it is unlikely that Fisher Glacier could dam the nearby Alsek River and cause a glacier lake outburst flood.

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Preface

This thesis is written in hybrid format under the supervision of Dr. Luke Copland, with a plan to publish Chapter 2 as a paper in collaboration with numerous co-authors. Chapter 1 introduces the subject through a background and literature review, and details the thesis structure and research objectives. The article manuscript is presented in Chapter 2 (Reconstructing the surge history and dynamics of Fisher Glacier, Yukon, from 1948-2022). All references are presented together at the end and all tables and figures are presented directly within the manuscript text.

The manuscript (Chapter 2) is presented in a format ready for submission to a peer-reviewed journal such as Journal of Glaciology. The proposed title and authorship are:

Partington, G., Copland, L., Lauzon, B., Medrzycka, D., Main, B., Kochtitzky, W., Dow., C. (in prep). Reconstructing the surge history and dynamics of Fisher Glacier, Yukon, Canada, from 1948-2022.

I was the primary author, with input and suggestions from my thesis supervisor, Dr. Luke Copland. I was also responsible for collecting and analysing data, and processing most of the datasets in Chapter 2. In addition to helping with writing the initial manuscript and providing guidance throughout the project planning phase, Dr. Luke Copland also contributed through organising fieldwork logistics and planning, and provided *in situ* photographs of Fisher Glacier from August 2015. Benoît Lauzon assisted with troubleshooting and technical help at many steps throughout the data analysis process, and provided co-registered MMASTER DEMs and trend DEMs for Fisher Glacier. The MMASTER DEMs provided to Benoît were initially processed by Dr. William Kochtitzky who derived them from ASTER satellite imagery using the MMASTER Python package on the Canada Compute Cedar super computer. Dr. Dorota Medrzycka offered invaluable training expertise during fieldwork and guided me through the process of generating DEMs and orthophotos from historical air photographs. Brittany Main provided Radarsat-2 SAR winter velocity mosaics for Fisher Glacier. Finally, Dr. Christine Dow contributed through organising logistics and planning for fieldwork, and provided weather station data from Lowell Glacier.

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List of Abbreviations

AAR	Accumulation Area Ratio
a.s.l.	Above Sea Level
ASTER	Advanced Spaceborne Thermal Emission and Reflection Radiometer
Auto-RIFT	Autonomous Repeat Image Feature Tracking
AWS	Automatic Weather Station
CAA	Canadian Arctic Archipelago
CNES	Centre National d'Études Spatiales
CT	Central Terminus
DBL	Dynamic Balance Line
DEM	Digital Elevation Model
EDW	Elevation-Dependent Warming
EOS	Earth Observing System
GCP	Ground Control Point
GDEM	Global Digital Elevation Model
GTT	Glacier Termini Tracking
ITS_LIVE	Inter-mission Time Series of Land Ice Velocity and Elevation
KH-x	Keyhole satellite mission
MEaSURES	Making Earth Science Data Records for Use in Research Environments
MMASTER	MicMac ASTER
NAPL	National Air Photo Library
NASA	National Aeronautics and Space Administration
NOAA	National Oceanic and Atmospheric Administration
NT	Northern Terminus
PF	Pothole Field
RCAF	Royal Canadian Air Force
RGI	Randolph Glacier Inventory
RMSE	Root Mean Square Error
SAR	Synthetic Aperture Radar
SfM	Structure from Motion
SPOT	Satellite Pour l'Observation de la Terre
ST	Southern Terminus
TP	Tie Points
USGS	United States Geological Survey
UTM	Universal Transverse Mercator
w.e.	Water equivalent
WGS	World Geodetic System

Chapter 1: Background and research objectives

1.1 Background

Surge-type glaciers are characterized by dynamic instabilities that result in periodic alternating cycles of fast flow, termed the active phase, and slow flow, termed the quiescent phase (Meier & Post, 1969; Sevestre & Benn, 2015). The active phase generally lasts for <1 to ~5 years for a temperate glacier, and ~2 to 20 years for a polythermal glacier, and is much shorter than the quiescent phase, which can be as short as 5-10 years for a temperate glacier and as long as 50-500 years for a polythermal glacier (Meier & Post, 1969; Kochtitzky et al., 2019; Nolan et al., 2021). In addition, active phase velocities are usually at least an order of magnitude higher than quiescent flow speeds (Meier & Post, 1969), and quiescent phase velocities are almost always slower than velocities on neighbouring non-surge-type glaciers (Burgess et al., 2013).

Surge-type glaciers are composed of an ice reservoir where ice mass builds up during quiescence (usually indicated by an increasing glacier thickness over time) and an ice receiving area where the mass is transferred to during a surge (Meier & Post, 1969). It is important to note that these two areas typically do not correspond to the accumulation and ablation zones of a glacier (e.g., the ice reservoir and receiving areas can both be located in the ablation zone of a glacier) (Meier & Post, 1969; Kochtitzky et al., 2019).

Due to the difficulty and impracticality of inserting instruments at the glacier bed, and because surge periodicity can be very long, only a few well-documented examples of repeat surge events on a single glacier have been recorded (e.g., Bevington & Copland (2014) for Lowell Glacier, Yukon; Eisen et al. (2001 & 2005) for Variegated Glacier, Alaska; Abe et al. (2016) and Kochtitzky et al. (2019) for Donjek Glacier, Yukon; Nolan et al. (2021) for Turner Glacier, Alaska), making surge mechanisms difficult to determine, particularly outside of Yukon/Alaska. The main consensus is that two general mechanisms drive glacier surges: the thermal switch and the hydrologic switch (Kamb et al., 1985; Murray et al., 2003). Some surges seem to be exclusively driven by one or the other, whereas others seem to display characteristics of both mechanisms. Based on the locations where these mechanisms were first described and are often found, surges driven by the thermal switch are also termed 'Svalbard-type', and surges driven by the hydrologic switch are also termed 'Alaskan-type' (Abe et al., 2016). The following paragraphs review what each of these mechanisms propose.

The hydrologic switch is based on the linked cavity theory, whereby surges are initiated through a reorganization of subglacial meltwater channels (Kamb et al., 1985; Fowler, 1987). During quiescence, the subglacial drainage system consists of one (or a few) large tunnels which efficiently drain most of the water from surface and basal melt. These tunnels have a much lower water pressure than the ice overburden pressure but remain open by frictional heating from flowing water. This tunnel system eventually collapses due to the increasing mass of overlying ice which builds up during the quiescent phase due to the lack of ice motion, and which overwhelms the ability of flowing water to keep the subglacial tunnels open (Kamb et al., 1985). The subglacial drainage system then transitions to an inefficient system of small subglacial cavities linked by narrow passageways that remain open due to high water pressure rather than by frictional heating (Kamb et al., 1985). The high basal water pressure in the linked cavity system reduces basal shear stress and drives high basal sliding, which allows for rapid surface ice motion and therefore initiates a surge.

In contrast to the hydrologic switch, a surge mechanism controlled by a thermal switch has been proposed for surge-type glaciers in colder regions such as Svalbard (Jiskoot et al., 2000; Fowler et al., 2001; Murray et al., 2003). This hypothesis stipulates that changes in ice temperature near the bed initiate surges because glaciers with temperate beds can slide, whereas those with cold beds cannot. The positive feedback mechanism of the thermal switch over a soft bed works as follows (Fowler et al., 2001): (1) At the beginning of a surge cycle (i.e., at the end of a prior surge) the glacier is cold-based, at least at its margins. (2) Since the ice is frozen to the bed and velocities are extremely low, ice builds up in the reservoir zone and raises the pressure-melting point of ice at the base, and the thick ice insulates the bed from cold atmospheric temperatures. (3) As the bed warms and basal meltwater production increases, pore-water pressures reach near-flotation which reduces effective pressure and weakens the subglacial till layer and causes it to fail. (4) Basal motion increases as the till undergoes deformation, which also creates more basal water through friction. This positive feedback cycle causes the glacier to slide forward as it is no longer frozen to the bed. Basal motion causes glacier thinning, which results in heat loss and refreezing at the glacier bed which ultimately terminates the surge.

While some regions such as Yukon/Alaska tend to be heavily influenced by hydrological controls, and surges in areas like Svalbard and the Canadian High Arctic tend to be influenced by thermal controls (Jiskoot et al., 2000), there are numerous cases where these mechanisms do not

explain a glacier's surging behavior. It is now generally agreed upon that there is a spectrum of surging behaviors, and that many glaciers display a mix of both hydrologic and thermal mechanisms, and maybe other controls that have yet to be proposed.

Most recently, the Enthalpy Balance Theory has been proposed as a way of explaining glacier surging based on changes in energy content of a glacier, regardless of the exact mechanisms at play (Sevestre & Benn, 2015; Benn et al., 2019a; 2023). This theory stipulates that in order to maintain steady state, a glacier's ice discharge through any given cross-sectional area must: (a) evacuate exactly the mass gains from up glacier (balance velocity), and (b) maintain enthalpy, the energy flux into and out of the system. If this doesn't happen, then the glacier will undergo oscillations of fast and slow flow to return to a steady state (Sevestre & Benn, 2015). Under this model, low basal enthalpy production and efficient heat losses by conduction explain the scarcity of surge-type glaciers at the cold, dry end of the climatic spectrum (Sevestre & Benn, 2015), such as in Antarctica. In contrast, large glaciers in slightly warmer climates, such as in Arctic Canada and Svalbard, are more likely to surge as they cannot dissipate heat through conduction as their thickness provides more insulation. In the warmer end of the climate spectrum, such as in the European Alps, glaciers are likely temperate and rarely surge as they are able to discharge sufficient water to counter the excess energy from melting/radiation (Sevestre & Benn, 2015). Thus, in the optimal climate zone (mean annual temperature from -10°C to 0°C and mean annual precipitation from 165 mm yr^{-1} to 402 mm yr^{-1}), surge-type glaciers are common because *“neither heat conduction nor subglacial drainage systems are efficient enough to evacuate the enthalpy gains associated with their balance velocities, creating the possibility of heating-velocity feedbacks at the bed.”* (Sevestre & Benn, 2015, p. 657)

Benn et al.'s (2019a) theory suggests that surges are not solely caused by internal processes and can be affected by climatic fluctuations. According to Sevestre & Benn's (2015) results, climate seems to be a first-order, global control on the distribution of surge clusters, whereas glacier geometry is a second-order, regional control. In other words, climate dictates where surge-type glaciers are likely to be found, whereas glacier geometry influences which glaciers within these clusters are surge-type. Studies in Svalbard, Iceland, and the Karakoram have suggested that climate can affect the length of the quiescence period by altering glacier mass balance, thus influencing the geometric evolution that drives the surge cycle (Dowdeswell et al., 1995; Copland et al., 2011; Striberger et al., 2011). On a regional basis, positive mass balance seems to increase

surge frequency (Copland et al., 2011) while negative mass balance reduces it (Dowdeswell et al., 1995; Striberger et al., 2011). In support of this relationship, one theory stipulates that surges are initiated once an accumulation threshold is attained in the reservoir zone, since an increased slope affects the shear stress and the increase in overburden pressure can cause the ice to reach pressure melting point (Eisen et al., 2001). In such cases, climate can affect the accumulation rates, and therefore the surge recurrence interval. That being said, other studies reported no significant change in recurrence interval despite warming temperatures and sustained negative mass balance (Bevington & Copland, 2014; Kochtitzky et al., 2020a), proposing instead that glacier mass balance, which is controlled by climate, might control the spatial extent of each surge event, while other mechanisms are responsible for controlling the surge recurrence interval (Kochtitzky et al., 2020a).

Glacier geometry, here meaning area, length, slope, aspect, branchiness, and other environmental variables like lithology and bed type play a role in determining which glaciers are surge-type within a given region. Large area has been shown to be the most important variable in determining the likelihood of a glacier being surge-type, followed by long length and then shallow slope, all of which are somewhat correlated (Sevestre & Benn, 2015). In areas where both surging and non-surging glaciers are present, surging glaciers are almost always larger than non-surging ones (Jiskoot et al., 1998; Sevestre & Benn, 2015). In addition, glaciers with more branching also tend to be surge-type (Sevestre & Benn, 2015). Lithology and bedrock have also been identified as being significant contributors to the likelihood of a glacier being surge-type in Svalbard and the Yukon, although these variables were not significant in all areas (Jiskoot et al., 1998; Crompton et al., 2018). Ultimately, dynamic instabilities may be the result of an interplay between many variables, both internal and external.

1.2 Characteristics of glacier surging in Yukon/Alaska

The mountain ranges of Yukon/Alaska are home to many surge-type glaciers. Meier & Post (1969) identified 204 glaciers as being surge-type, all of which were located in the Alaska Range, the Chigmit, eastern Wrangell, eastern Chugach, and the St. Elias Mountains of Alaska, Yukon and British-Columbia. In the St. Elias Mountains specifically, 151 glaciers were classified as surge-type (Clarke et al., 1986). Decades later, and with access to better data, the number of surge-type glaciers in Yukon/Alaska as a whole was updated to 322 (Sevestre & Benn, 2015).

Only Svalbard has more confirmed surge-type glaciers than Yukon/Alaska (Sevestre & Benn, 2015). Some of the first detailed studies of glacier surges in Yukon/Alaska were carried out on Variegated Glacier in Alaska, from which the hydrologic switch surge mechanism was discovered (Kamb et al., 1985). This study took detailed *in situ* measurements that showed that water levels in boreholes drilled to the bed peaked at the same time as velocities during Variegated Glacier's 1982 surge. This lends itself to the theory that the glacier bed is subject to an inefficient drainage system, where high water pressures “float” the glacier off the bed and cause it to slide (Kamb et al., 1985).

From this study, and many others since, we can identify several behavioural properties that are now considered characteristic of Alaskan-type surges. For one, the hydrologic switch mechanism affects temperate glaciers, which have a warm bed. Active phases typically last 1-3 years and are followed by a 10-40 year-long quiescent phase (Meier & Post, 1969; Kamb et al., 1985; Lawson, 1997; Abe et al., 2016; Kochtitzky et al., 2019). Surges frequently initiate in the late autumn or winter (Kamb et al., 1985) and the initiation phase is rapid, with velocities increasing by at least an order of magnitude (i.e., the typical threshold when a surge is considered to have initiated, although it can be more or less than this). This 10-fold increase in velocities occurs in as little as a few days to a few weeks, and rarely exceeds a few months (Kochtitzky et al., 2019). With the hydrologic switch, the velocities often begin increasing in the mid- to upper region of the glacier, and subsequently move downglacier with a pronounced surge bulge (Kamb et al., 1985; Lawson, 1997). Surges frequently terminate in the late summer (Kamb et al., 1985; Lawson, 1997; Bevington & Copland, 2014; Abe et al., 2016) and the switch back to quiescence is usually abrupt and occurs more rapidly than the surge initiation (Murray et al., 2003). In addition, the surge brings about significant change to the glacier's geometrical properties as the downglacier transfer of mass leads to a thickening in the receiving zone and a corresponding thinning in the reservoir zone (Kamb et al., 1985; Kochtitzky et al., 2019). This transfer of mass typically results in a significant and rapid terminus advance and is usually accompanied by several morphological features such as extensive crevassing, serac formation, sheared margins, and looped moraines at the snout of tributary glaciers that feed into the main trunk or along medial moraines (Kamb et al., 1985; Copland et al., 2011; Bevington & Copland, 2014; Kochtitzky et al., 2019).

Although the characteristics described above are typically associated with Alaskan-type surging glaciers, they are by no means absolute criteria as most of these glaciers display a variety of behavioural exceptions which do not correspond to this list of characteristics.

For example, in Alaska, Variegated Glacier's surges are subject to short pulses/mini-surges during the surge buildup where velocities will rapidly increase within a few hours for a very short period of time (usually less than a day), before slowing back down again over a period of 10-20 hours (Kamb et al., 1985). For example, a rapid transition from 0.4 m d^{-1} to $1\text{-}3 \text{ m d}^{-1}$ over a period of a few hours and lasting less than a day was measured in 1979 during the buildup to Variegated Glacier's 1982 surge (Kamb et al., 1985). Its active phases are also characterized by oscillations of differing lengths (some lasting about 40 hours, others less than a day, and still others classified as slowdown oscillations) (Kamb et al., 1985).

Moving inland to the Yukon, Klutlan Glacier's surges seem to initiate in the central trunk and then propagate both downglacier (velocities of at least 4 m d^{-1}) as well as a little upglacier (Altena et al., 2019; Samsonov et al., 2021).

Walsh Glacier's flow dynamics during a surge display many of the typical characteristics of Alaskan-type surges as described above. However, it has an anomalously long surge recurrence interval of >40 years (Meier & Post, 1969; Samsonov et al., 2021). Additionally, the surge only influences the mid- and upper parts of the glacier, with the surge front ending at 20 km from the terminus, and no change in velocity or terminus advance observed below this point (Samsonov et al., 2021).

Lowell Glacier generally behaves like a "typical" Alaskan-type surging glacier. It has a surge duration of 1-2 years with a recurrence interval of 12-20 years (average 15.25 years) (Bevington & Copland, 2014). A detailed study on the glacier found that even with sustained negative mass balance, it shows no statistically significant trend in recurrence interval (although the recurrence interval does seem to be shortening slightly), but there has been a gradually decreasing terminus extent with each subsequent surge (Bevington & Copland, 2014). The magnitude of Lowell Glacier's surges therefore seems to be responding to changes in climate.

Donjek Glacier has relatively high quiescence velocities (typically these are slower on surge-type glaciers than on neighbouring non-surging glaciers, but this is not the case for Donjek Glacier) (Burgess et al., 2013). At the terminus, velocities are 0.5 m d^{-1} during quiescence and they increase to $2\text{-}4.5 \text{ m d}^{-1}$ during a surge (Abe et al., 2016; Kochtitzky et al., 2019). Thus, active phase

velocities are much slower than most surging glaciers in the region. Comparatively, many glaciers have virtually stagnant termini during quiescence and velocities of up to $10+ \text{ m d}^{-1}$ during the active phase (e.g., Variegated Glacier; Lawson, 1997; Lowell Glacier; Bevington & Copland, 2014; Tweedsmuir Glacier, Krimmel & Meier, 1975). Another particularity is that Donjek Glacier is 55 km long, but its surges only affect the lower 21 km of the glacier (Abe et al., 2016; Kochtitzky et al., 2019). Like Lowell Glacier, Donjek Glacier also shows a trend of a decreasing magnitude of surge extent with each successive surge (Kochtitzky et al., 2019). Another unusual characteristic is that most of its surges initiated in summer, which is more typical of polythermal glaciers (Kochtitzky et al., 2019; 2020b). Finally, Donjek Glacier also stands out as one of the surge-type glaciers in the St. Elias Mountains with the smallest amounts of mass moved during each surge (after Trapridge Glacier), and shortest recurrence interval (12 years) (Kochtitzky et al., 2019).

Back in Alaska, Turner Glacier is a 30-km long tidewater surging glacier with a 2-year active phase length and an extremely unusual 5-year recurrence interval (Nolan et al., 2021). It is characterized by high velocities, with $<1 \text{ m d}^{-1}$ during quiescence and up to 26 m d^{-1} during the active phase (Nolan et al., 2021).

Although hydrological controls are the most common influence on glacier surges in the St. Elias Mountains, there are a few cases of polythermal glaciers displaying “slow surges” that are believed to be thermally controlled (Clarke, 1976; Frappé et al., 2007; De Paoli & Flowers, 2009). For example, Trapridge Glacier, Yukon, is a small polythermal glacier with a warm-bedded central base bounded by cold ice (Fowler et al., 2001). It last surged from ~1980 to ~2000, thus displaying a 20-year long active phase where velocities peaked at 42 m yr^{-1} (0.12 m d^{-1}), before slowing back down to 9 m yr^{-1} (0.025 m d^{-1}) by 2005 (Frappé et al., 2007). It had last surged from sometime before 1939 to ~1951 (Frappé et al., 2007), therefore suggesting a ~30-year repeat interval (De Paoli & Flowers, 2009). Although its quiescent phase length is not unheard of in hydrologically controlled surges, its active phase dynamics are what set it apart from the other glaciers mentioned above, and are more similar to thermally controlled surges observed in locations such as Svalbard and the Canadian Arctic (e.g., Dowdeswell et al., 1991; Jiskoot et al., 2000; Van Wychen et al., 2022; Lauzon et al., 2023).

Similarly, North Glacier (unofficial name), a small glacier in a tributary valley of Kaskawulsh Glacier also seems to undergo slow surges (De Paoli & Flowers, 2009). An analysis of historical air photographs showed that this small glacier has a surge periodicity of 30-40 years.

Active phase velocities in the central region of the glacier were measured at around 20-35 m yr⁻¹ (0.05-0.10 m d⁻¹) and are accompanied by morphological features indicative of surging, notably an undulating surface and crevassing (De Paoli & Flowers, 2009). At the time of the study, the lower region had not started moving significantly, although terminus advance has been documented in previous surges through historical air photos (De Paoli & Flowers, 2009). It is unclear whether this is a typical surge for this glacier, or if changes in climate (and therefore mass accumulation and glacier geometry) are affecting its dynamics in the most recent surge (De Paoli & Flowers, 2009).

Evidently, a variety of surge mechanisms are at play in the St. Elias Mountains. Even within the broad category of Alaskan-type surging glaciers, there is an incredible diversity in their surge dynamics. This highlights the importance of in-depth studies on individual glaciers, which allow for a better understanding of glacier surging in the region and provides greater insight than what can be learned from regional- or global-scale studies. It also calls attention to the argument that dynamic instabilities occur over a wide spectrum, and that the word “surge” and its “Alaskan-type” versus “Svalbard-type” subcategories may be too simplistic in classifying surge-type glaciers (Van Wychen et al., 2021; 2022; Benn et al., 2023).

1.3 Research Objectives

The primary objective of my thesis is to reconstruct the surge history of Fisher Glacier and determine the controls on its surges. This is done by: (a) Measuring the frequency, duration, and location of past surges, and (b) Assessing controls on the surge cycle (e.g., potential water inputs, inferred thermal conditions, surge velocity patterns) to determine which factors control surging.

My study is focused on Fisher Glacier as it is one of the last large surge-type outlet glaciers on the eastern side of the St. Elias Icefield that has yet to be studied in detail. An analysis of multi-decadal changes in this glacier can therefore provide important information to better characterize glacier dynamics in this region, better understand surging dynamics in Yukon/Alaska, and better understand the impact of land-terminating glaciers on Yukon’s contribution to sea level change.

I accomplish my objectives using primarily remotely sensed data, including air photos from the 1940s onwards, declassified intelligence satellite images since the 1960s, optical satellite images since the 1970s, digital elevation models, and glacier velocity datasets derived from both optical and SAR imagery. I use these to determine changes in terminus extent, ice surface

velocities, surface elevation, and the identification of surface features that are typically observed on surge-type glaciers. Together, this information is used to identify when surges have occurred at Fisher Glacier over the past ~75 years and provide information concerning Fisher Glacier's behavior and evolution through each surge and the subsequent quiescent period.

Chapter 2: Reconstructing the surge history and dynamics of Fisher Glacier, Yukon, 1948-2022

2.1 Introduction

Surge-type glaciers are characterized by semi-regular alternations between active phases (fast motion of up to $1000+ \text{ m yr}^{-1}$) and quiescent phases (slow motion, occasionally $<10 \text{ m yr}^{-1}$) (Meier & Post, 1969; Cuffey and Paterson, 2010). We know that many glaciers in the Yukon display surge-type behaviours, particularly those found in the St. Elias Mountains along the Yukon-Alaska border (Meier & Post, 1969; Clarke and Holdsworth, 2002; Sevestre & Benn, 2015). For example, studies on Lowell Glacier have shown that it surges regularly, with a recurrence interval of 12 to 20 years and active phases lasting approximately 1 to 2 years (Bevington & Copland, 2014). Further north, Donjek Glacier also demonstrates regular surge behavior, with a repeat interval of about 12 years and active phases lasting ~ 2 years (Abe et al., 2016; Kochtitzky et al., 2019). Fisher Glacier is another large surge-type outlet glacier that is located on the eastern side of the St. Elias Icefield, to the south of Lowell Glacier. However, no detailed studies have focused on Fisher Glacier's surge behavior, despite evidence that it surged as recently as 2016.

Surges, and the accompanying terminus advance that frequently ensues, can have significant impacts on the hydrology of the surrounding environment, which can in turn affect both the built (e.g., nearby roads, bridges, and towns) and natural (e.g., fish migration and spawning habitats) environments. In southwestern Yukon, past advances of Lowell Glacier have repeatedly dammed the Alsek River and created Neoglacial Lake Alsek, with the most recent lake forming around 1900 (Clague & Rampton, 1982). The blockage of the river by Lowell Glacier created a glacier-dammed lake that flooded the surrounding area, including the current townsite of Haines Junction (Clague & Rampton, 1982). When it retreated after the active phase, a sudden release of water resulted in a glacier lake outburst flood which caused damage to infrastructure downstream and loss of life (Cruikshank, 1981). Presently, the Alaska Highway crosses the Alsek River and would be at risk of damage if a glacier-dammed lake were to form again (Post & Mayo, 1971). To the south, Tweedsmuir Glacier has also been reported to have blocked the Alsek River and created a large glacier-dammed lake (Post & Mayo, 1971; Clarke & Holdsworth, 2002). Being located between Lowell Glacier and Tweedsmuir Glacier, Fisher Glacier also has the potential to impede the flow of the Alsek River if a significant surge event were to occur (Clarke & Holdsworth, 2002).

Cruikshank (1981) mentions a “Glacial Lake Fisher” which formed after the last glacial maximum ~12,500 years ago, although no information about it, nor any other lakes created by Fisher Glacier in the recent historical past, was passed down through First Nations oral tradition (Cruikshank, 1981).

The objectives of this research are: (a) to reconstruct the surge history of Fisher Glacier by measuring the frequency, duration, and location of past surges, and (b) to assess controls on the surge cycle (e.g., using potential water inputs, inferred thermal conditions, ice velocity patterns) to determine which factors control surging, and how these may change over time in a warming climate. This is undertaken using a variety of remote sensing datasets and field observations from the past ~75 years, including historical air photos, satellite images, digital elevations models and surface velocities, as described in detail below.

2.2 Study Area

2.2.1 St. Elias Mountains

The St. Elias Mountains straddle the border between southeastern Alaska, southwestern Yukon, and northwestern British Columbia (Meier & Post, 1969; Williamson et al., 2020) (Figure 1). Glaciers in this region were first studied in detail during the Icefield Ranges Research Project in the 1960s (Wood, 1963). This region is known for its high density of surge-type glaciers (Meier & Post, 1969; Clarke et al., 1986; Sevestre & Benn, 2015), and has been characterized as one of the largest surging glacier clusters in the world (Sevestre & Benn, 2015; Benn et al., 2019a). Many of its large outlet glaciers are characterized by ‘Alaskan-type’ (hydrologically controlled) surges, with relatively short active and quiescent phases (Clarke et al., 1986; Bevington & Copland, 2014; Kochtitzky et al., 2019), whereas some smaller tributary glaciers seem to demonstrate ‘Svalbard-type’ (thermally controlled) surge characteristics such as long active phases and slow surge initiation and termination (Frappé & Clarke, 2007; De Paoli & Flowers, 2009; Flowers et al., 2011).

The total glacierized area in northwestern North America (here meaning Alaska and northwest Canada) is 87,860 km² (Berthier et al., 2010). Of this, the glaciers of the St. Elias Mountains cover 32,900 km², with glacier elevations varying from 0 to 5930 m above sea level (a.s.l.), with a mean of 1600 m a.s.l. (Arendt et al., 2008). The St. Elias Mountains therefore contain

~38% of the ice-covered area in the Yukon/Alaska region and form the largest non-polar icefield in the world (Young et al., 2021).

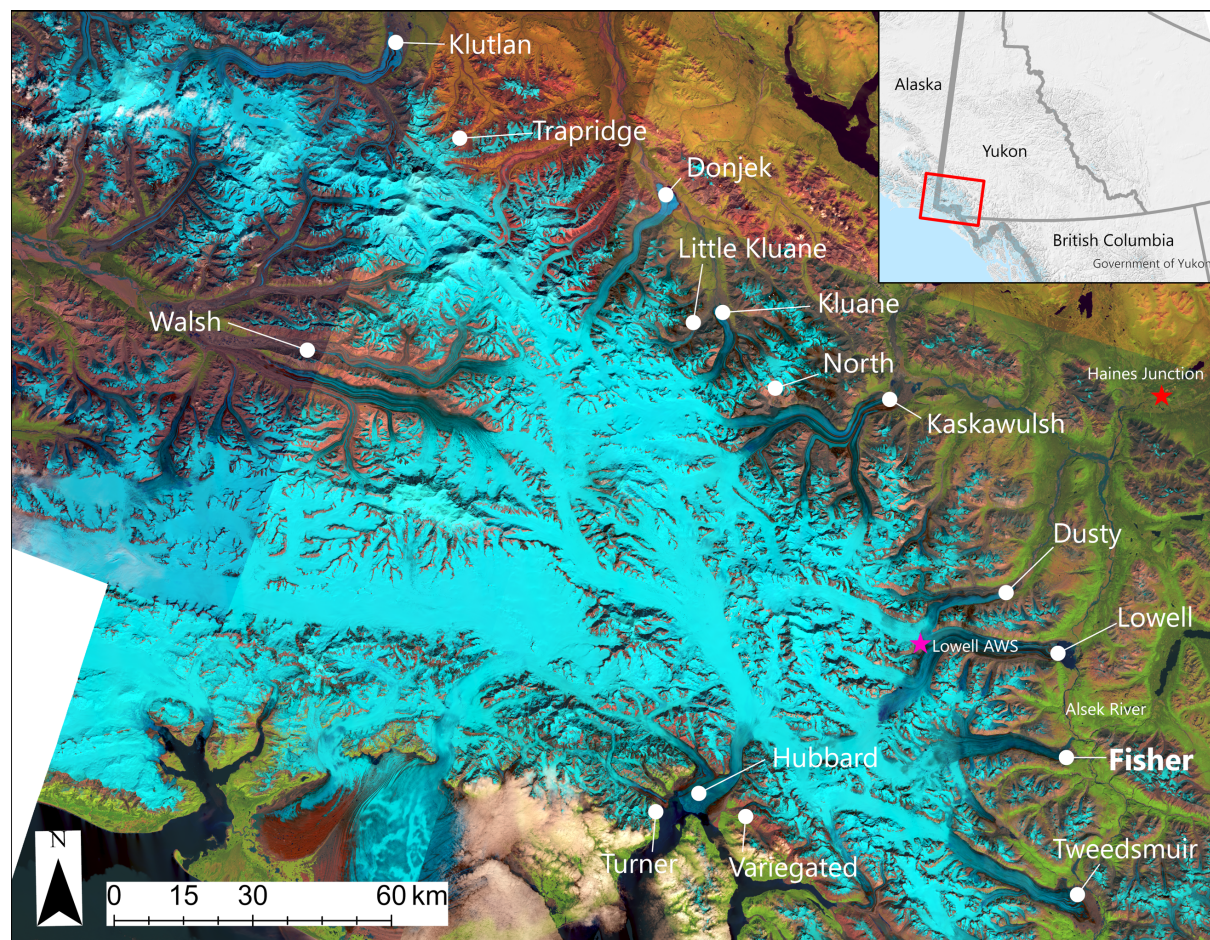


Figure 1. The St. Elias Mountain range showing the location of Fisher Glacier (bold) and other glaciers referred to in the text. The location of the weather stations located in Haines Junction (red star) and on the nunatak between Dusty Glacier and Lowell Glacier (pink star) are also shown. Background is a mosaic of three Landsat 8 images (8 August 2017, 2 August 2022) and two Landsat 9 images (4 July 2022, 12 August 2022), all downloaded from USGS EarthExplorer. Inset from GeoYukon.

Both surge-type and normal (non surge-type) glaciers in northwestern North America have undergone recent changes, notably in thinning and mass balance (Arendt et al., 2008; Berthier et al., 2010; Box et al., 2018; Ciraci et al., 2020; Hugonnet et al., 2021). In fact, the St. Elias and Wrangell Mountains sustained ice loss of $21.66 \pm 4.4 \text{ km}^3 \text{ yr}^{-1}$ w.e. and had an average mass balance of $-0.47 \pm 0.09 \text{ m yr}^{-1}$ w.e. from 1962-2006 (Berthier et al., 2010). A separate study found that the St. Elias Mountains showed a negative mass balance trend of $-20.6 \pm 3.0 \text{ Gt yr}^{-1}$ from

2003-2007 (Arendt et al., 2008). The calving flux from tidewater (ocean-terminating) and lacustrine (lake-terminating) glaciers is a significant contributor to ice loss in this region, corresponding to 36% of the annual mass change throughout the Wrangell-St. Elias Mountains (Burgess et al., 2013). Like the St. Elias Mountains, the Yukon/Alaska region as a whole showed a similar negative trend of $-72.5 \pm 8 \text{ Gt yr}^{-1}$ from 2002-2019 (Ciraci et al., 2020).

The St. Elias Mountains' annual air surface temperatures have been increasing since 1979 at all elevations from 2000-6000 m, with statistically significant elevation-dependent warming (EDW) at elevations of $>3000 \text{ m}$ (Williamson et al., 2020). In their paper, Williamson et al. (2020) found a 1.5 amplification factor in the warming rate between 2000-2500 m and 5500-6000 m elevation bins ($0.019^\circ\text{C yr}^{-1}$ and $0.028^\circ\text{C yr}^{-1}$, respectively). In addition, the maximal rate of warming in the St. Elias was 1.6x larger than the global-mean warming rate found by NOAA ($0.028^\circ\text{C yr}^{-1}$ vs. $0.017^\circ\text{C yr}^{-1}$) for 1970-2015 (Williamson et al., 2020). While longwave water vapor feedback appears to be a primary driver of EDW in the St. Elias Mountains, it is unclear what other factors are also significant, such as from snow albedo feedback, aerosol radiative forcing, clouds, and tropospheric warming (Williamson et al., 2020).

Although each study shows slightly differing values stemming from differences in time period, variations in methodology and corrections, and differences in spatial coverage, there is consensus that the Arctic has been a significant contributor to sea level rise throughout the 20th and 21st centuries. Notably, southern Arctic regions such as Alaska and the Yukon are undergoing the most negative mass changes and highest thinning rates and, as a result, are currently the greatest contributors to sea level rise, outside of the Greenland and Antarctic ice sheets (Box et al., 2018; Ciraci et al., 2020; Hugonnet et al., 2021). The negative mass balance rates measured at regional and global scales are supported by many studies on individual glaciers which agree that glaciers in the area have sustained negative mass balance trends over the past several years (e.g., Foy et al. (2011) and Young et al. (2021) for Kaskawulsh Glacier; Kochtitzky et al. (2019) for Donjek Glacier).

2.2.2 Fisher Glacier

Fisher Glacier is a large valley glacier located in Kluane National Park in the southeastern portion of the St. Elias Mountains ($60^\circ04'56'' \text{ N}$, $138^\circ03'46'' \text{ W}$) (Figure 1). It is approximately 45 km long, an average of $\sim 3.8 \text{ km}$ wide, and its main trunk covers an area of $\sim 230 \text{ km}^2$, although it

drains an area of $>400 \text{ km}^2$ when considering the many small tributaries that feed into it (Figure 2a). Its terminus is located around 500 m a.s.l., and it drains eastward into the Alsek River, which flows mainly south and ends in the Gulf of Alaska. The top of the accumulation zone extends up to an elevation of $>1500 \text{ m}$ and it shares an accumulation zone with Lowell (Nàhüdäy) Glacier, which flows parallel to Fisher Glacier $\sim 20 \text{ km}$ to the north (Figures 1 and 2). It has an average surface slope of $\sim 1.3^\circ$.

The inland, eastern side of the St. Elias Mountain range is characterized by a subarctic climate regime due to low temperatures caused by the pronounced profile of the mountain range, which blocks warm and humid air from the Pacific coast (Marcus & Ragle, 1970; Williamson et al., 2020). These low winter temperatures have allowed large glaciers formed during the Wisconsin Glaciation to persist in all of the main valleys in this region (Cruikshank, 1981). The mean annual air temperature measured at Haines Junction from 1944-2023 is -2.4°C , and the region received an average precipitation of 154 mm yr^{-1} throughout this time. The Haines Junction weather station ($60^\circ 46' 21'' \text{ N}$, $137^\circ 34' 49'' \text{ W}$) is operated by Environment and Climate Change Canada and is located at an elevation of 595 m a.s.l. (i.e., similar to Fisher Glacier's terminus), approximately 80 km to the NNE (red star on Figure 1). A weather station located on a nunatak at 1248 m a.s.l. on Lowell Glacier ($60^\circ 18' 45'' \text{ N}$, $138^\circ 33' 36'' \text{ W}$; pink star on Figure 1) recorded a mean annual air temperature of -4.1°C over 2019-2023.

Previous studies on Fisher Glacier are limited. One of the earliest mentions of the glacier is by Clarke & Holdsworth (2002, page J312), who state that “*Fisher Glacier [...] surged around 1970, but little information about this event is available.*”, but provide no further details. Burgess et al. (2013) provide a velocity map which shows winter velocities of around $0.1\text{-}0.3 \text{ m d}^{-1}$ from 2007-2011, while Van Wychen et al. (2018) show similar winter velocities of $\sim 40 \text{ m yr}^{-1}$ in 2007-08 and $\sim 60 \text{ m yr}^{-1}$ in 2009-10 (0.11 and 0.16 m d^{-1} , respectively). More recently, Altena et al. (2019) mentioned that an increase in speed was observed on Fisher Glacier from 2013-2018, reaching velocities of about 4 m d^{-1} , suggesting that a surge occurred sometime during this period.

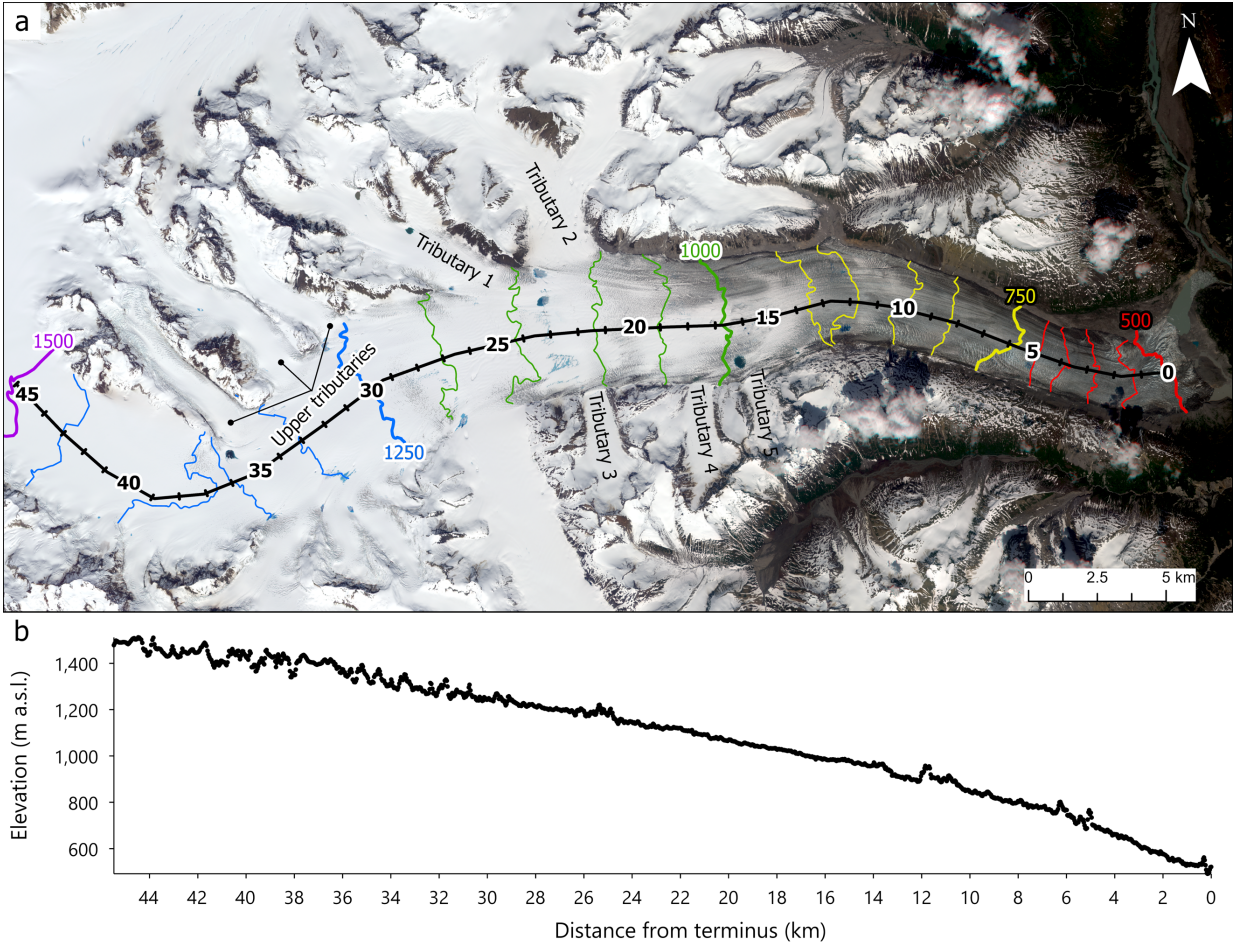


Figure 2. (a) Overview of Fisher Glacier showing some of the main features discussed in the text. Black line marks the centerline, with kilometer markers indicating distance from the 2015 terminus. Coloured contour lines represent 50 m a.s.l. hypsometry bands on the main glacier trunk, with each colour representing a 250 m elevation range. Base image: RapidEye-1, 2016-06-22. (b) Profile of the glacier surface along the centerline, using elevations extracted from the ASTER GDEM v3.

2.3 Methods

2.3.1 Changes in terminus extent

Although not all surges are accompanied by a significant terminus advance (e.g., Walsh Glacier, Alaska; Fu & Zhou, 2020), rapid changes in terminus extent are typically considered as one of the main indicators of glacier surging (Meier & Post, 1969; Copland et al., 2011). Terminus changes were measured using the *Glacier Termini Tracking* (GTT) toolbox in ArcGIS Pro (Urbanski, 2018). The GTT toolbox uses digitized terminus extents to determine two-dimensional rates of change by measuring the distance between points along the terminus outline in subsequent

extents (Urbanski, 2018). The tool was run over two time periods: first on an annual scale for the entire study period, and then on an approximately monthly scale (at least for the summer months) from 2013-2017 to cover a period of rapid terminus advance. This tool uses the date information contained in the attribute fields to determine the time elapsed between each terminus extent, and calculates terminus change in m month^{-1} . Terminus changes were later converted to m yr^{-1} by multiplying the GTT values by 12. The tool automatically adds points along each terminus line at a user-designated interval. From these points, the tool can quantify the mean changes in terminus position for each year by calculating the mean point separation distance along the entire terminus (Figure 3a). The point interval (distance between points along a line) was set at 100 m, which provides sufficient detail for Fisher Glacier's terminus width of >3 km. The results of the GTT two-dimensional analysis of terminus change were then run through the *Statistics for Glacier Termini Changes Tool* to obtain various statistics such as max, min, and mean monthly retreat rates and their standard deviation.

This series of steps was repeated four times: once for the entire length of the terminus to get an idea of total mean retreat, and then once for each of three subsections of the terminus titled the Northern Terminus, Central Terminus, and Southern Terminus (Figure 3b). For the GTT analysis on the terminus subsections, the point interval was set at 50 m. These three subsections were chosen because the ice flows differently in each one, which correspondingly displayed very different rates of change.

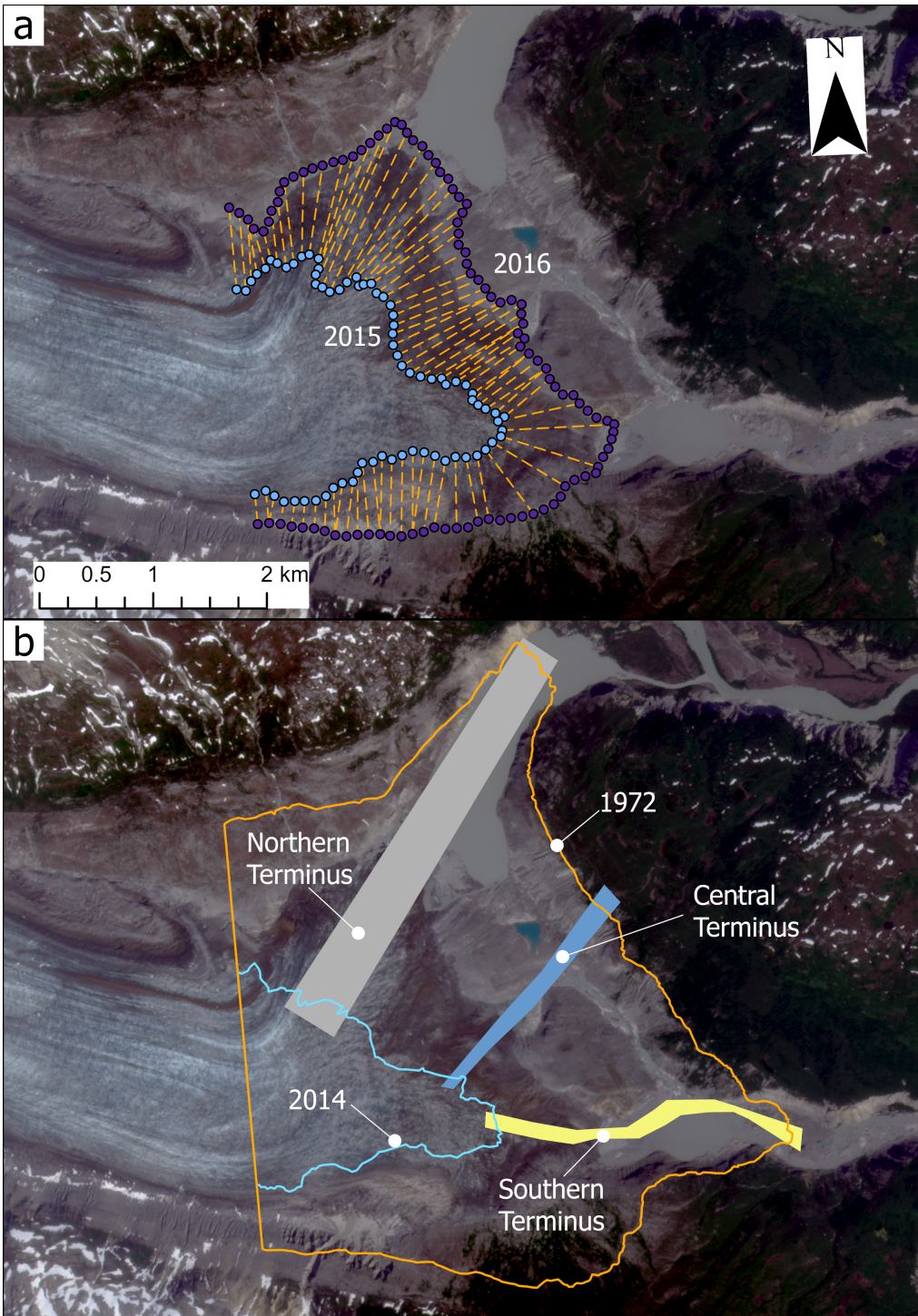


Figure 3. (a) Example of the GTT tool's output for change between the 2015 and 2016 terminus outlines; the average length of the orange dashed lines is measured to obtain the mean terminus-wide change. (b) The three subsections of the terminus on which the GTT tool was run, along with two examples of terminus polygons, from 1972 and 2014. The straight orange line indicates the common line to which all terminus extents were clipped. Base image: PlanetScope, 2022-07-04.

For this project, terminus extents and retreat rates were quantified for the period 1948-2022 for the years when satellite imagery, historical air photographs or declassified intelligence satellite imagery were available (Table 1). Shorter-term (sub-annual) terminus change rates were also calculated using the GTT toolbox from 13 August 2013 to 6 August 2017 to provide greater temporal resolution during a period of rapid terminus advance (Table 2). The upglacier boundary of these terminus extents was clipped to a common line in the ablation zone, located ~3.4 km upglacier from the maximum terminus extent, past which the terminus never retreated (Figure 3b).

Terminus extents for Fisher Glacier were manually delineated and digitized in ArcGIS Pro for 1948-2022 from a total of 54 images (Table 1). For the pre-satellite era (<1972), historical air photographs, taken by the Royal Canadian Air Force (RCAF), were acquired from the National Air Photo Library in Ottawa, Ontario and the Yukon Energy, Mines & Resources Library in Whitehorse, Yukon. Air photo surveys of Fisher Glacier were not conducted regularly during the pre-satellite era, so there are large gaps in the record, particularly in the 1960s. The 1948, 1954, 1972, and 1976 termini were mapped from nadir air photographs taken by the RCAF from a cruising altitude of 20,000 ft (6096 m), 10,500 ft (~3200 m), 38,358 ft (11,692 m), and 40,000 ft (12,192 m) above sea level, respectively.

For the satellite era, a summer (May-September) cloud-free optical satellite image was acquired for each year from 1973-2022. Images from July to September were prioritized when available to ensure consistent temporal separation between images and minimal snow cover. Terminus outlines from 1963 and 1977 were delineated from two declassified satellite images from the Keyhole satellite system's KH-5 Argon and KH-9 Hexagon projects, respectively, (<https://www.usgs.gov/centers/eros/science/usgs-eros-archive-declassified-data-declassified-satellite-imagery-3>), obtained from the United States Geological Survey (USGS). The 1977 image was manually georeferenced to a 3 m resolution PlanetScope image from 4 July 2022 with 8 ground control points (GCPs) and an affine transformation. The 1963 intelligence satellite image from the KH-5 Argon project was manually georeferenced in the same way as the KH-9 image, this one with 9 GCPs and an affine transformation.

Table 1. Satellite imagery and historical air photographs used for the delineation of annual terminus extents from 1948-2022.

Acquisition date	Image ID	Platform	Ground resolution (m)
1948-06-12	A11521	RCAF	1.47
1954-08-30	A14459	RCAF	2.62
1963-08-29	DS09058A032MC026_a	KH-5	140.2
1972-08-08	A23001	RCAF	3.06
1973-07-22	LM01_L1TP_066018_19730722_20200909_02_T2	Landsat 1	60
1974-08-21	LM01_L1GS_065018_19740821_20200908_02_T2	Landsat 1	60
1975-09-22	LM01_L1TP_066018_19750922_20200908_02_T2	Landsat 1	60
1976-07-24	A24520	RCAF	3.5
1977-09-03	DZB1213-500139L002001	KH-9	7.5
1978-09-24	LM03_L1TP_066018_19780924_20200906_02_T2	Landsat 3	60
1979-08-24	LM02_L1TP_067018_19790824_20200906_02_T2	Landsat 2	60
1980-08-17	LM02_L1TP_066018_19800817_20200905_02_T2	Landsat 2	60
1981-08-29	LM02_L1GS_065018_19810829_20200905_02_T2	Landsat 4	60
1982-09-17	LM04_L1TP_061018_19820917_20200904_02_T2	Landsat 4	60
1983-08-28	LM04_L1TP_060018_19830828_20200903_02_T2	Landsat 4	60
1984-07-28	LT05_L2SP_061018_19840728_20200918_02_T1	Landsat 5	30
1985-07-15	LT05_L2SP_061018_19850715_20200918_02_T1	Landsat 5	30
1986-07-18	LT05_L2SP_061018_19860718_20200917_02_T1	Landsat 5	30
1987-08-22	LT05_L2SP_061018_19870822_20201014_02_T1	Landsat 5	30
1988-06-21	14772278806212014541P1	SPOT1	10
1989-08-14	14772278908142058142X1	SPOT1	20
1990-08-11	24772279008112047231X1_1T	SPOT2	20
1991-06-26	24772279106262012222P1	SPOT2	10
1992-08-20	24772279208202017401P1	SPOT2	10
1993-09-02	24772279309022047302P1_1T	SPOT2	10
1994-09-04	34772279409042046282P1	SPOT3	10
1995-09-04	34772279509042027291P1	SPOT3	10
1996-08-17	34772279608172033311P1	SPOT3	10
1997-06-30	LT05_L2SP_061018_19970630_20200910_02_T1	Landsat-5	30
1998-08-04	LT05_L2SP_061018_19980804_20200908_02_T1	Landsat-5	30
1999-07-14	LE07_L2SP_061018_19990714_20200918_02_T1	Landsat-7	15
2000-08-10	LE07_L2SP_060018_20000810_20200918_02_T1	Landsat-7	15
2001-08-10	001-004_S4_477-226-9_2001-08-10-21-09-47_HRVIR-1_I_MX_TT	SPOT4	20
2002-06-13	LE07_L2SP_060018_20020613_20200916_02_T1	Landsat-7	15
2003-09-11	AST_L1T_00309112003204125_20150430201243_93084	ASTER	15
2004-08-23	004-005_S5_477-227-0_2004-08-23-20-16-48_HRG-1_J_MX_PQ	SPOT5	10
2005-08-16	001-004_S5_477-226-6_2005-08-16-20-29-23_HRG-2_J_DT_PQ	SPOT5	10
2006-06-24	002-004_S4_477-227-0_2006-06-24-20-58-37_HRVIR-2_I_DT_LM	SPOT4	20
2007-06-18	LE07_L2SP_061018_20070618_20200913_02_T1	Landsat-7	15 (30)
2008-06-13	LE07_L2SP_060018_20080613_20200912_02_T1	Landsat-7	15 (30)

2009-09-07	001-005_S4_477-226-8_2009-09-07-20-30-01_HRVIR-2_I_MX_TT	SPOT4	20
2010-09-15	766821_2010-09-15_RE4_3A	RapidEye-4	5
2011-08-30	766821_2011-08-30_RE5_3A	RapidEye-5	5
2012-08-30	766821_2012-08-30_RE4_3A	RapidEye-4	5
2013-08-13	766821_2013-08-13_RE4_3A	RapidEye-4	5
2014-07-31	766821_2014-07-31_RE3_3A	RapidEye-3	5
2015-07-25	766821_2015-07-25_RE1_3A	RapidEye-1	5
2016-09-05	20160905_193702_0e0e_3B	PlanetScope	3
2017-08-06	20170806_194305_1036_3B	PlanetScope	3
2018-07-21	20180721_195627_100c_3B	PlanetScope	3
2019-08-30	20190830_195625_0e2f_3B	PlanetScope	3
2020-07-03	20200703_201034_1006_3B	PlanetScope	3
2021-07-30	20210730_195235_33_2233_3B	PlanetScope	3
2022-07-04	20220704_192917_05_2421_3B	PlanetScope	3

Landsat 1-4 Multispectral Scanner (MSS) images and Landsat 5 Thematic Mapper (TM) images were obtained from USGS EarthExplorer to cover the period from 1973 to 1987, for the most part as precision and terrain corrected files, which means that no further georeferencing or pre-processing was required. Landsat 4 TM images were not available for Fisher Glacier during the study period, so Landsat 4 MSS images were used instead to cover the period 1981-1983. Historical SPOT 1-3 imagery from 1988-1996 were also downloaded from the USGS EarthExplorer, while SPOT 4-5 images from 2001-2009 were downloaded from CNES's Spot World Heritage Programme (<https://regards.cnes.fr/user/swh/modules/60>). Most of the SPOT images were downloaded as L1A products, so they required slight geometrical corrections. In such cases, the *Shift* function in ArcGIS Pro's georeferencing menu was used to align the terminus area of the SPOT image with a recent PlanetScope or RapidEye image. Landsat 5 and 7 images (downloaded from USGS EarthExplorer) and one ASTER scene (downloaded from NASA Earthdata search portal) were used to fill the gaps up to 2009 when no SPOT or other higher resolution images were available. RapidEye 1-4 and PlanetScope images were acquired courtesy of Planet Images, through their Education and Research Program, and covered the period 2010-2022.

For a period of rapid terminus advance around 2014-2016, and the years shortly before and after it, additional optical satellite imagery was used to quantify the short-term evolution in terminus position. In addition to the RapidEye and PlanetScope satellites, this included imagery from Landsat 7 and 8, which were obtained from the USGS EarthExplorer. Ultimately 16 optical satellite images from May to October were used for this period (Table 2).

To ensure consistency, all termini were delineated manually by the same individual, to the nearest pixel. This ensures that in cases where the true extent of the ice may be difficult to discern (e.g., dead ice interpreted to be glacier ice, or instances where the glacier surface is debris-covered and may be mistaken for bare ground), the choice of terminus position is consistent across the images and thus relative change can still be considered. In addition, as older images have poorer ground resolution (e.g., 60 m for Landsat 1 compared to 3 m for PlanetScope), the accuracy of older terminus outlines generally decreases, and uncertainties increase. False colour composites were created from the lower resolution Landsat imagery (i.e., Landsat 5 and earlier) to better distinguish the ice from surrounding terrain. The bands used in the red, green and blue channels, respectively, to produce these composites were bands 5, 4, 3 for Landsat 5 TM, bands 4, 3, 2 for Landsat 4 MSS, and bands 7, 6, 5 for Landsat 1 and 2 MSS.

Table 2. Satellite imagery used for the delineation of sub-annual terminus extents from 2013-2017.

Acquisition date	Image ID	Platform	Ground resolution (m)
2013-08-13	766821_2013-08-13_RE4_3A	RapidEye-4	5
2014-06-16	766821_2014-06-16_RE1_3A	RapidEye-3	5
2014-07-31	766821_2014-07-31_RE3_3A	RapidEye-3	5
2014-09-02	766821_2014-09-02_RE3_3A	RapidEye-3	5
2014-09-25	LE07_L1TP_061018_20140925_20200905_02_T1	Landsat 7	15
2015-05-21	766821_2015-05-21_RE3_3A	RapidEye-3	5
2015-06-23	766821_2015-06-23_RE2_3A	RapidEye-2	5
2015-07-25	766821_2015-07-25_RE1_3A	RapidEye-1	5
2015-09-13	LC08_L1TP_060018_20150913_20200908_02_T1	Landsat 8	15
2015-10-04	766821_2015-10-04_RE1_3A	RapidEye-1	5
2016-05-10	LC08_L1TP_060018_20160510_20200907_02_T1	Landsat 8	15
2016-06-10	LE07_L1TP_061018_20160610_20200902_02_T1	Landsat 7	15
2016-07-18	766821_2016-07-18_RE3_3A	RapidEye-3	5
2016-08-21	LC08_L1TP_061018_20160821_20200906_02_T1	Landsat 8	15
2016-10-01	766821_2016-10-01_RE2_3A	RapidEye-2	5
2017-08-06	20170806_194305_1036_3B	PlanetScope	3

Uncertainties associated with the terminus change rates are dependent on the ground resolution of the imagery that was used. A conservative maximum uncertainty of ± 2 pixels was considered to be an appropriate uncertainty estimate for terminus delineation (Kochtitzky et al., 2019; Lauzon et al., 2023), assuming that errors resulting from manual terminus delineation are random. This corresponds to 120 m for Landsat 1-4, 60 m for Landsat 5, 30 m for Landsat 7 (band 8 with 15 m resolution was used), Landsat 8 (panchromatic band) and ASTER. 20 m or 40 m for

SPOT 1-5 (depending on the image), 10 m for RapidEye 1-5, 6 m for PlanetScope, 15 m for the 1977 KH-9 image, and ~280 m for the 1963 KH-5 image. For the historical air photos, rolls A11521, A14459, A23001 and A24520 had uncertainties of 2.94 m, 5.24 m, 6.12 m and 7 m, respectively.

Uncertainties for the advance and retreat calculated by the GTT toolbox were calculated using the following formula:

$$Uncertainty = \frac{\sqrt{(pixel\ length)^2 + (pixel\ width)^2}}{number\ of\ years} \quad (1)$$

The pixel length and width in the uncertainty formula above were calculated as the mean pixel size for both images involved in the calculation. For instance, the uncertainty for the rate of change in terminus length between 1972 (RCAF; 3.06 m resolution) and 1973 (Landsat 1; 60 m resolution) was calculated with a pixel length/width of: $(3.06\ m + 60\ m) / 2 = 31.53\ m$.

In addition, the manual georectification error associated with historical air photographs and intelligence satellite imagery was added as a numerator in Equation 1. This error was calculated by measuring the offset between stable ground features located near the terminus on the air photos or intelligence imagery and a well georectified modern satellite image (i.e., RapidEye or PlanetScope since 2010). Following the example above, the additional georectification error for the 1972 RCAF air photo was 60 m (no additional error was calculated for the 1973 Landsat image since it was not manually georeferenced). Thus, uncertainty for the terminus change from 1972-1973 was calculated as:

$$Uncertainty = \frac{\sqrt{(31.53)^2 + (31.53)^2} + 60}{0.95} = 110\ m$$

With the 60 being the manual georectification error in meters and the denominator being 0.95 because there is 0.95 years between 1972-08-08 and 1973-07-22.

Changes in terminus area in km^2 were calculated by creating polygons of the digitized termini, and time series of terminus area change were plotted to show the evolution through a full surge cycle. Following methodology developed by Krumwiede et al. (2014), uncertainties for area changes (A_{er}) were assumed to be ± 1 pixel and calculated from:

$$A_{er} = (n \cdot m) \cdot 0.000001 \quad (2)$$

Where n is the number of pixels that make up the perimeter of the terminus area (approximated by dividing the outline length by the image pixel size), m is the spatial resolution of the sensor bands

expressed as the area of a pixel (e.g., 900 m² for a 30 × 30 m pixel), and it is multiplied by 0.000001 to convert it from m² to km².

2.3.2 *Ice surface velocities*

Surface velocities of Fisher Glacier were derived from two main sources: (i) annual and summer data from 1985 to 2018 were obtained from the NASA Inter-mission Time Series of Land Ice Velocity and Elevation (ITS_LIVE) website (<https://its-live.jpl.nasa.gov/>), while (ii) winter velocities for 2014-2017 were derived from RADARSAT-2 synthetic aperture radar (SAR) data, discussed further below.

The NASA ITS_LIVE dataset provides a globally comprehensive multi-sensor record of land ice velocity and elevation with high temporal resolution (Gardner et al., 2019). Annual velocity mosaics (1985-present), together with velocities derived from individual scene pairs, are produced by the auto-RIFT feature tracking processing chain (Gardner et al., 2018) using optical imagery from Landsat 4, 5, 7, and 8, as well as Sentinel 1 and 2. Files from the Alaska and Western North America region were downloaded and clipped to Fisher Glacier's region (RGI 6.0 polygon of Fisher Glacier) to obtain annual and shorter-term mapping of ice flow velocities. The annual velocity mosaics at 240 m resolution were used for most periods. These are created from a weighted average of all available image-pairs that have a center date that falls within that calendar year (Gardner et al., 2018). Since the annual mosaics are composed of many scene pairs, they have lower uncertainties than the individual scene pairs, but can miss details for velocity changes that occur over sub-annual periods. Therefore, velocity scene pairs at 120 m resolution (2013 and later) and 240 m resolution (before 2013) were used to derive short-term velocities for 2010-2017 during a period of rapidly changing flow speed (Table 3). Filters for scene pair acquisition were set to a time separation of 14-120 days, with >30 percent coverage of the area of interest. The velocity pair results were then manually sorted and selected for best coverage both spatially and temporally. This resulted in a mean image time separation of ~40 days for most image pairs.

Table 3. Individual NASA ITS LIVE image pairs used to cover the 2013-2016 period of enhanced flow speeds of Fisher Glacier, and the buildup phase from 2010-2012.

Date 1	Date 2	Image-pair ID	Platform
2010-02-18	2010-03-22	LE07_L1TP_061018_20100322_20160916_01_T1_X_LE07_L1 TP_061018_20100218_20160916_01_T1_G0240V01_P064	Landsat 7
2010-03-15	2010-04-23	LE07_L1TP_061018_20100423_20160915_01_T1_X_LE07_L1 TP_060018_20100315_20160916_01_T1_G0240V01_P062	Landsat 7
2010-07-28	2010-09-14	LE07_L1TP_061018_20100914_20160914_01_T1_X_LE07_L1 TP_061018_20100728_20160915_01_T1_G0240V01_P053	Landsat 7
2011-03-09	2011-04-19	LE07_L1TP_060018_20110419_20160913_01_T1_X_LE07_L1 TP_061018_20110309_20160913_01_T1_G0240V01_P073	Landsat 7
2011-05-29	2011-07-23	LT05_L1TP_061018_20110723_20160831_01_T1_X_LT05_L1 TP_060018_20110529_20160901_01_T1_G0240V01_P055	Landsat 5
2012-03-04	2012-04-05	LE07_L1TP_060018_20120405_20160912_01_T1_X_LE07_L1 TP_060018_20120304_20160912_01_T1_G0240V01_P060	Landsat 7
2013-07-28	2013-08-13	LC08_L1TP_061018_20130728_20200912_02_T1_X_LC08_L1 TP_061018_20130813_20200913_02_T1_G0120V02_P067(1)	Landsat 8
2013-08-13	2013-08-29	LC08_L1TP_061018_20130813_20200913_02_T1_X_LC08_L1 TP_061018_20130829_20200912_02_T1_G0120V02_P066(1)	Landsat 8
2013-10-16	2014-02-05	LC08_L1TP_061018_20131016_20200912_02_T1_X_LC08_L1 TP_061018_20140205_20200912_02_T1_G0120V02_P048	Landsat 8
2014-02-05	2014-03-25	LC08_L1TP_061018_20140205_20200912_02_T1_X_LC08_L1 TP_061018_20140325_20200911_02_T1_G0120V02_P073	Landsat 8
2014-11-13	2015-02-01	LC08_L1TP_060018_20141113_20200910_02_T1_X_LC08_L1 TP_060018_20150201_20200909_02_T1_G0120V02_P084	Landsat 8
2015-03-05	2015-04-22	LC08_L1TP_060018_20150305_20200909_02_T1_X_LC08_L1 TP_060018_20150422_20200909_02_T1_G0120V02_P079	Landsat 8
2015-08-23	2015-09-16	S1A_IW_SLC__1SSV_20150823T153143_20150823T153210_0 07392_00A2AE_4BE9_X_S1A_IW_SLC__1SSV_20150916T15 3144_20150916T153211_007742_00AC25_876B_G0120V02_P 082	Sentinel- 1A
2015-09-16	2015-10-10	S1A_IW_SLC__1SSV_20150916T153144_20150916T153211_0 07742_00AC25_876B_X_S1A_IW_SLC__1SSV_20151010T153 145_20151010T153212_008092_00B584_E55F_G0120V02_P08 2	Sentinel- 1A
2016-05-13	2016-06-06	S1A_IW_SLC__1SSV_20160513T153205_20160513T153236_0 11242_01101E_9FAF_X_S1A_IW_SLC__1SSV_20160606T153 206_20160606T153237_011592_011B69_EA6E_G0120V02_P0 71	Sentinel- 1A
2016-06-30	2016-07-24	S1A_IW_SLC__1SSV_20160630T153141_20160630T153209_0 11942_012676_BCCB_X_S1A_IW_SLC__1SSV_20160724T15 3143_20160724T153211_012292_0131E0_5AA4_G0120V02_P 081	Sentinel- 1A
2016-08-06	2016-09-23	S1A_IW_SLC__1SSV_20160806T024630_20160806T024657_0 12474_0137F7_4662_X_S1A_IW_SLC__1SSV_20160923T0246 32_20160923T024659_013174_014F17_8E88_G0120V02_P084	Sentinel- 1A
2016-09-10	2016-10-16	S1A_IW_SLC__1SSV_20160910T153145_20160910T153213_0 12992_014908_519F_X_S1A_IW_SLC__1SSV_20161016T1531 51_20161016T153219_013517_0159F7_D64C_G0120V02_P076	Sentinel- 1A

Since optical image matching used in ITS_LIVE requires surfaces with distinctive features to track, it typically works best in bare ice areas of the ablation zone of glaciers and is rarely used to derive winter velocities when the surface is typically covered in snow and daylight is limited. To derive winter velocities, we therefore used SAR-based speckle tracking of RADARSAT-2 ultrafine data from 2014-2017, subsampled to 50 m resolution (Table 4). SAR velocities derived from images collected in neighbouring paths/swaths within 7 days of the original acquisition were combined to provide better spatial coverage across the glacier (combined pairs are identified with a superscript number in Table 4). In all cases, velocities for overlapping areas were nearly identical, so the mean velocity was used. In such cases, the dates referred to in the text are the first date of the first velocity pair and the last date of the second velocity pair. Speckle tracking of SAR imagery was only used to derive winter velocities (January to April) to avoid incoherence between image pairs caused by melt in the summer months (Waechter et al., 2015).

Once all velocities had been gathered, they were extracted along the glacier centerline (Figure 2a) using the *Profile Tool's Terrain Profile* function in QGIS. This tool overlays a point or line feature (glacier centerline, in this case) over a raster and extracts the values underlying it. In this way, the centerline velocity values were extracted from the annual and individual image-pair rasters.

As it was not possible to calculate velocities from the historical air photos for the pre-satellite era, point velocities were determined from manual tracking of the displacement of distinctive looped moraines and moulins or potholes between 1954 and 1976. Using ArcGIS Pro, a point was placed at the tip of each looped moraine or in the center of each moulin/pothole, and the *Measure* tool was used to measure the displacement between the same feature on successive images. In total, 10 moulins or potholes (near the glacier centre) and 4 looped moraines (near the southern glacier margin) were identified and tracked over multiple images. Total displacement was measured from 1954-1972, and then from 1972-1976, and velocities were standardized to annual values. Uncertainties associated with image resolution were calculated in the same way as for the terminus change described above (1.8 m for 1954-1972 and 7.4 m for 1972-1976). While this method provides a general idea of ice velocities during this time, the temporal resolution is poor and it is not possible to distinguish between quiescent and active phase velocities if both occurred during a measurement period.

Table 4. Radarsat-2 ultrafine image pairs used to derive winter velocities.

Beam Path/Swath	Date 1	Date 2	Original resolution (m)	Mean velocity error (m yr ⁻¹)
U10W2	2014-03-13 ¹	2014-04-06	3	2.7
U17W2	2014-03-20 ¹	2014-04-13		
U10W2	2015-02-12 ²	2015-03-08	3	3.7
U17W2	2015-02-19 ²	2015-03-15		
U10W2	2016-02-07 ³	2016-03-02	3	2.5
	2016-03-02	2016-03-26		
U17W2	2016-02-14 ³	2016-03-09	3	2.8
U10W2	2017-01-08 ⁴	2017-02-01		
U17W2	2017-01-15 ⁴	2017-02-08		

^{1,2,3,4} Superscript numbers indicate which velocity pairs were combined in Figure 16.

2.3.3 Interpretation of surface feature change

During a surge, a glacier's surface often changes significantly due to a shift in supraglacial drainage system organisation, changing ice velocities, and downglacier transfer of mass. Surface features can therefore be used to identify surge-type glaciers and can contribute to understanding surge dynamics. Some of the most common surface features of surge-type glaciers are (Meier & Post, 1969; Lawson, 1997; Copland et al., 2011):

1. Distorted/looped surface moraines, which are created from rapid changes in ice motion over short distances, and can be located in the middle (i.e., medial moraine) or along the margins where tributary glaciers feed into the main trunk. When the main glacier's flow velocities are very slow during quiescence, faster flowing tributary glaciers can push into the main trunk, creating a moraine at the location where both ice masses meet. When the main trunk surges, it carries the moraine downvalley, folding it and creating distinctive looped features (Cuffey & Paterson, 2010);
2. Extensive crevassing. Crevasses are fracture features that result from changing stresses, or tension, in the ice and are often created by changing flow velocities (Hambrey & Alean, 2004; Cuffey & Paterson, 2010). Different types of stresses (e.g., compressive vs. extensive) lead to the formation of different types of crevasses defined by their

geometry, the main ones being transverse, longitudinal, splaying, and *en échelon* (Hambrey & Alean, 2004; Colgan et al., 2016).

3. Seracs are another type of feature created by tensional failure of the ice (Hambrey & Alean, 2004). They typically form in areas where several sets of crevasses intersect (Hambrey & Alean, 2004) or where the ice surface is convex and/or bends with the surrounding topography, which causes the ice to stretch and forms pinnacles (Allaby, 2020).
4. Sheared margins are another type of fracture feature that form along the margins of the glacier when the ice in the center flows faster than the ice along the sides, creating a shearing stress that fractures the ice. They can also be created by a lowering of the glacier surface in the reservoir zone, which exposes cliff faces that were previously ice-covered, and/or an ice speedup in the main trunk that cuts off a smaller tributary.
5. Formation or drainage of short-lived supraglacial lakes and streams as the drainage system at the glacier bed and surface changes. Formation of water-filled crevasses or lakes are indicative of high basal water pressures and/or blocked passageways to the glacier bed (e.g., moulins). Conversely, these features drain when basal water pressure drops (Murray et al., 2003). Such features can therefore be used to infer water pressure changes and drainage system reorganisation at the glacier bed.
6. Potholes, which are depressions or holes in the glacier surface, often filled with water, tend to form in clusters during the quiescent phase on surge-type glaciers (Sturm, 1987).

For this study, surface features on Fisher Glacier were identified visually in satellite imagery and air photographs, and traced out and tracked manually in ArcGIS Pro. A Gigapan (large panoramic) photo of the ablation zone taken from a bedrock knoll on the north side of the glacier (60°06'21.996" N, 138°05'54.96" W) by Dr. Luke Copland on 11 August 2015, and repeated on 25 July 2021, also offered an opportunity to analyze surface features from an oblique point of view that differs from what can be observed in nadir satellite imagery, and allowed for a detailed surface comparison to be made between 2015 and 2021. Additionally, numerous oblique photographs taken with handheld cameras from a helicopter (2015, 2021), and helio-courier aircraft (2021), offered excellent views of the various features at the terminus and in the ablation area.

The timing of crevasse appearance, their location, shape/type, and extent on the glacier surface, was used to provide insight into ice dynamic processes and a means to narrow down surge initiation and termination times and locations. The drainage of supraglacial lakes is also connected to crevasse formation and changes in basal water pressures (Murray et al., 2003), and was assessed visually to help reconstruct surges and infer possible surge mechanisms. As potholes are associated with the quiescent phase of the surge cycle (Sturm, 1987), their evolution on the glacier surface through time was visually assessed and characterized using available imagery. This visual analysis of changes in crevasses, supraglacial drainage features, and potholes is of particular importance in identifying older surges which occurred during times when glacier velocities and surface elevation changes were difficult, or impossible, to determine.

2.3.4 Surface elevation changes

Surface elevation and ice volume changes provide important information about the state and evolution of surge-type glaciers because mass tends to accumulate in the reservoir zone during the quiescent phase (i.e., long term increase in surface elevation), and subsequently move downglacier during the active phase (Meier & Post, 1969; Kamb et al., 1985; Dowdeswell et al., 1991; Kochtitzky et al., 2019). In this study DEMs were both downloaded from existing sources, and newly derived from historical air photos and satellite images, as detailed further below.

2.3.4.1 Historical DEMs from air photos

The historical air photographs acquired from the National Air Photo Library and the Yukon Energy, Mines and Resources Library were used to create high-resolution orthoimages and DEMs of Fisher Glacier using the Structure from Motion (SfM) technique, which allows for the generation of DEMs from 2D imagery. SfM has been successfully applied to many types of images (e.g., nadir, oblique) in glaciology, including those obtained from fixed wing aircraft and helicopter (Thomson & Copland, 2016; Medrzycka et al., 2023), drone/UAV (Shugar et al., 2017), or even handheld at ground level (Vargo et al., 2017; Lewińska et al., 2021).

In this study Agisoft Metashape Professional software was used to process the images for SfM analysis. Prior to importing the photos, they were visually assessed to remove any that were overexposed. This was common in photos taken over snow-covered areas at higher elevations. The photos were then imported into Metashape and masks manually added to remove any non-desired

objects such as borders and markings added to the images. Once imported, Metashape's automated photo quality estimation feature assessed the quality of each image by calculating the sharpness level of the most focused part of each one and assigned them a value from 0-1.0. Photos with a value below 0.5 were discarded. Fiducial points were then manually applied to the marks on the images to tell the software which way the photos should be oriented when looking for tie points (TPs). Next, GCPs were manually added. Each image was assigned a minimum of 5 GCPs when possible, although a select few images in the snow-covered areas at higher elevations had less than this due to a lack of distinctive features in the terrain. Additionally, all GCPs overlapped on at least two images. GCPs were chosen by comparing the air photos to a mosaic of cloud-free 3-m resolution PlanetScope images from 4 July and 12 August 2022, and identifying distinctive features on stable ground. For each GCP, its coordinates (derived from the PlanetScope images) and elevation in m a.s.l. (derived from the ArcticDEM) were provided to Metashape.

With the calibrated fiducial marks and GCPs, the camera alignment was optimized. Next, the photo alignment (sparse point cloud) took place, which typically requires multiple iterations where the GCPs and TPs are added or removed to minimize the reprojection error (i.e., a measure of location accuracy of pixels) (Thomson & Copland, 2016; Medrzycka et al., 2023). At this step, outlier points were omitted from the sparse point cloud. A dense point cloud was then created using the depth map from the previous step, with quality set at "Ultra high" and filtering on "mild" so that smaller features were not filtered out as outliers. The resulting point cloud was again manually filtered to remove any remaining outlier points. The point cloud was then used to reconstruct a 3D model which was exported as a DEM. The final step was to build the texture, after which the final product was exported as an orthophoto.

The DEMs and orthophotos were exported from Metashape with a WGS 1984 coordinate system and subsequently projected to WGS 1984 UTM Zone 7N in ArcGIS Pro. The processing statistics for the final products are provided in Table 5. In the end, three historical DEMs were created from photo rolls A14459 (1954), A23001 (1972), and A24520 (1976).

Table 5. Metashape SfM processing statistics for historical DEMs.

	A14459	A23001	A24520
Survey date	1954-08-30	1972-08-08	1976-07-25
Total images used	16	17	16
Ground resolution (m)	2.62	3.06	3.09
Tie points	10,068	22,671	24,134
GCPs	19	36	31
Dense point cloud points	120,776,217	98,817,846	138,505,331
Reprojection error (pix)	1.85	9.22	2.54
Control point RMSE (m)	12.2	74.7	12.6
Roll image numbers	63-65, 104-107, 111-114, 151-155	84-85, 90-92, 124-127, 131-134, 165-167, 212	176-181, 204-208, 221-225

2.3.4.2 Satellite-derived DEMs

Existing DEMs were acquired from two main sources:

1. ArcticDEM strips were downloaded from the University of Minnesota (<https://www.pgc.umn.edu/data/arcticdem/>). The ArcticDEM is based on stereo WorldView-1, 2, and 3 imagery and, with up to 3 m resolution, it is the highest resolution DEM of the study area that is publicly available. However, it contains many holes in its coverage which limits its use in certain areas.
2. ASTER GDEM v3 (<https://asterweb.jpl.nasa.gov/gdem.asp>) was used for co-registering the DEMs that were created specifically for this project. The ASTER GDEM has 30 m spatial resolution and is not affected by jitter artifacts that typically plague ASTER imagery as the GDEM v3 benefits from a refined production algorithm which improves spatial resolution and horizontal and vertical accuracy (Shean et al., 2016; Girod et al., 2017).

The pre-existing DEMs described above are often generated from many satellite images collected over a multi-year period, and in some cases the exact dates are not provided with the DEM product. Custom DEMs must therefore be generated to obtain surface elevation data for known, specific periods of time. For this project, satellite based DEMs were created from Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) stereo satellite imagery. The ASTER sensor is housed onboard the Terra (EOS AM-1) satellite and offers the largest consistent multi-temporal dataset of stereo images, with availability since 1999. The data are affected by a high frequency jitter that causes an attitude perturbation on the satellite, which can be minimized by using MicMac ASTER (MMASTER) software developed by Girod et al.

(2017) to produce DEMs. With an accuracy better than ± 10 m, MMASTER DEMs provide more accurate glacier elevation data than other software options (Girod et al., 2017).

ASTER images covering Fisher Glacier and its surroundings were downloaded from the NASA Earthdata Search portal (<https://search.earthdata.nasa.gov/search>) as Level 1A reconstructed unprocessed instrument data products. Filters were set to only select images containing 80% cloud cover or less and taken during daylight hours. This yielded 99 results, which corresponded to 70 images as some of the strips were stitched together as a single scene. The resulting 70 ASTER images were converted to DEMs using the MMASTER processing chain (Girod et al., 2017). Batch processing of the DEMs was run on the Compute Canada Cedar supercomputer. The DEMs were then manually inspected one at a time to remove those containing visible noise due to cloud cover or bad lighting, as well as those that did not cover the region of interest. To reduce uncertainties caused by differences between DEMs, only 1 DEM was selected per year. An exception to this was for 2016 where two DEMs were retained; the first covering the top half of the glacier and the second covering the bottom half. This process yielded a total of 9 DEMs that were sufficiently noise-free, from 2003, 2004, 2006, 2014, 2016, 2018, 2019 and 2020.

2.3.4.3 DEM co-registration and trend analysis

Once the historical and MMASTER DEMs had been created the next step is co-registration, which ensures that the pixels in each DEM represent the same location on the Earth's surface (Nuth & Kääb, 2011). Three potential biases to account for in this step are in the geolocation of the data (x, y, and z matrices), any elevation-dependent bias, and bias related to acquisition geometry of the data (Nuth & Kääb, 2011). For example, uneven distribution of GCPs in the x, y, z plane can lead to an elevation dependent bias due to poorly resolved stereo orientation which causes distortion of the z-scale (Nuth & Kääb, 2011). Thus, it is only once stable terrain has been matched between DEMs (no horizontal offsets between them) that any elevation differences on the ice surface will be accurately observable when differencing them. A DEM co-registration Python package (<https://github.com/iamdonovan/dem-coregistration>) created by Dr. Robert McNabb was used to iteratively co-register the air photo DEMs created in Metashape and the ASTER DEMs created with MMASTER. This Python package is based on the co-registration algorithm developed by Nuth & Kääb (2011).

Prior to co-registration, the DEMs were resampled to 30 m to match the ASTER GDEM v3. Some of the DEMs required additional filtering to remove smaller patches of noise. This was accomplished through the ArcGIS Pro Python console by differencing the DEMs with the ASTER GDEM v3 and then filtering for values over a certain threshold. For the MMASTER DEMs, the threshold was set at 100-125 m (depending on the DEM). For the air photo DEMs, the threshold value was set at 200-250 m to account for greater errors in this data source and a longer time separation between the differenced DEMs, which could lead to greater changes over the glacier surface. Once all DEMs were filtered to remove small clouds or other noise, they were co-registered to the ASTER GDEM v3 using the co-registration package described above.

Once co-registration was completed, trends in surface elevation change were computed using the `make_stack.py` script from the `pygeotools` Python package by Dr. David Shean. `Make_stack.py` outputs a raster showing the change in elevation standardized to values of m yr^{-1} . It requires at least 3 DEMs, which allows it to reduce uncertainties by calculating the observed changes over stable ground and adjust the corresponding change over the glacier surface. The tool uses a “stack” of rasters to create a raster time series object from which many statistics are calculated for each cell, such as mean, minimum, maximum, and standard deviation. Trends in surface elevation change were computed for 3 periods: 2003-2014, 2016-2020, and 2003-2020, to enable quantification of elevation differences around a period of rapid glacier change in 2014-2016.

Since `make_stack.py` requires multiple DEMs from different years, it was not used for the air photo DEMs, nor for the MMASTER DEMs covering the period of rapid glacier change in 2014-2016 (since there were only two DEMs: one for 2014 and another for 2016). In these two cases, the DEMs were differenced using the *Minus* tool from the *Spatial Analyst* toolbox in ArcGIS Pro, and the resulting dDEM were standardized to annual values.

Once the trend DEMs had been filtered and standardized to annual values, the mean elevation change per 10 m and 50 m surface elevation band (hypsothetic band) was computed using the *Contour* tool and *Zonal Statistics* in ArcGIS Pro. Uncertainties were measured by calculating the mean value over stable ground.

2.3.4.4 Glacier trimlines as a means of deriving elevation change

As a means to validate the results shown by the historical and 2014-2016 dDEMs, a secondary method for measuring ice surface elevation change was employed by measuring the elevation of trimlines (ice margins) along the valley walls in historical imagery. In the context of this study, trimlines are the location where the glacier ice intersects the lateral moraines or valley walls along the glacier margin. This method assumes that changes along the glacier margin are representative of changes across the glacier's width and are considered as a minimum estimate of ice thickness loss (Curley et al., 2021). Although this method does not offer high spatial resolution across the glacier surface, it provides a broad estimate of surface elevation change when other data sources are limited or non-existent.

The ice extent along two sections on the northern margin of Fisher Glacier was manually traced in ArcGIS Pro (Figure 4). We used the same imagery that was used for the terminus analysis (i.e., RCAF air photos from 1954, 1972, and 1976, KH-9 image from 1977, RapidEye imagery from 2010 and 2016, and PlanetScope image from 2022). Changes were quantified along two sections which had clear boundaries where the ice met the northern valley wall: the first section was 1.6 km-long, near the terminus (green dots in Figure 4), and the second section was 2.4 km-long, in the lower ablation area (blue dots in Figure 4). Ice margin elevations were manually traced and transects were generated every 100 m, perpendicular to the centerline (dashed lines in Figure 4). Ice margin elevations at specific points were then extracted at each intersection between the 100 m transects and the ice margin outlines (larger points in Figure 4). The average elevation for each year was calculated to give an approximate elevation change. The 2014 MMASTER DEM (i.e., the lowest and smallest glacier extent on record) was used as the background DEM from which the ice margin elevations were extracted based on the location of the glacier margin in historical and recent imagery.

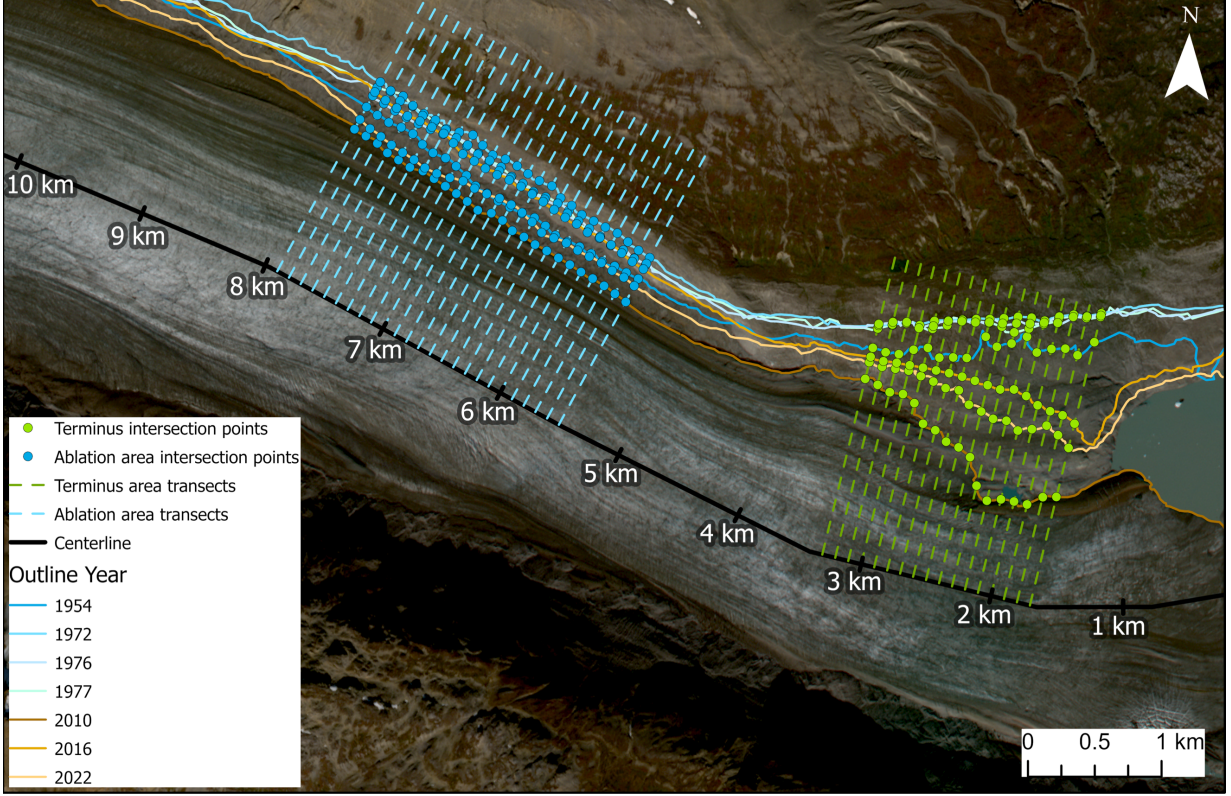


Figure 4. Illustration of the outlines used to derive ice elevations along the northern margin of Fisher Glacier during the historical (1954-1977) and recent (2010-2022) periods. The 100 m transects are represented by dashed lines and the intersection points at which elevations were extracted are displayed in the same colour. The centerline with 1 km markers from the 2015 terminus is also provided for reference. Base image: RapidEye-4, 2010-09-15.

Uncertainties of the ice margin elevations were determined by shifting the trimlines up- and down-valley (in this case, by x distance North, and x distance South) by a distance of two image pixels and extracting the elevations at these new locations (following the methodology of Curley et al. (2021)). The difference in elevation was then used as the uncertainty. We used the same outline uncertainty as was used for the terminus outlines (i.e., 2 image pixels = 5.24 m up- and down-valley displacement for 1958, 6.12 m for 1972, 7.00 m for 1976, 15.00 m for 1977, 10.00 m for 2010 and 2016, and 6.00 m for 2022). The average difference in mean elevation for all years combined was 3.1 m (0.62 m yr^{-1}) for the lower ablation area section and 2.2 m (0.34 m yr^{-1}) for the terminus section.

2.4 Results

2.4.1 Terminus changes from 1948-2022

2.4.1.1 Terminus area changes

Historical air photographs and intelligence satellite imagery from 1948 to 1963 show that the terminus area of Fisher Glacier decreased steadily over this time (Figure 5a), with a total reduction of $3.03 \pm 1.85 \text{ km}^2$ (from $13.08 \pm 0.02 \text{ km}^2$ in 1948 to $10.05 \pm 1.83 \text{ km}^2$ in 1963). Between 1963 and 1972 the area increased by $5.81 \pm 1.89 \text{ km}^2$, reaching $15.86 \pm 0.06 \text{ km}^2$ in 1972, the largest over the study period. Over the following ~42 years the terminus underwent consistent retreat, until it reached its minimum extent of $2.41 \pm 0.04 \text{ km}^2$ in 2014, corresponding to an annual average decrease of $-0.32 \pm 0.0024 \text{ km}^2 \text{ yr}^{-1}$. The rate of retreat over this period was fairly constant (Figure 5a).

The period 2014 to 2016 was marked by a rapid increase in terminus area (Figure 5b). Fisher Glacier's trend of decreasing terminus area persisted through the first half of 2014, until it reached its minimum on 31 July 2014. However, by 2 September 2014 it had increased by $0.13 \pm 0.08 \text{ km}^2$ in only 34 days, at a rate of $1.39 \pm 0.86 \text{ km}^2 \text{ yr}^{-1}$. The rate of advance then slowed through the fall and winter of 2014-15, increasing by $0.13 \pm 0.17 \text{ km}^2$ by 21 May 2015 (mean rate of $0.19 \pm 0.24 \text{ km}^2 \text{ yr}^{-1}$). After May 2015 the rate of change began climbing steadily through the summer and fall of 2015, reaching a peak of $5.86 \pm 0.24 \text{ km}^2 \text{ yr}^{-1}$ between 4 October 2015 and 10 May 2016 (i.e., during the winter). The rate of advance then suddenly slowed in summer 2016, reaching a rate of $0.41 \pm 2.61 \text{ km}^2 \text{ yr}^{-1}$ between 18 July and 21 August 2016, after which the terminus became stagnant, or showed very slight retreat, until 6 August 2017. The period of rapid change lasted for a total of ~2.1 years (31 July 2014 to 21 August 2016), during which the terminus increased in area by a total of $5.41 \pm 0.23 \text{ km}^2$.

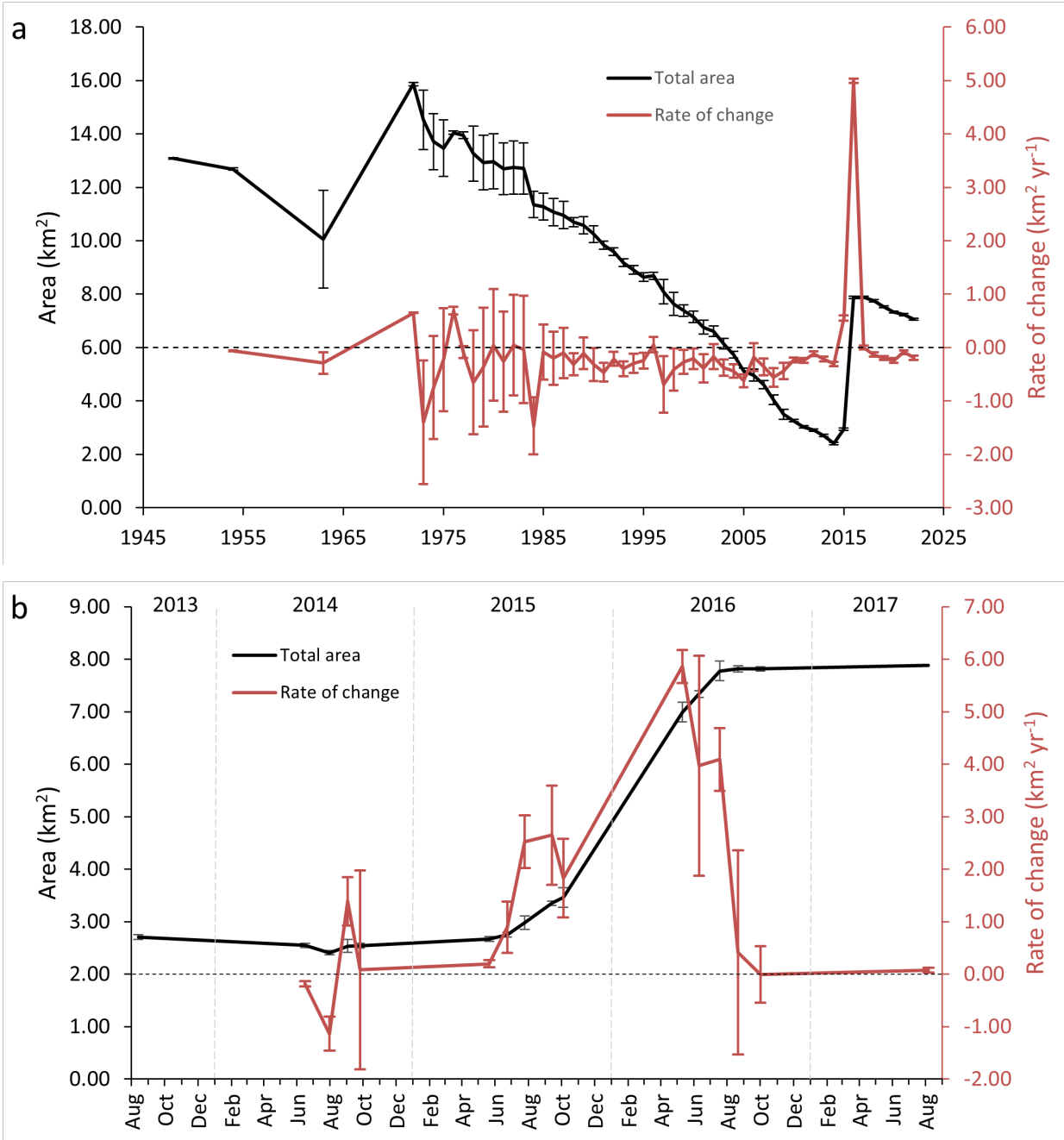


Figure 5. (a) Time series of total terminus area (black line) and annual rate of change (red line) for Fisher Glacier from 1948-2022 (b) Time series of total terminus area (black line) and annual rate of change (red line) during the period of rapid terminus advance from 13 August 2013 to 6 August 2017.

By 2017, terminus area changes returned to being negative, with this trend being maintained to the end of the study period in summer 2022. The average rate of terminus change

was $-0.14 \pm 0.01 \text{ km}^2 \text{ yr}^{-1}$ between 2017 and 2022, with a total terminus area decrease of $-0.82 \pm 0.08 \text{ km}^2$ over these five years (Figure 5a).

2.4.1.2 Terminus length changes

The satellite imagery and historical air photographs show that Fisher Glacier's terminus underwent three periods of sustained retreat, which were separated by two short periods of rapid advance over the study period (Figures 6a and 7). The first period of retreat lasted from 1948 to 1963, with a terminus-wide mean retreat rate of $-7 \pm 5 \text{ m yr}^{-1}$ from 1948-1954, increasing to $-37 \pm 20 \text{ m yr}^{-1}$ from 1954-1963. The glacier retreated by an average of 375 m across its width, although there were very large spatial variations as both water-terminating Northern and Southern sections of the terminus retreated by $>1 \text{ km}$, while the Central land-terminating section only retreated by an average of $-77 \pm 14 \text{ m}$ (Table 6).

From 1963 to 1972, Fisher Glacier underwent a significant advance, averaging $+63 \pm 25 \text{ m yr}^{-1}$ across its terminus (Figure 6b). However, it should be emphasized that there is a nearly 9-year gap in image availability during this advance and based on Fisher Glacier's recent behavior, and that of other glaciers in the vicinity, it is likely that the advance lasted for a shorter period than this. This means that peak advance rates were probably much higher than those reported here based on a 9-year average.

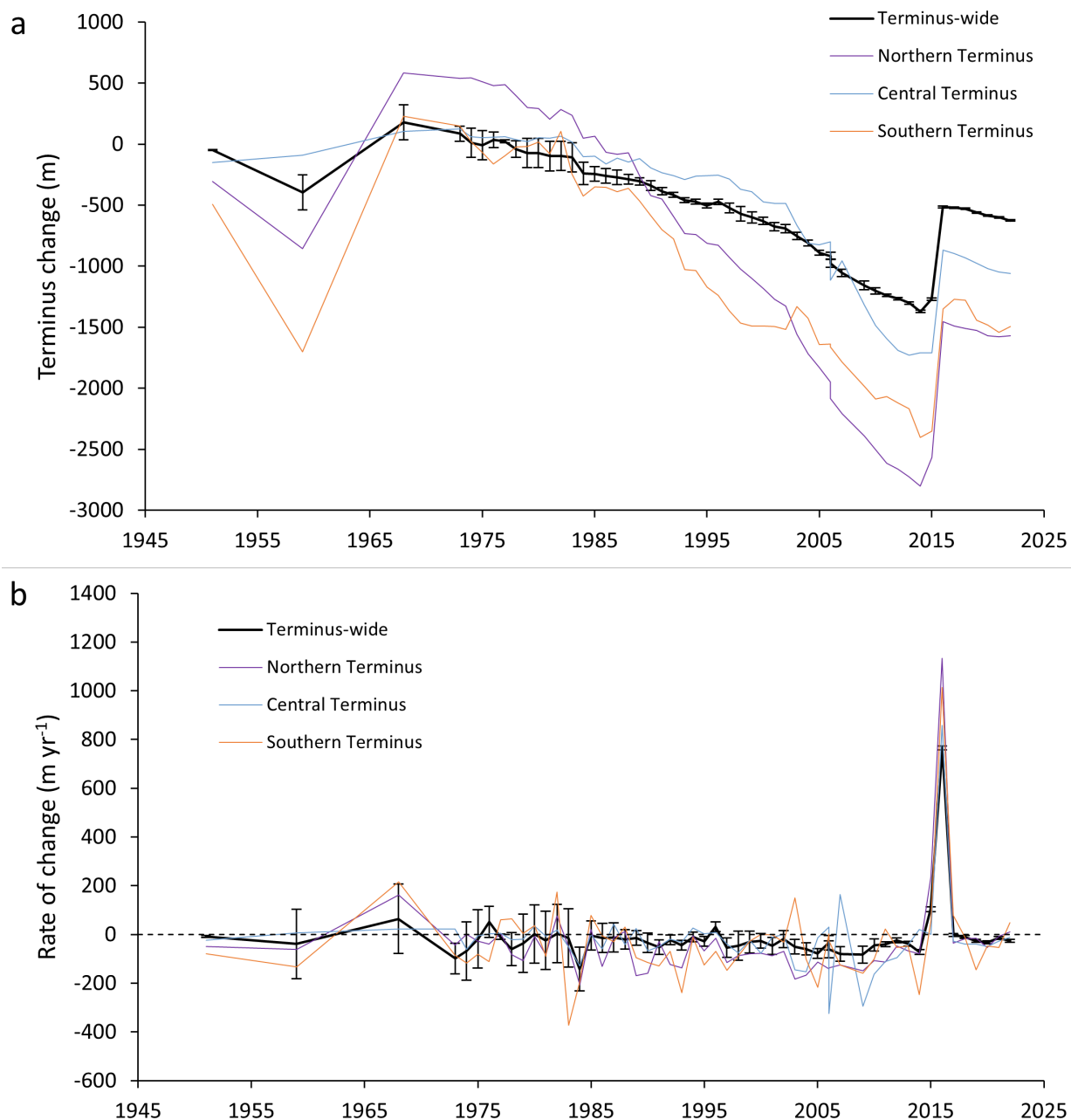


Figure 6. Time series of: (a) change in terminus length of Fisher Glacier in relation to its 1948 extent, and (b) annualized rate of change, from 1948 to 2022. Error bars are only included on the terminus-wide line because they are the same for all lines.

After 1972 the terminus underwent sustained retreat for the next ~42 years, attaining a terminus-wide average of -2.05 km ($-49 \pm 1 \text{ m yr}^{-1}$) (Figure 6b; Figure 7c). Much like the retreat from 1948-1963, this did not occur evenly across the terminus, with the Northern Terminus (NT) section retreating the most (3567 m ; -85 m yr^{-1}) and the Central Terminus (CT) section retreating

the least (1930 m; -46 m yr^{-1}), while the Southern Terminus (ST) section was in between the two others (2564 m; -61 m yr^{-1}) (Table 6). The surrounding topography of each section is quite different, with the CT being land-terminating and, at its maximum extent, the glacier snout pushing up against a $\sim 300 \text{ m}$ high hill. In contrast, both the NT and ST are water-terminating, although the NT terminates in deeper water in a permanent lake, whereas the ST terminates in seemingly shallower water (Figure 3b).

Table 6. Mean total terminus change of Fisher Glacier in distance (m) and area (km^2) per successive period of advance/retreat from 1948 to 2022.

	Terminus- wide	NT	CT	ST	Uncertainty (m)	Area (km^2)	Area Uncertainty ($\text{km}^2 \text{ yr}^{-1}$)
1948-1963	-375	-1184	-77	-1644	142	-3.0	1.85
1963-1972	+571	+1435	+185	+1913	143	+5.8	1.89
1972-2014	-2058	-3567	-1930	-2564	8	-12.1	0.03
2014-2017	+868	+1635	+836	+1144	8	+5.4	0.80
2016-2022	-100	-163	-201	-145	6	-0.8	0.01

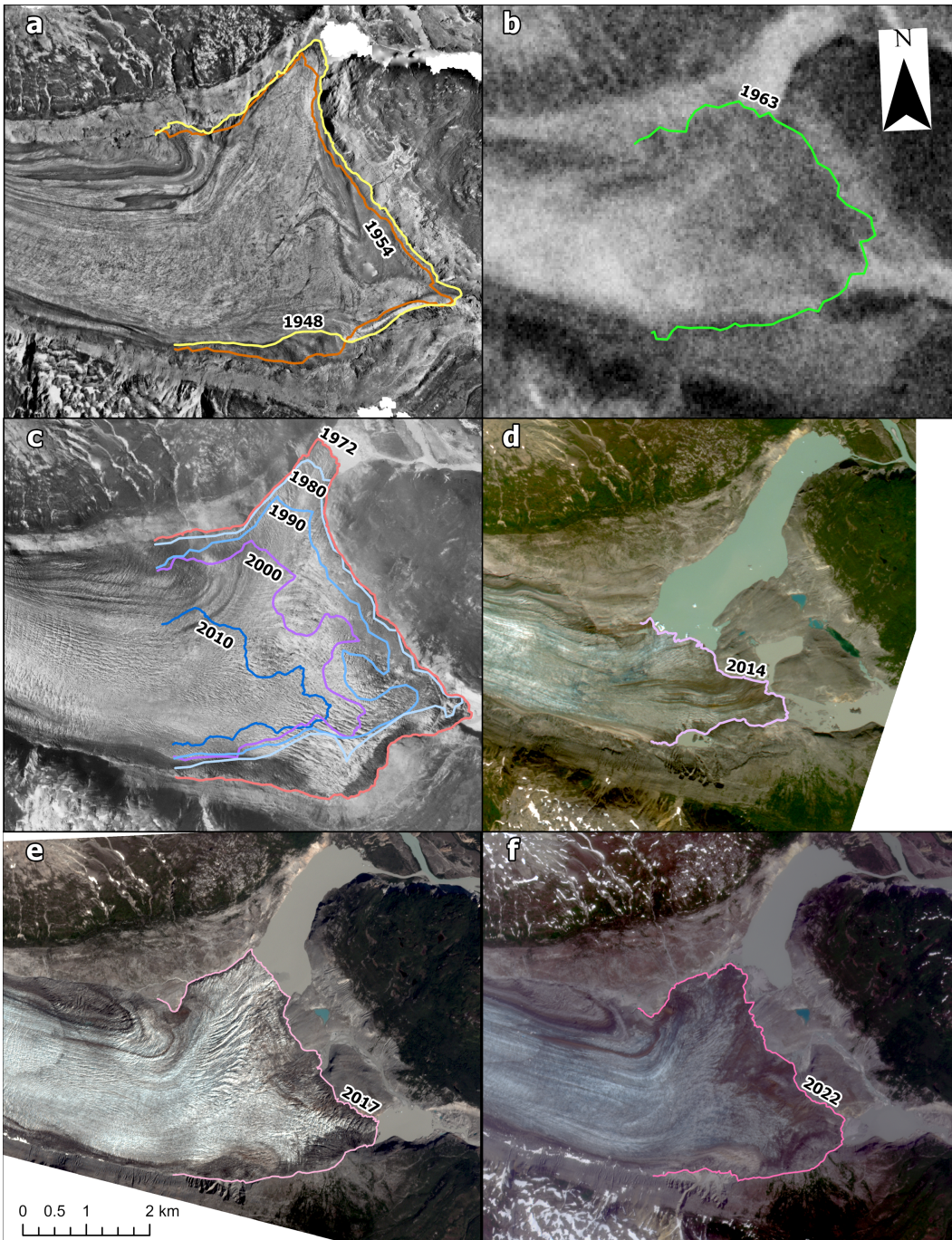


Figure 7. Terminus extents during each of Fisher Glacier's periods of advance and retreat between 1948 and 2022: (a) Period of retreat from 1948-1954. Base image: 1948-06-12 air photo. (b) KH-5 image from 1963-08-29, after which the terminus began advancing. (c) Air photo from 1972-08-08 at the glacier's maximum extent, with decadal terminus outlines showing retreat patterns up to 2010-09-15. (d) RapidEye-3 image from Fisher Glacier's minimum extent on 2014-07-31, before the recent advance. (e) 2017-08-06 PlanetScope image, just after the most recent period of advance ended. (f) 2022-07-04 PlanetScope image showing the current (summer 2022) terminus extent.

As of summer 2014, Fisher Glacier's terminus began a new period of advance (Figure 8a, Figure 9a, b). Between 31 July and 2 September 2014, the terminus underwent a rapid but small advance of an average of 18 ± 78 m, equivalent to a rate of $+204 \pm 78$ m yr⁻¹, although some areas such as the NT were still retreating. Over fall and winter 2014/15, terminus change rates generally stayed above zero, but remained low until May 2015 (Figure 8b). The first significant terminus advance then occurred between 21 May and 23 June 2015, although it was mainly constrained to the NT (Figure 8a). The NT advanced by 28 ± 10 m in 33 days (a rate of 635 m yr⁻¹). The rate of advance of the NT kept climbing throughout summer 2015, with the other subzones also having a small peak in July 2015 (Figure 8).

The winter of 2015/16 is when the terminus underwent its greatest advance (Figure 8; Figure 9b, c). Between 4 October 2015 and 10 May 2016, it advanced by an average of 526 ± 20 m, with local amounts of 592 ± 20 m (ST), 657 ± 20 m (CT), and 1236 ± 20 m (NT). The terminus-wide annual rate of advance was at 877 ± 24 m yr⁻¹, with the NT reaching up to 2059 ± 24 m yr⁻¹. The rate of advance remained high throughout early summer 2016, until sometime between 18 July and 21 August 2016 when the rate of advance rapidly dropped. The terminus was virtually stagnant by October 2016.

After the period of rapid terminus advance from 2014-2016, Fisher Glacier's terminus stabilized and began retreating very slowly over the following 5 years (Figure 8b). Since the terminus stabilized in late summer 2016, it has retreated at an average rate of only 17 m yr⁻¹ (Figure 6) and, as of the most recent measurement on 30 July 2022, remains at nearly the same extent as it was at the end of the last advance (compare Figure 7e and f).

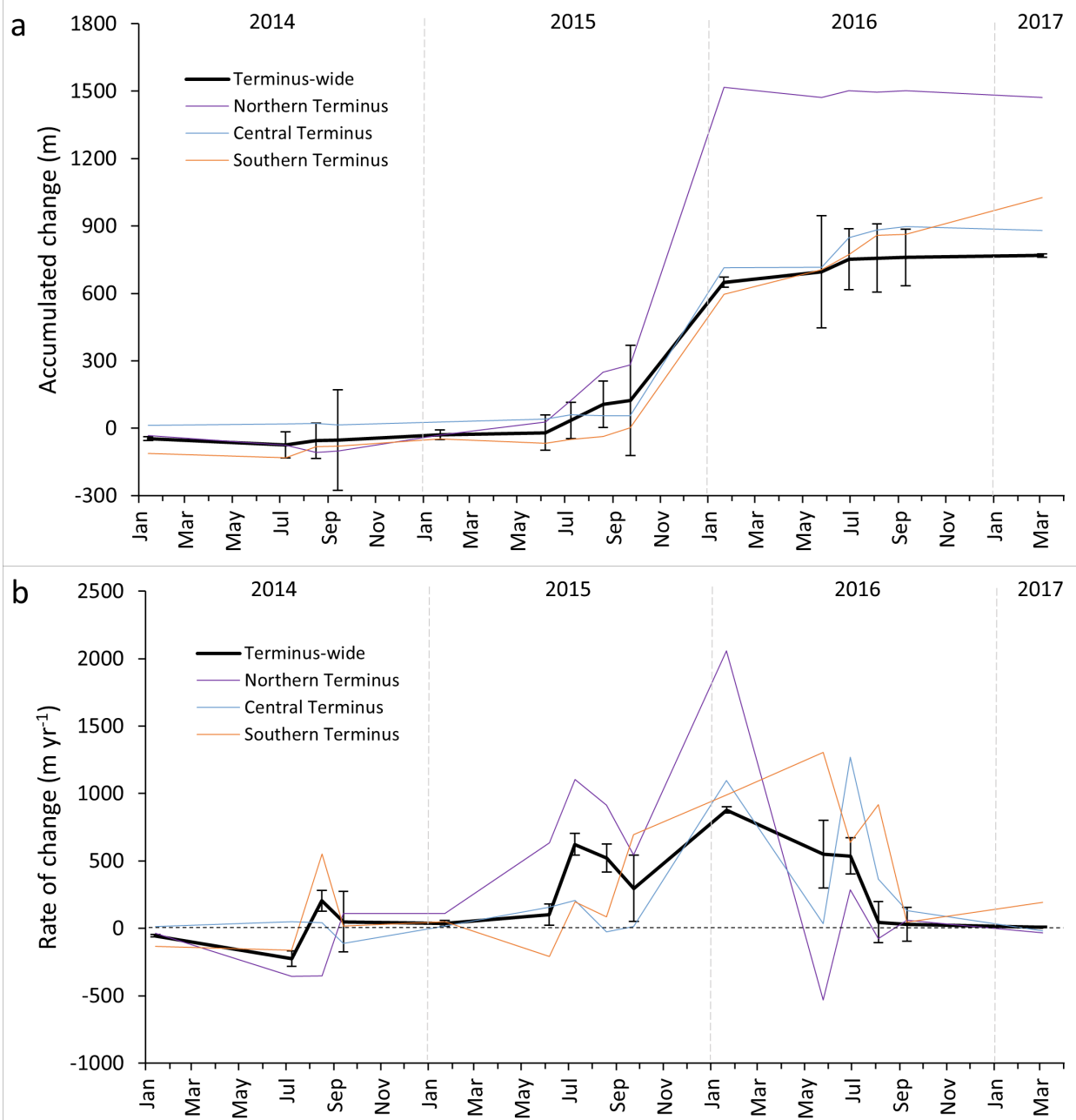


Figure 8. Time series of (a) change in terminus length of Fisher Glacier, and (b) annualized rate of change, during the period of rapid terminus advance from 13 August 2013 to 6 August 2017. Error bars are only included on the terminus-wide line because they are the same for all lines.

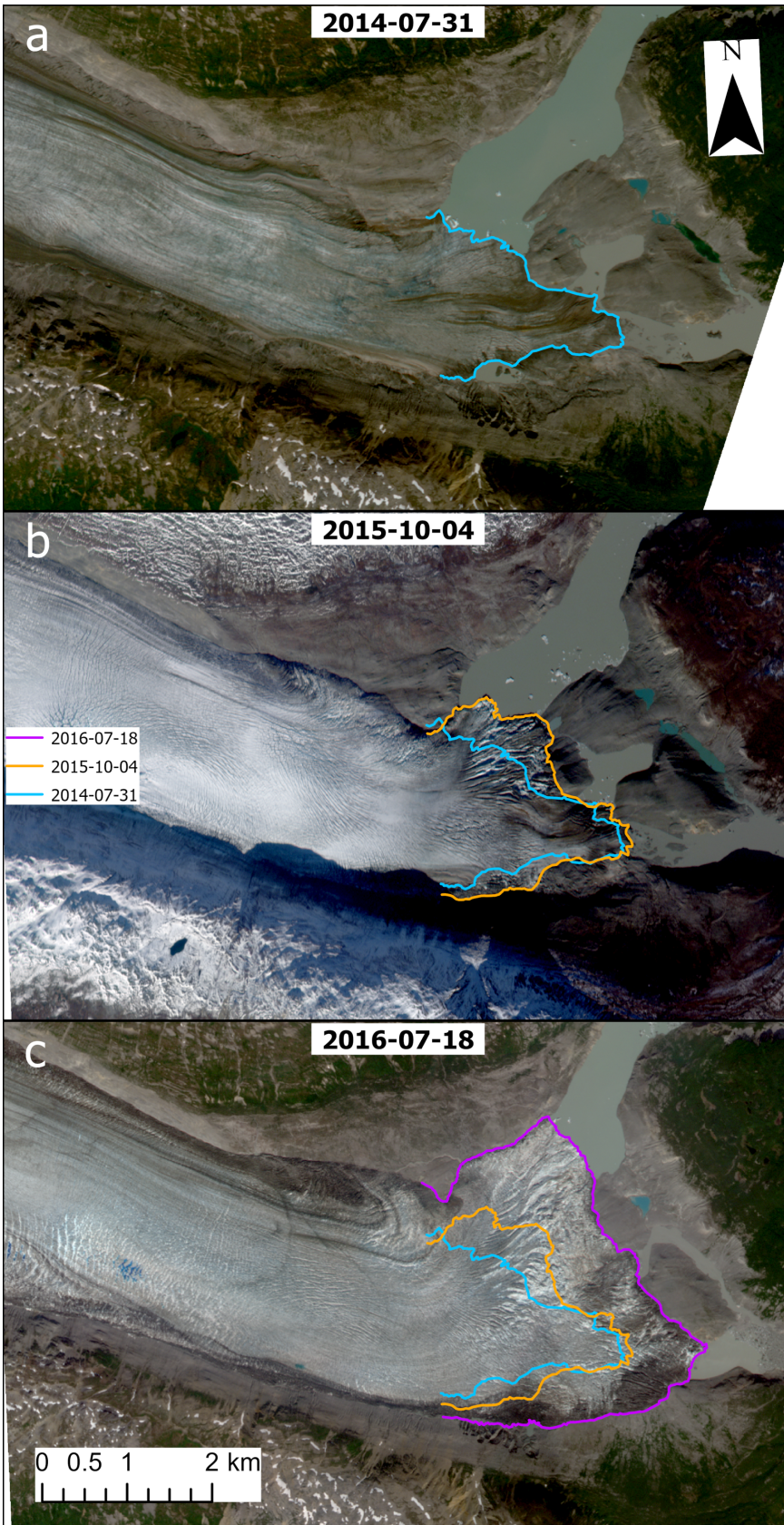


Figure 9. Terminus extents of Fisher Glacier at three key periods during 2014-2016: (a) RapidEye-3 image from 2014-07-31 when Fisher Glacier was at its minimum extent; (b) RapidEye-1 image from 2015-10-04 and corresponding terminus outline (orange), shortly after the first significant terminus advance; (c) RapidEye-3 image from 2016-07-18 when the terminus was near its maximum extent and had stopped advancing, with the outlines from (a) and (b) for comparison.

2.4.2 Velocity changes from 1954 to 2022

Manual tracking of surface features from historical air photos from 1954, 1972, and 1976 provide point surface velocities in the mid-region (ablation area) of the glacier, ~10-20 km from the terminus (Table 7, Figure 10, Figure 11a). Potholes and moulins located along the glacier centerline progressed downglacier by an average of $233 \pm 1.8 \text{ m yr}^{-1}$ from 1954 to 1972 (Table 7, Figure 10). The surface velocity then decreased to an average of $22 \pm 7.4 \text{ m yr}^{-1}$ from 1972 to 1976 (Table 7). Due to the lack of detail concerning how velocities varied during 1954-1972, it is possible that short-term velocities were much higher during part of this period if a surge occurred. The likelihood of this is assessed in the Discussion.

Table 7. Displacement of potholes/moulins and looped moraines on Fisher Glacier over two historical periods. See Figure 10 to see pothole/moulin numbering and Figure 11a for looped moraine numbering. Potholes 7-10 were only tracked from 1972-1976, whereas looped moraines were only tracked from 1954-1972.

Pothole/moulin	1954-1972		1972-1976	
	Total displacement (m)	Point velocity (m yr^{-1})	Total displacement (m)	Point velocity (m yr^{-1})
1	4178	233	105	26
2	4165	232	102	26
3	4134	230	66	17
4	4135	230	91	23
5	4251	237	70	18
6	4263	237	86	22
7	N/A	N/A	75	19
8	N/A	N/A	86	22
9	N/A	N/A	75	19
10	N/A	N/A	109	27
Looped moraine				
1	2789	155	N/A	N/A
2	2382	132	N/A	N/A
3	2441	136	N/A	N/A
4*	2665	148	N/A	N/A

*New looped moraine that appeared between the two dates. Distance was measured from center of Tributary 3 (see Figure 2a for tributary locations).

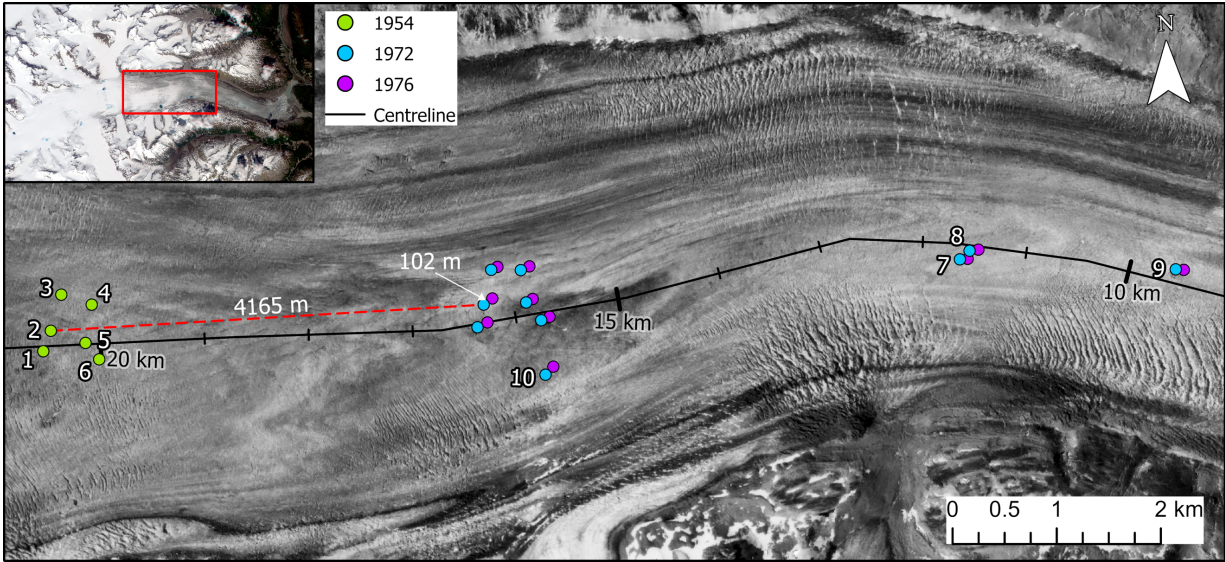


Figure 10. Moulin and potholes tracked to derive point velocities from 1954-1976 on Fisher Glacier. Tracked features are numbered 1-10 and symbol colour represents the year for a given feature (see legend). Features 7-10 were only tracked from 1972-1976. Red dashed line shows an example of how feature #2 was tracked. The black line is the glacier centerline, with distances upglacier from the terminus. The displacement between features is presented in Table 7. Base image: RCAF air photo, 1972-08-08. Inset image: RapidEye-1, 2016-06-22.

Displacement of looped moraines located on the southern margin of the glacier from 1954-1972 show slower velocities than what were measured from potholes along the centerline (Table 7, Figure 11a). The margins of the glacier flowed at $143 \pm 1.8 \text{ m yr}^{-1}$ from 1954-1972, approximately 90 m yr^{-1} slower than the centerline. All of the looped moraines moved at approximately the same speed, ranging from $133\text{-}154 \pm 1.8 \text{ m yr}^{-1}$ (Table 7). The three looped moraines that existed in 1954 moved downglacier by an average of $2054 \pm 5.6 \text{ m}$ (Table 7, Figure 11a). In 1972, a new looped moraine (looped moraine 4; Figure 11a) had been produced and had moved downglacier by $2665 \pm 5.7 \text{ m}$. The distance travelled downglacier for the new looped moraine was measured from its tip to the center of Tributary 3, from which it originated. The movement of the looped moraines could not be tracked in the 1976 image because the southern margin of the glacier is overexposed and lacks sufficient detail to confidently map the features.

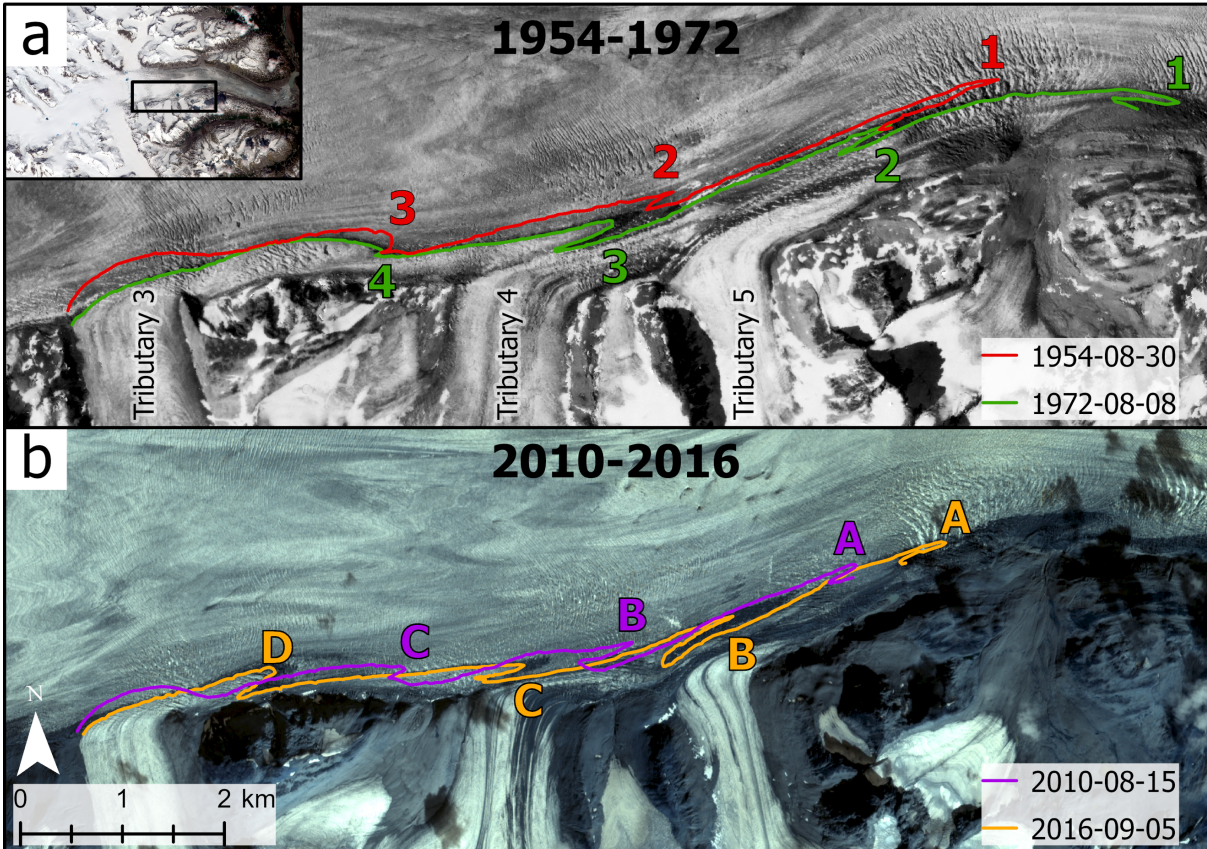


Figure 11. Looped moraine displacement on Fisher Glacier during the two main periods of significant change: (a) 1954-1972 and (b) 2010-2016. Although most of the looped moraines in (a) likely correspond to those in (b), we identify them separately in each part to avoid incorrect identification of similar-looking features. In both cases, a newly-formed looped moraine can be seen in the later date. Base images are RCAF air photo A23001 (1972-08-08) and a PlanetScope image (2016-09-05) for (a) and (b), respectively.

NASA ITS_LIVE average annual velocity mosaics show that surface velocities gradually increased, though remained low, from 1985-2010 (Figure 12). Over this time, the mean annual centerline velocity (measured over a distance of 0-45 km along the centerline from the terminus, including 500 m either side) increased by a few meters per year (Figure 13). Figure 12 illustrates the gradual up- and down-glacier spread during the multidecadal slow velocity increase. The average annual centerline velocities were $<50 \text{ m yr}^{-1}$ from 1985 to 1999. From 2000 to 2006 velocities were $<60 \text{ m yr}^{-1}$, and they remained below 90 m yr^{-1} until 2011. Between 2011 and 2012, average velocities increased from $83 \pm 10 \text{ m yr}^{-1}$ to $106 \pm 10 \text{ m yr}^{-1}$ (Figure 13). This single-year increase by 22 m yr^{-1} contrasted with the trend observed throughout the previous decades, which saw an average annual increase of only $\sim 2 \text{ m yr}^{-1}$ (Figure 13). Figure 14 shows that the velocity

speedup in 2011 occurred in the central ~20 km of the glacier. Over the next year (from 2012 to 2013), the average velocity increased by another 22 m yr^{-1} , for an average of $128 \pm 2 \text{ m yr}^{-1}$ in 2013 (Figure 13). The speedup increased substantially in 2014 (mean $209 \pm 1 \text{ m yr}^{-1}$) and 2015 (mean $359 \pm 15 \text{ m yr}^{-1}$) as Fisher Glacier began a period of rapid acceleration (Figure 13, Figure 14).

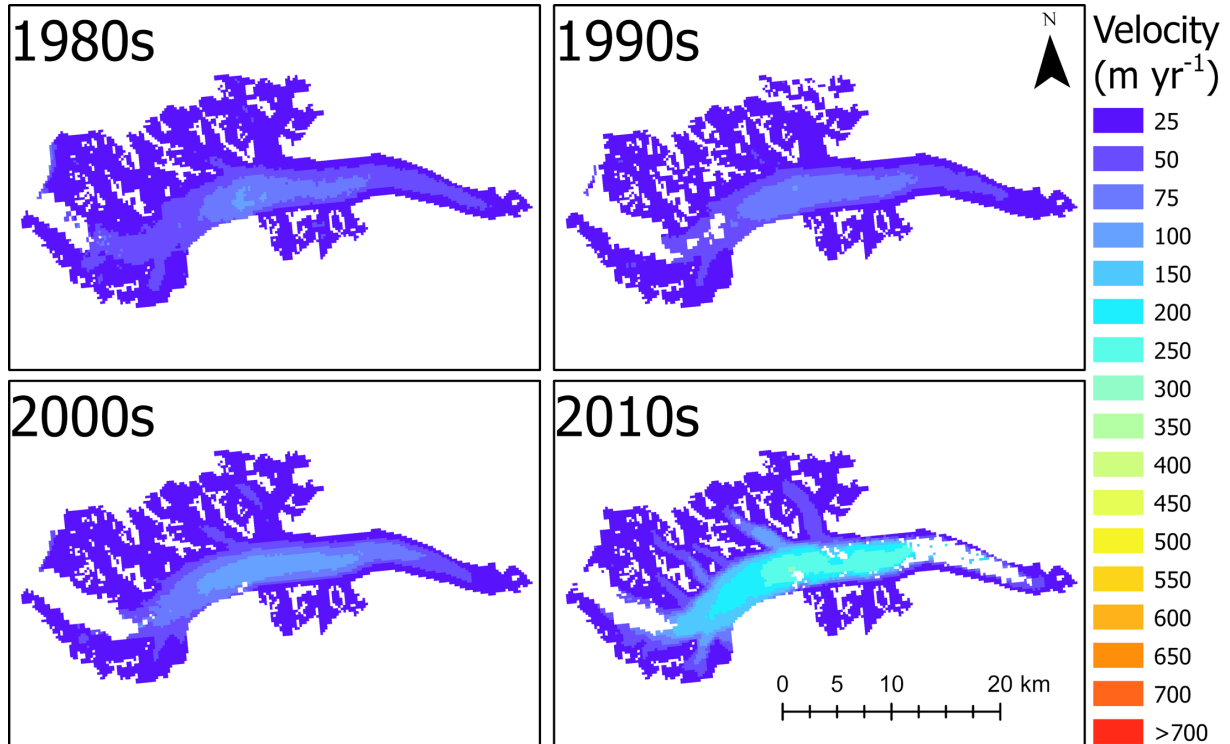


Figure 12. Decadal surface velocity mosaics (m yr^{-1}) of Fisher Glacier covering the period 1985-2018 (1980s = 1985-1989; 1990s = 1990-1999; 2000s = 2000-2009; 2010s = 2010-2018). These mosaics display the mean velocity derived from all ITS_LIVE annual mosaics available for each decade. All data obtained from the NASA MEaSUREs ITS_LIVE project.

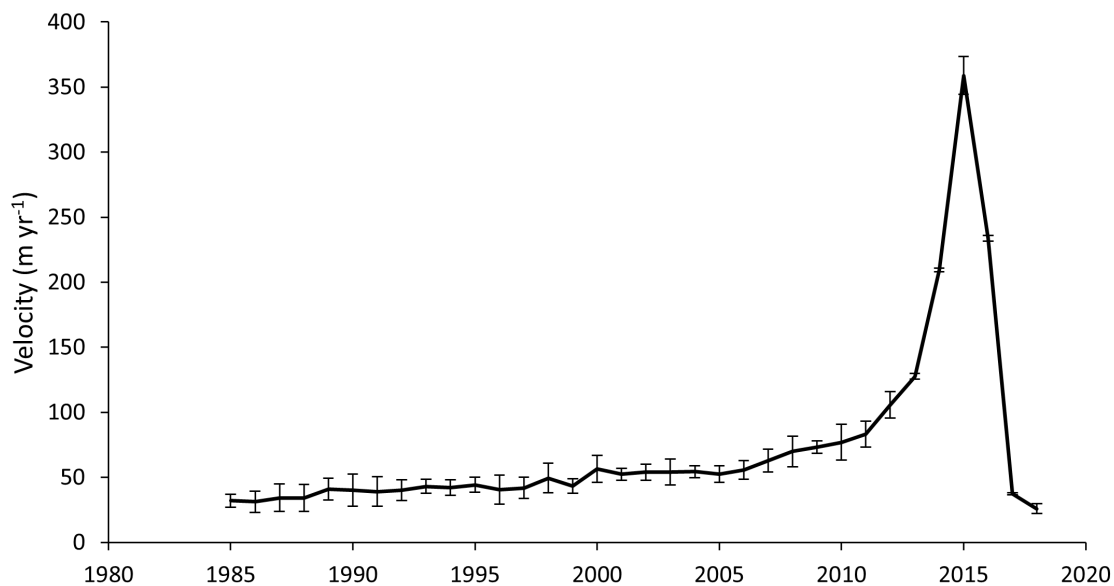


Figure 13. Average annual centerline velocity (1985-2018) of Fisher Glacier, calculated from the mean of all cells within 500 m on either side of the centerline over a distance of 0-45 km upglacier from the terminus. Figure 2 shows the centerline.

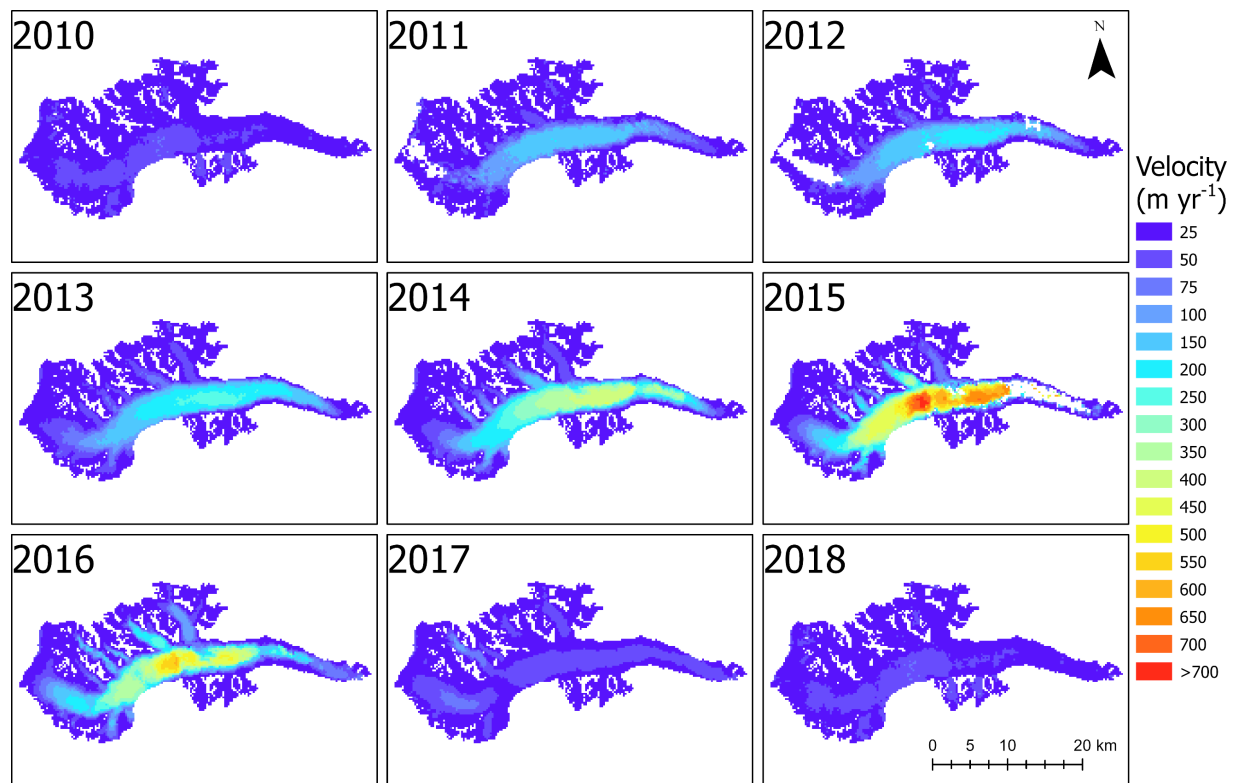


Figure 14. Annual surface velocity mosaics (m yr⁻¹) of Fisher Glacier covering the period 2010-2018. All data obtained from the NASA MEaSUREs ITS_LIVE project.

The annual centerline velocities increased yearly since 1985, starting at the 25 km mark and slowly spreading up- and downglacier (Figure 15). The increase is very gradual at first, with higher speeds associated with the pre-surge buildup starting around 2008, when the velocities surpassed 100 m yr⁻¹ in the region from 22-29 km from the terminus. Because these values cover a small portion of the glacier, the local speedup was not captured in the glacier-wide average of Figure 13. In Figure 15, the following years' peak velocities (i.e., 2009) can be seen slowly increasing, both up and down glacier from where they first started. In 2012, they passed the 150 m yr⁻¹ mark, and in 2013 they climbed above the 200 m yr⁻¹ mark (Figure 15).

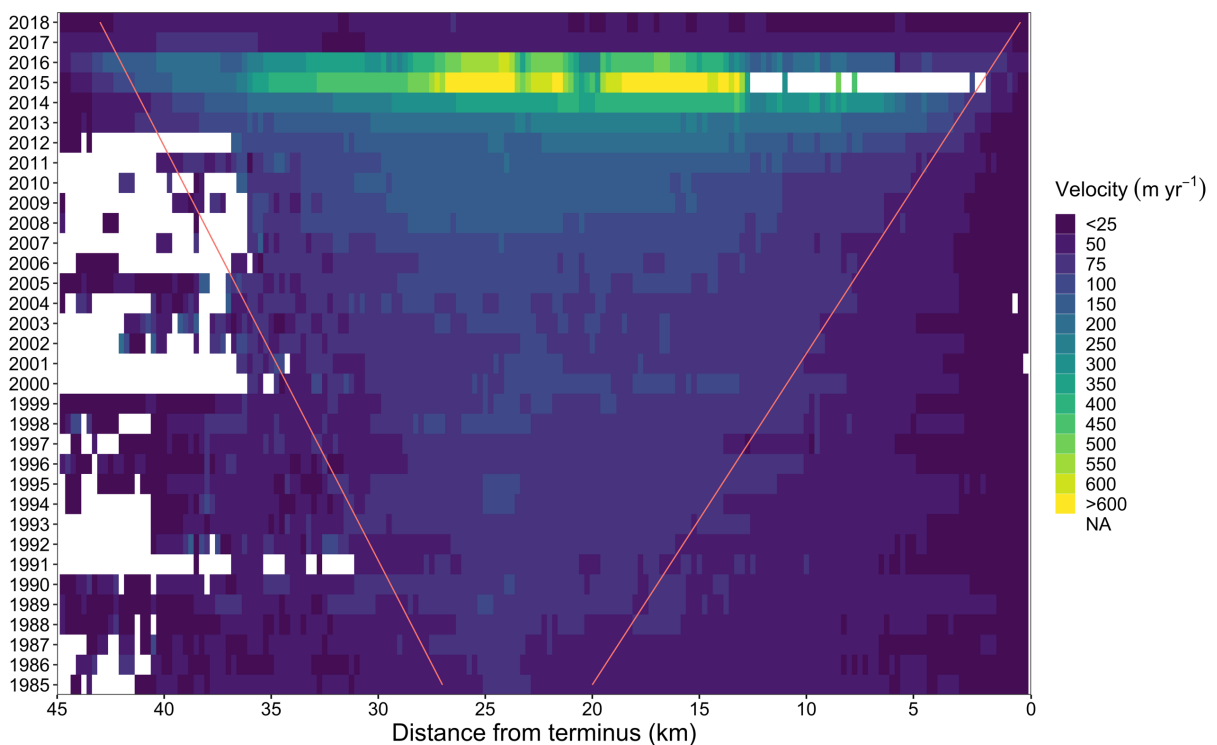


Figure 15. Annual ice velocities from 1985-2018 in 250 m distance bins along the centerline of Fisher Glacier. Velocities were extracted from the ITS_LIVE annual mosaics. Red lines are the approximate boundary between <50 m yr⁻¹ and >50 m yr⁻¹, showing the long-term patterns of the speedup starting from the middle part of the glacier.

The velocity scene pairs obtained from ITS_LIVE and SAR sources show further detail concerning the evolution of velocities throughout the period of rapid surface movement that lasted from 2013/14 to 2016 (Figure 16).

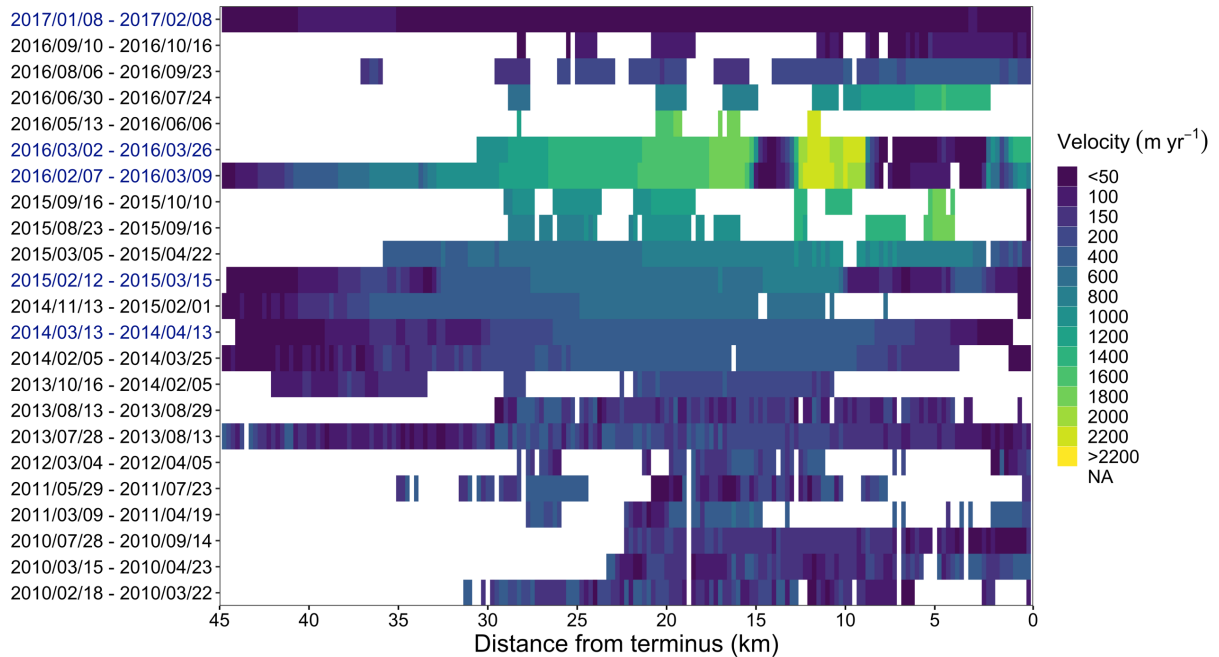


Figure 16. Ice velocities along the centerline of Fisher Glacier at key moments during the period of rapid velocity increase from 2014-2016, together with the buildup and slowdown phases. Velocities were primarily obtained from ITS_LIVE scene pairs and were supplemented with wintertime SAR-derived velocities (identified in blue on the y axis).

In the summer of 2013 (28 July to 13 August 2013), velocities in the midsection of the glacier were an average of $\sim 175 \text{ m yr}^{-1}$, with small areas reaching up to $>300 \text{ m yr}^{-1}$. These enhanced velocities were limited spatially to the middle section, from around 16 km to 29 km from the terminus. The glacier then slowed a little during the late summer and winter of 2013/14. In the early spring of 2014 (5 February to 25 March), the velocities increased again in the mid-glacier with an average of 238 m yr^{-1} between 12.5 km and 27.5 km (showing an increase both up- and down-glacier). The speedup also began to be noticeable downglacier at this time, with velocities of $>200 \text{ m yr}^{-1}$ reaching to 10 km from the terminus, and velocities of $>100 \text{ m yr}^{-1}$ being observed in the upper terminus area from $\sim 4\text{-}9 \text{ km}$ from the terminus, which had previously been nearly stagnant ($<50 \text{ m yr}^{-1}$). One month later, in late spring 2014, the $>200 \text{ m yr}^{-1}$ velocities had spread both up- and downglacier by an additional kilometer (Figure 16). There is a gap in the data throughout summer 2014, but by winter 2014/15 the lower 25 km of the glacier had a mean velocity of 477 m yr^{-1} , with a maximum velocity of 557 m yr^{-1} at 12 km; the more moderate velocities of $200\text{-}400 \text{ m yr}^{-1}$ spread further upglacier, reaching the 37 km mark.

Throughout spring 2015 (February to April), velocities kept increasing both upglacier and downglacier. The lower 12 km saw velocities of $>800 \text{ m yr}^{-1}$, while speeds of $>400 \text{ m yr}^{-1}$ spread upglacier to $\sim 33 \text{ km}$. In late summer and fall 2015 (August to October) spatial coverage is patchy, but velocities clearly kept increasing along the entire length of the glacier, which had an average of $\sim 1000 \text{ m yr}^{-1}$. The lower glacier (4-6 km from the terminus) was flowing the fastest at a mean of 1630 m yr^{-1} .

The winter and spring of 2016 is when the greatest velocities occurred (Figure 16). Between 7 February and 9 March 2016 the extremely rapid velocities finally reached the terminus, which had previously remained below 100 m yr^{-1} . At this time, the terminus attained a mean velocity of $\sim 700 \text{ m yr}^{-1}$ and increased to $>1100 \text{ m yr}^{-1}$ from March 2 to 26, 2016. February 2016 also marked the time where the upper glacier reached its peak velocities. Velocities of $>700 \text{ m yr}^{-1}$ were observed up to 36.5 km, speeds of $>400 \text{ m yr}^{-1}$ reached up to 38.75 km, and speeds of $>200 \text{ m yr}^{-1}$ reached up to 41.25 km from the terminus. Therefore, the velocity changes affected approximately 91% of Fisher Glacier's length (Figure 14; Figure 16). Surface velocities remained high throughout the early summer of 2016. A velocity measurement of 2157 m yr^{-1} , recorded between 13 May to 6 June 2016, is the highest recorded (Figure 16; Figure 17). The highest velocities occurred around 12 km from the terminus (Figure 16).

At the end of June 2016, flow speeds in the mid-glacier (12-21 km) dropped by $\sim 1000 \text{ m yr}^{-1}$ in the span of a month (from a mean of 1758 m yr^{-1} in early June to 735 m yr^{-1} in July). Between July and September 2016, the velocities dropped again, this time to a mean of 162 m yr^{-1} in the mid-glacier. The lower ablation area and terminus velocities lagged behind slightly, as they remained at speeds of $>1300 \text{ m yr}^{-1}$ until late July, before rapidly slowing to $\sim 200 \text{ m yr}^{-1}$ in August/September (Figure 16). Unlike the slow buildup that initiated this period of rapid flow, the termination occurred abruptly. The velocities dropped from a mean of $>1700 \text{ m yr}^{-1}$ to $<100 \text{ m yr}^{-1}$ in less than 4 months in 2016 (Figure 16, Figure 17). By January/February 2017, the mean ice velocity was only 34 m yr^{-1} , and this remained similar through 2018 (Figure 14, Figure 15).

During the short period of rapid velocities from 2013-2016, the three pre-existing looped moraines along the southern ice margin near Tributaries 3-5 progressed downglacier by an average distance of $1017 \pm 8 \text{ m}$, while a new looped moraine was created by the rapid ice velocities (Figure 11b).

ITS_LIVE point data from 2017-2022 shows that Fisher Glacier undergoes regular midsummer speedups (typically July and August) that average $64 \pm 19 \text{ m yr}^{-1}$ (Figure 17). However, these faster flow speeds are short-lived and the mean velocity during the rest of the year remains low ($47 \pm 28 \text{ m yr}^{-1}$ from 2017-2022). Additionally, the mid-summer speedups are significantly slower than speeds observed during the periods of flow instability. In comparison, Figure 17 shows that the period of rapid flow from 2013-2016 had a mean velocity of $348 \pm 40 \text{ m yr}^{-1}$. Figure 17 also shows that there was a particularly elevated summer speedup in summer 2021, though it is unclear what may have caused such a speedup and it was not reflected in the changes in terminus length. Therefore, the values showing velocities $>1000 \text{ m yr}^{-1}$ are most likely outliers or very short and local speed events (they are almost exclusively reported at Point 4, ~27 km upglacier from the terminus).

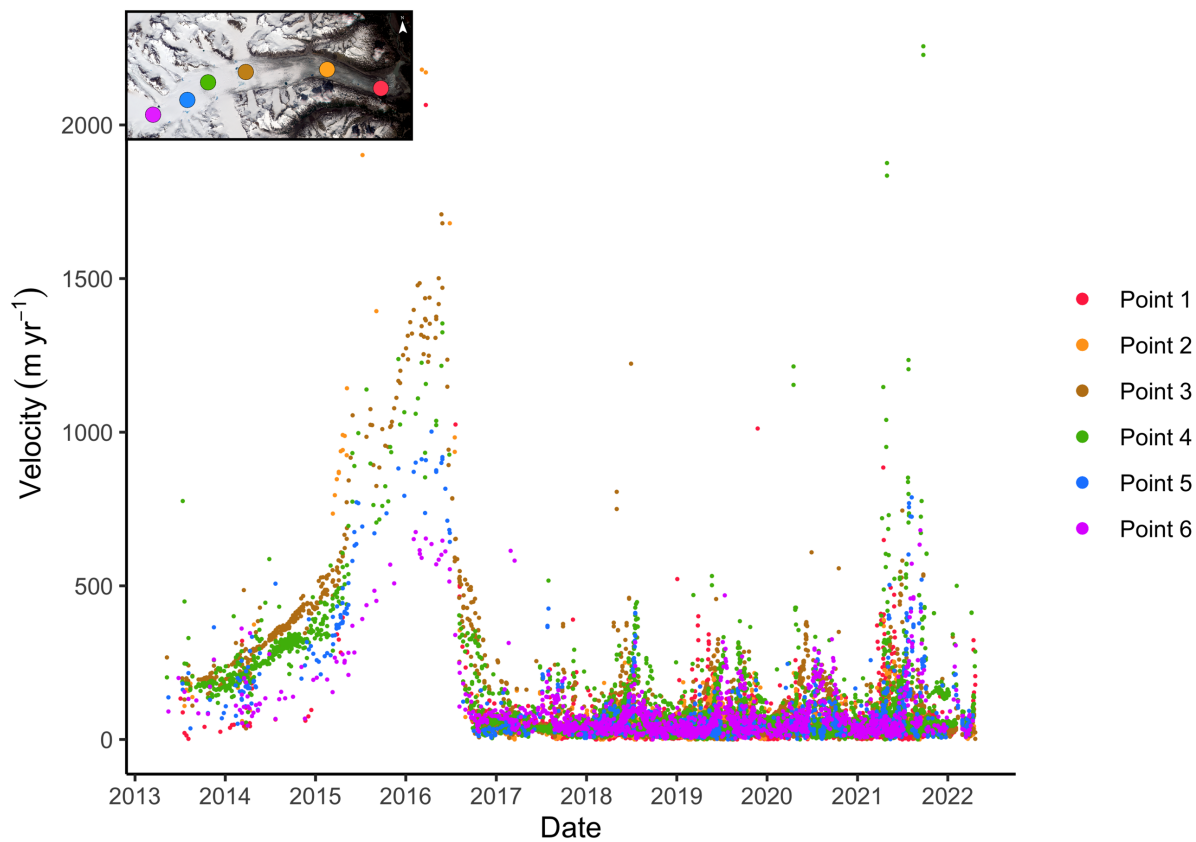


Figure 17. Point velocities from 6 locations along the centerline of Fisher Glacier. Inset map shows the point locations with corresponding colours. Point locations 1-6 are located at 2 km, 10 km, 22 km, 27.5 km, 31.5 km, and 37 km from the terminus, respectively. Data downloaded from the NASA MEaSUREs ITS_LIVE project.

2.4.3 Surface feature changes

2.4.3.1 Surface feature changes between 1948 and 2010

In the 1948 and 1954 historical air photos, the glacier surface appears smooth (e.g., as seen in Figure 18a in a close-up of the terminus). In 1954, potholing is present from ~15 km from the terminus to at least 30 km upglacier, beyond which no imagery is available. The area from 15-25 km is where the pothole fields are most dense (Figure 19) and occurs in the region where two large tributaries join from the northern margin, and three smaller tributaries join from the southern margin (Figure 2, Figure 19d). The potholes generally aggregate in clusters, and form three distinct fields just past the mouth of tributary glaciers #2, #3, #4 and #5 (Figure 19d), although only Tributary 2 (the largest one) displays traces of potholing directly on its surface.

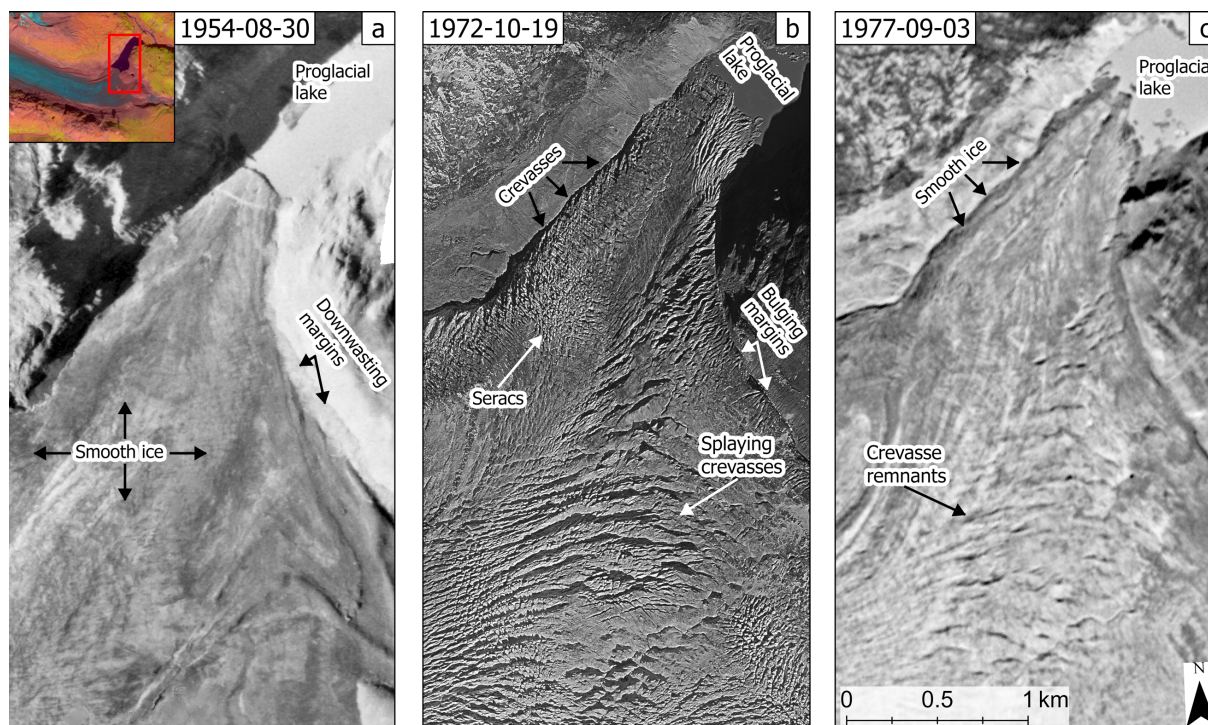


Figure 18. Close-up images of the northern section of Fisher Glacier's terminus in: (a) 1954, (b) 1972, and (c) 1977 that display surface feature changes. The inset map shows a view of Fisher Glacier's lower ablation area with a box showing the section of the terminus displayed in this figure. Images were obtained from the NAPL (A11459) and the USGS declassified satellite imagery archive (D3C1204-100125A001, DZB1213-500139L002001), respectively.

Pothole Field 1 (referred to as PF1, PF2, etc. from here onwards) is the largest and has just under 100 potholes (Figure 19a). They vary in shape from circular to oval, and in width from 10 m to ~130 m. PF2 has 38 potholes of similar size and shape to PF1. PF3 has 27 potholes, although some are larger than in the other fields, possibly due to coalescence of smaller potholes over time.

In addition to these large clusters, smaller groupings and individual potholes can also be observed on the glacier surface above 20 km.

In the 1972 image, nearly all the potholes have disappeared due to the extensive crevassing that covers the glacier, with only a few of the largest potholes in PF3 still visible (Figure 19b). In the lower ~7 km of the glacier crevasses are widespread across the entire ice surface at this time (Figure 18b, Figure 20d), whereas above this some areas are less crevassed, although not completely unaffected (particularly along the center of the glacier) (Figure 20c). The terminus in 1972 is characterized by a chaotic assemblage of longitudinal, transverse, and splaying crevasses which intersect to form serac fields in certain areas (Figure 18b, Figure 20d). Additionally, the ice margins, which were in a state of downwasting in 1954, are bulging out in 1972 (Figure 18b, Figure 20c, d). Furthermore, the texture of the snow up to 33 km from the terminus, and on the upper tributaries, suggests the presence of snow-covered crevasses in the upper glacier (Figure 20b). The intense surface changes in 1972 therefore extended far up the glacier's length, reaching at least as far as the upper tributaries (indicator box "b" in Figure 20a).

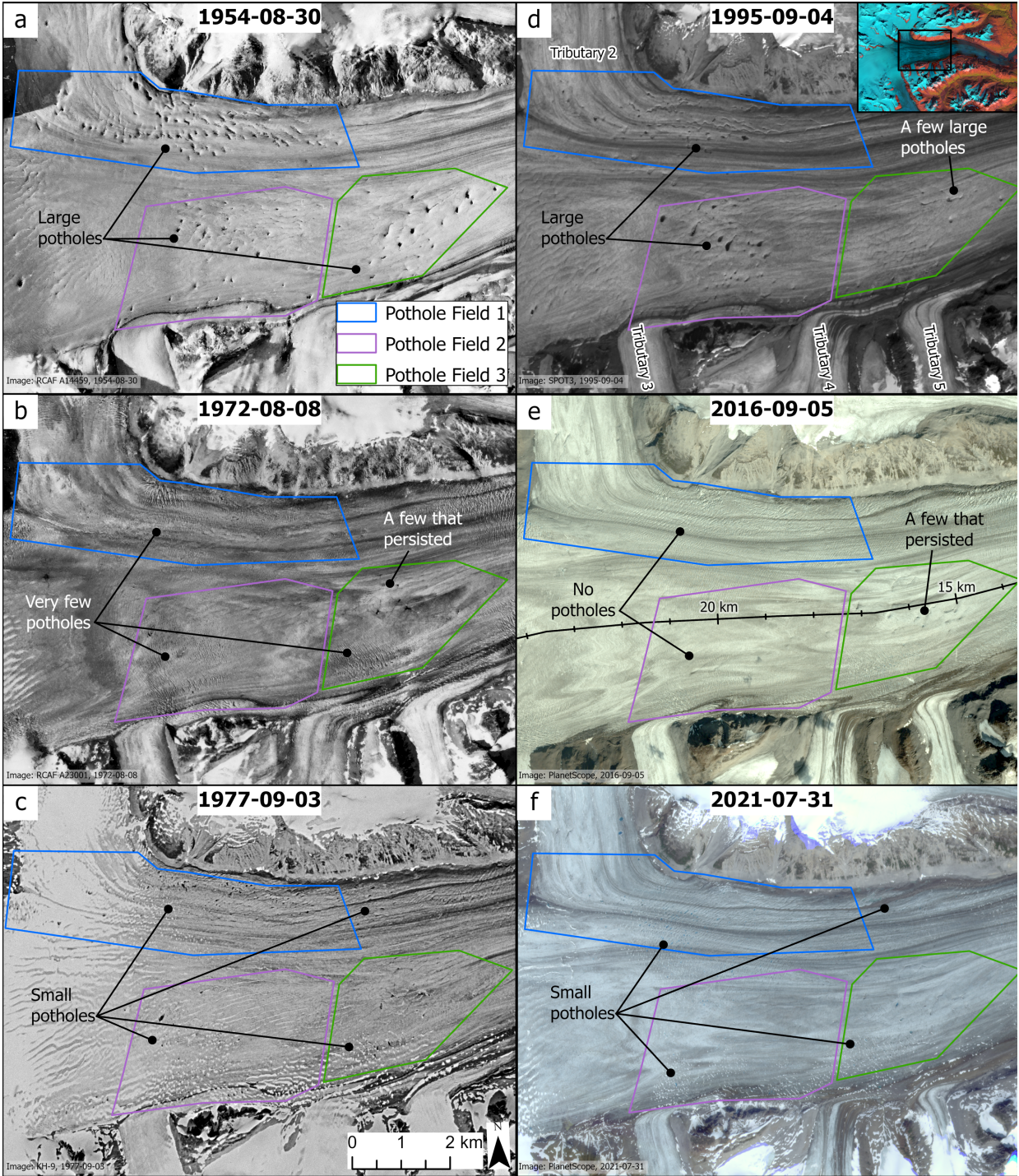


Figure 19. Evolution of potholes on Fisher Glacier through two full appearance/disappearance cycles in the three main pothole fields in (a) 1954, large potholes present; (b) 1972, almost no potholes; (c) 1977, small potholes reappearing; (d) 1995, large potholes; (e) 2016, almost no potholes (centerline distance markers shown on black line); and (f) 2021, small potholes reappearing. The inset map shows the location of the pothole fields near the mid-point of the glacier.

In 1972, a supraglacial lake measuring 450×220 m, that was not present in 1954, was visible near the snout of Tributary 1. By 1976 it had shrunk significantly to 190×125 m. In the 1976/77 and later imagery, the lake is no longer visible until it returned in approximately the same location in 2016. Of note is that Tributary 1 was one of the only tributaries that displayed a significant speed up during the period of rapid velocity change from 2013-2016 (Figure 14).

The images from 1976 and 1977 show that the glacier surface had smoothed-out, although remnants of large crevasses from 1972 were still present along the glacier margins and at the terminus (Figure 18c). It did not take long for the potholes to reappear, because in 1977 PF1 was already covered by >100 small potholes, although the field covered a smaller area, and the potholes were about half the size of those in 1954 (Figure 19c). In 1976/77, PF2 and PF3 had only partially started reappearing (Figure 19c).

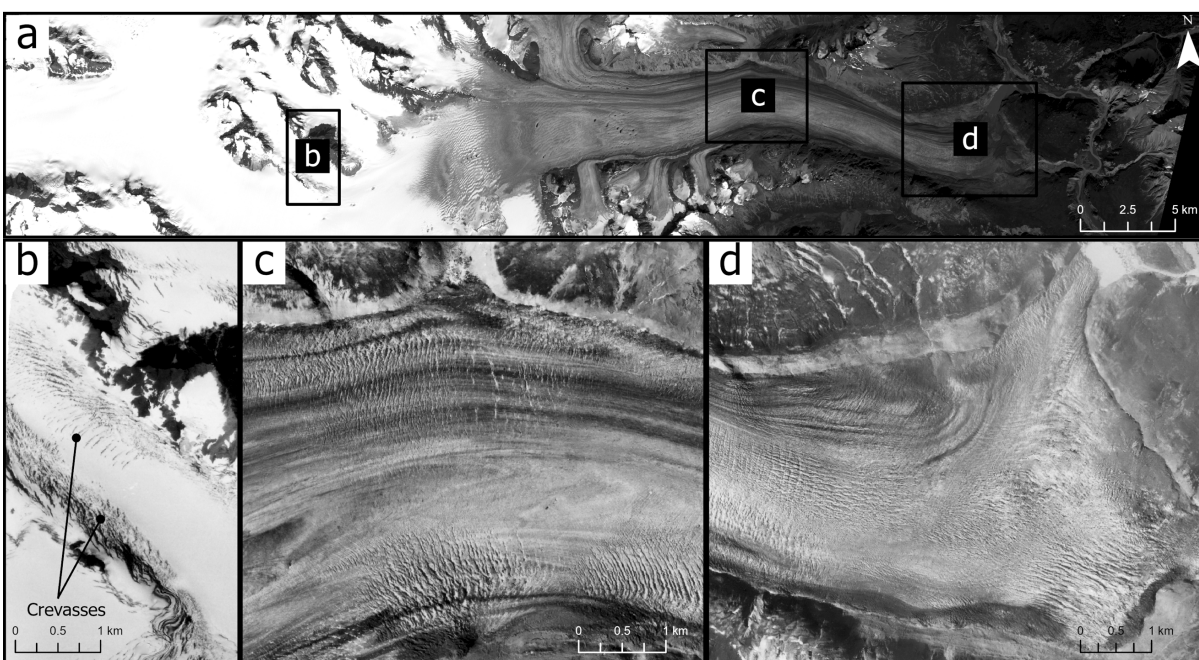


Figure 20. Crevasses at different locations on Fisher Glacier in 1972. (a) Image showing the extents of areas shown in (b)-(d). Base image: SPOT 3, 1995-09-04. (b) Crevasses forming in one of the upper tributaries of the upper glacier. (c) Crevasses along the margins in the ablation area at ~12 km. Note that large crevasses are mostly limited to the margins, with the center having a generally smoother surface. (d) Widespread crevasses covering the entire terminus and lower ablation area. Image in (b)-(d) is RCAF air photo A23001 (1972-08-08).

The Landsat 1-4 imagery from 1978-1983 is too coarse resolution to distinguish any significant feature changes. However, as of 1984 (and subsequent years) the finer resolution of

Landsat 5 and other imagery allows us to see that potholes on Fisher Glacier now extended far above the snowline to 37 km (Figure 21a), although the densest distribution was still located beneath Tributary 2 (i.e., around PF1) (Figure 21b). PF2 and PF3 reappeared at some point between 1977 and 1984, although they were not as densely populated as in 1954 (Figure 19d). The glacier surface displayed little to no significant change throughout the rest of the 1980s and 1990s. By the early 2000s, the total number of potholes had reduced, although those that remained were generally larger than before, likely due to the coalescence of previous smaller potholes in the mid-region of the glacier (Figure 21b-c). In 2010, PF1 only had ~40 potholes while PF3 had just over 20 and PF2 no longer existed (this can be seen on the background images in Figure 11b).

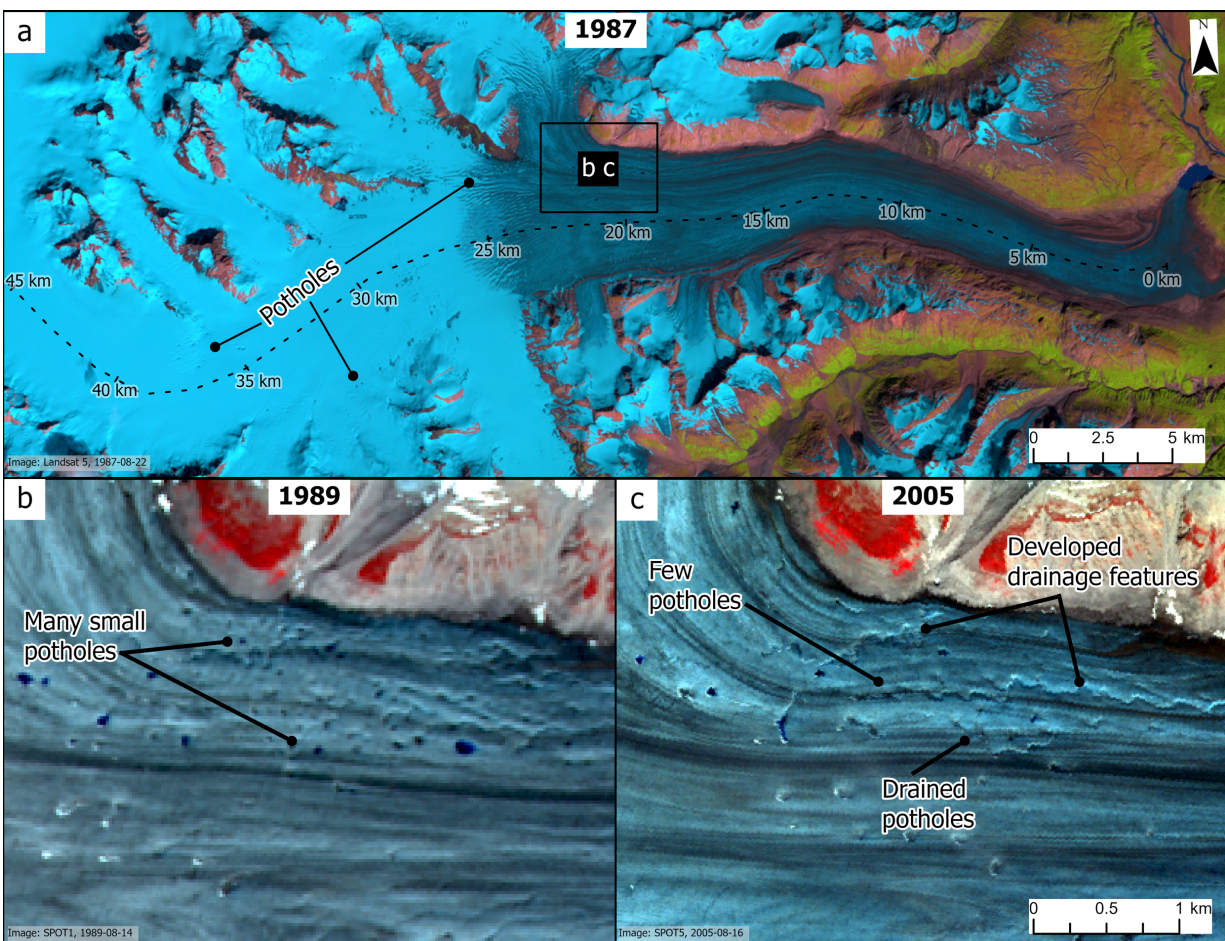


Figure 21. Pothole evolution on Fisher Glacier from 1987 to 2005, during a period of slow stable flow. (a) Image from 1987 shows pothole distribution in the upper glacier. Box indicates extent of (b) and (c). (b) Close-up of Pothole Field 1 in 1989: potholes were numerous, water-filled, and generally of small size. (c) Close-up of Pothole Field 1 in 2005: many potholes had disappeared or drained, and many new supraglacial drainage features have developed.

2.4.3.2 Surface feature changes between 2010 and 2021

The period ~2014-2016 was marked by a series of rapid and pronounced surface feature changes. The potholes almost entirely disappeared or drained between 2011 and 2014. At the end of July 2014 (i.e., date of minimum terminus extent), a RapidEye-3 image shows that the glacier surface was still smooth throughout its entire length. However, the first imagery available in the spring of 2015 shows the beginnings of a number of surface changes, including new crevassing, the formation and drainage of supraglacial lakes, and disappearance of potholes, that would continue until late 2016.

Between 2 September 2014 and 16 May 2015, a number of new supraglacial lakes formed in the snow on the glacier surface (Figure 22a). In only 5 days, between 16-21 May 2015, *Lake 1* more than doubled in size from 0.015 km² to 0.036 km². Over the same period, *Lake 2* also more than doubled in size, increasing from 0.024 km² to 0.058 km². *Lake 3* measured 0.024 km² on 21 May and had rapidly increased to 0.167 km² by 22 June 2015 (Figure 22a). After this rapid formation, all three lakes had disappeared by 8 August 2015. Rapid and short-lived changes in surface hydrology such as these were common nearly everywhere on the glacier's surface between 2015 and 2016 (Figure 22). For example, another small lake (0.021 km²) developed along the margin (60°06'59" N, 138°09'57" W) sometime before 3 July 2015, grew to 0.044 km² by 31 July 2015, and then halved in size to 0.021 km² by 8 August 2015. By summer 2016, this lake had completely disappeared as the ice pushed up higher along the margin and became increasingly crevassed. It was also common for many of the large crevasses to be water-filled in spring (May) and drain through the summer. In most cases, the drainage features that developed on the surface always reappeared in approximately the same place, regardless of how far the ice had moved downglacier.

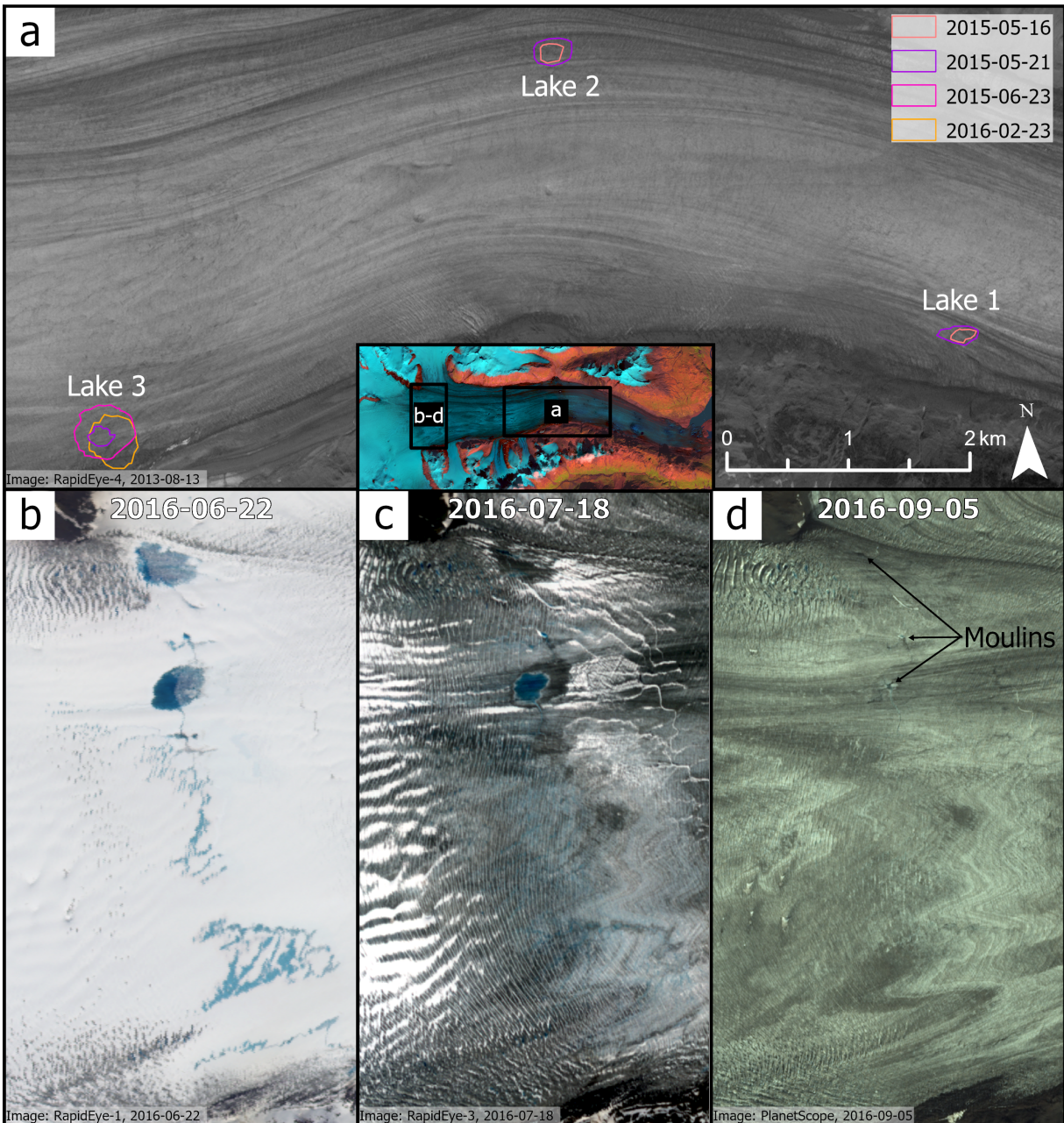


Figure 22. Examples of rapid changes to surface drainage features on Fisher Glacier from 2014-2016. (a) Evolution of supraglacial lakes #1-3. The date of each lake outline is identified by colour. (b)-(d) Changes undergone by a large drainage network on 22 June, 18 July, and 5 September 2016, respectively. The drainage feature consists of a series of interconnected lakes cutting across the glacier's width. Panel (d) shows the moulins through which the water drained. The scale is the same in all figure parts. Inset map shows the location of all panels (b-d are all the same extent).

Extensive crevassing began spreading across the lower glacier during winter 2014/15, or in spring 2015, as crevasses opened from the terminus up to ~15 km upglacier (Figure 23b, e, f).

At this point, the crevasses were mostly located along the glacier margins. A Gigapan (large panoramic) photo, and oblique photos captured from a handheld camera by Dr. Luke Copland in August 2015, confirm the extensive crevassing in the lower ablation area and terminus (Figure 23). A section of the 2015 Gigapan photo (Figure 23e) shows that the glacier's surface is very broken up and characterized by undulating hills on a surface that is typically smooth and flat when not surging (Figure 23h). Large seracs are also observed both at the terminus (Figure 23a) and a little upglacier in the lower ablation area (Figure 23g). The terminus is heavily fractured and terminates in the northern proglacial lake with a defined calving front characterized by an ice cliff (Figure 23a). By the end of summer 2015, the large crevasses at the terminus and along the margins in the ablation zone were generally about 20-30 m wide. The smaller crevasses, which at this point were generally found towards the center of the trunk, measured a maximum of about 8-12 m in width.

Crevasse formation was amplified across most of the glacier between summer 2015 and summer 2016 (compare Figures 23 and 24). RapidEye-1 imagery from 22 June 2016 shows very large crevasses and serac formation at the terminus, with crevasses generally measuring 20-40 m at their widest point, although some reached widths of >60 m in the area where the ice is extremely broken up and splayed (Figure 24d). The crevasses upglacier from the terminus had now spread across the entire glacier width in most areas and extended up to 37 km upglacier from the terminus, similar to 1972 (Figure 24b, c; compare with Figure 20). This included intense crevassing on the tributaries high in the accumulation zone (Figure 24b). In the ablation area, the larger crevasses were typically 20-30 m wide, while the smaller ones near the center of the glacier were generally <10 m wide (Figure 24c).

A variety of crevasse types developed. Along most of the glacier trunk, transverse crevasses were most common (similar to 1972; compare Figures 20 and 24). The shape of the crevasses suggests that the forces exerted on the ice varied depending on the surrounding topography (e.g., those in Figure 24c are formed by compressive stress along the bend in the southern margin, whereas those in Figure 24b are formed by tensile stress as the ice is pulled down from the tributary). The terminus displayed a chaotic mix of crevasses and seracs oriented in nearly every direction, and, notably, was characterized by large splaying crevasses at the edge of the terminus. The longitudinal crevasse orientation at the terminus and lower ablation area is indicative of compressive stress being applied to the ice.

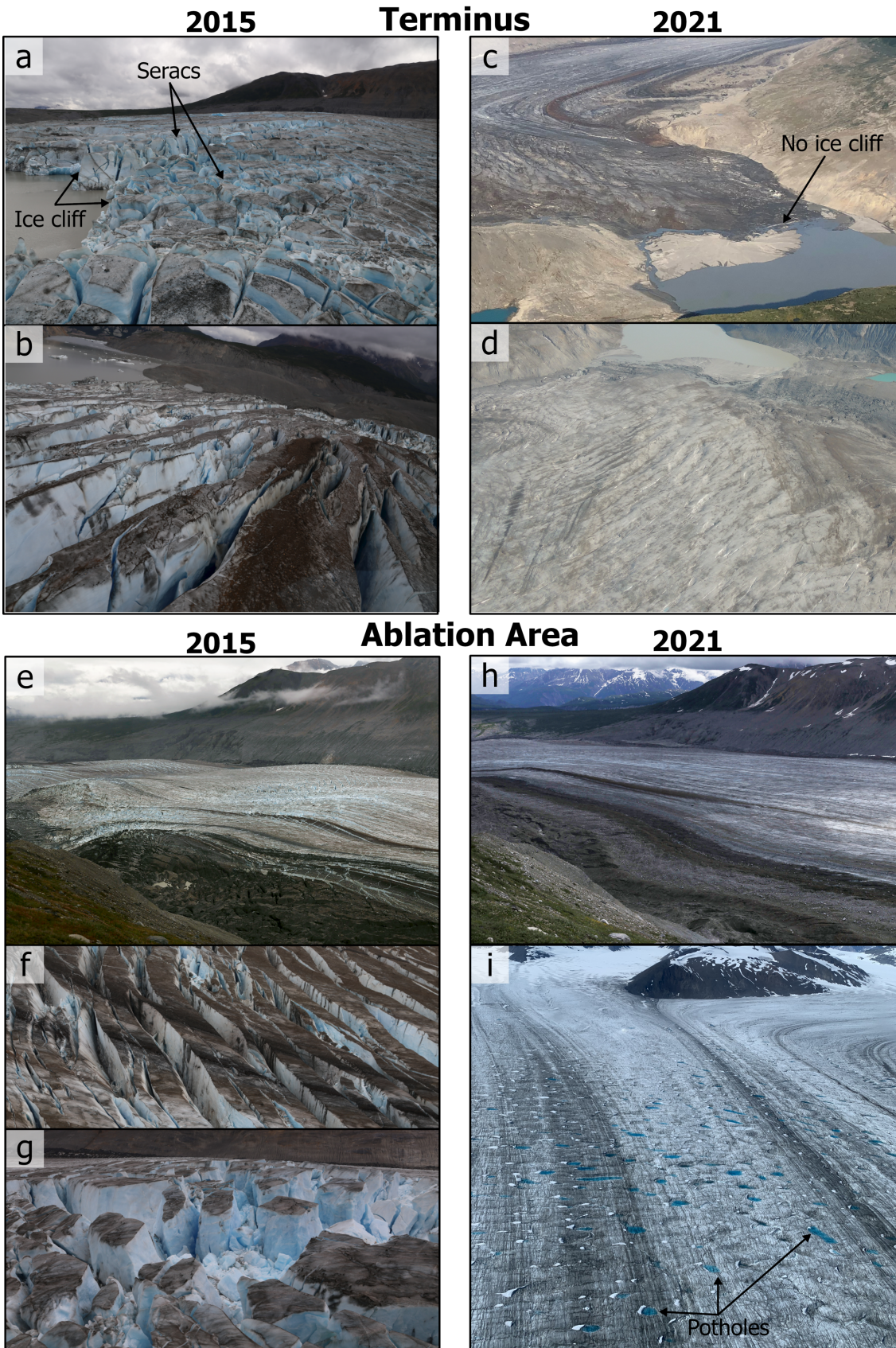


Figure 23 (previous page). Surface feature changes at the terminus (a-d) and in the ablation area (e-i) of Fisher Glacier on 2015-08-11 (left column) and 2021-07-25 (right column), taken from aircraft and a hill on the northern margin. (a) August 2015 image detailing the seracs and ice cliff that formed where the terminus meets the proglacial lake along the northern section of the terminus. (b) August 2015 image showing large crevasses that formed at the terminus, a little further upglacier from what is seen in (a). (c) July 2021 image showing where the northern terminus meets the proglacial lake. Note the absence of a pronounced ice cliff and the disappearance of the crevasses and seracs. (d) Image displaying the smooth and flat surface of the terminus in July 2021. (e) Section of the August 2015 Gigapan photo, looking generally South and downglacier. The ice surface is broken up with extensive crevassing and is characterized by a hilly, undulating topography. (f) Close-up on crevasses in the ablation area. (g) Close-up on seracs in the ablation area. Blue ice is indicative of recently exposed areas, suggesting an actively changing landscape. (h) Section of the July 2021 Gigapan photo, looking generally South and downglacier. The ice surface is smooth, flat, and generally a little higher than it was in (e). (i) Potholes in PF1 on an otherwise featureless/flat surface. For context, Tributary 2 can be seen veering off to the right in the top right corner of the image.

A June 2016 satellite image also shows the formation of a number of new supraglacial lakes (e.g., the lake along the southern margin in Figure 24c) and the reappearance of *Lake 3* that had first formed in spring 2015 (orange outline in Figure 22a; slightly displaced from where it was the previous year). Of particular interest is a large drainage network composed of supraglacial lakes interconnected by streams which cut across the glacier's width between Tributaries 1 and 2 (Figure 22b-d). This drainage feature coincided with the area where the velocities first began increasing (~ 23 km from the terminus) and drained at around the same time that the velocities in the mid-glacier dropped by 1000 m yr^{-1} in June/July 2016. Figure 22b-d demonstrates how most of the water had drained by 18 July 2016, and the feature had completely disappeared by 5 September 2016. In fact, nearly all surface drainage features on the glacier had drained or disappeared by 5 September 2016.

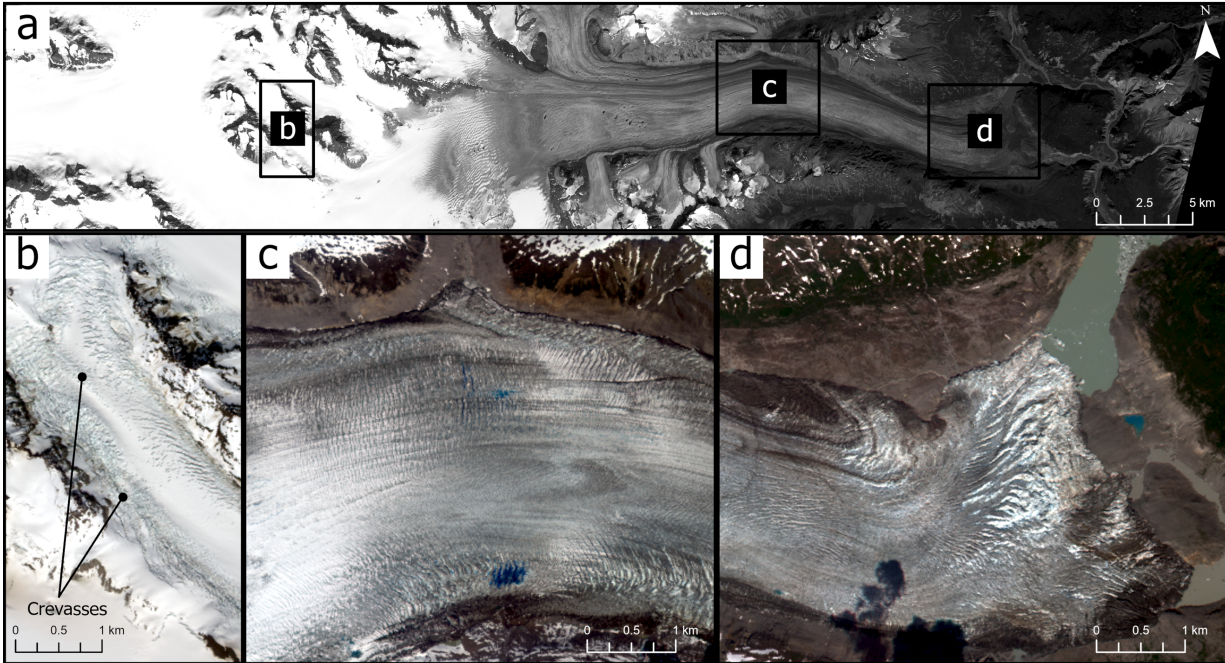


Figure 24. Crevasses at different locations on Fisher Glacier in 2016. (a) Image showing the extents of areas covered in (b)-(d). Base image: SPOT 3, 1995-09-04. (b) Crevasses forming in one of the upper tributaries of the upper glacier. (c) Crevasses along the margins in the ablation area. Note that large crevasses are mostly limited to the margins, with the center having smaller crevasses and some smooth sections. (d) Widespread crevasses covering the entire terminus and lower ablation area. (b)-(d) base image: RapidEye-1, 2016-06-22.

During the summer of 2017, the glacier surface looked much the same as it had in 2016. Most crevasses were still open, but not expanding, and there were a number of new surface drainage features. The photographs from 25 July 2021, four years later, shows that the ice was much smoother and only crevasse remnants can be distinguished on the ice surface (Figure 23c, d, h). The second Gigapan photo provides *in situ* confirmation of this change (Figure 23h). In 2021, we also see the reappearance of potholes covering an extensive region where PF1 used to be located (Figure 19f; Figure 23i).

2.4.4 Surface elevation changes from 1954 to 2022

End of summer snowlines from 2010-2022 suggest that the equilibrium line altitude (ELA) is located at a mean elevation of 1409 m a.s.l., ~38 km from the terminus (nearly at the head of the glacier). Therefore, Fisher Glacier has a low accumulation area ratio (AAR) of approximately 0.23.

DEMs of difference derived from historical images show that from 1954 to 1972 the area of Fisher Glacier below 1050 m a.s.l. experienced thickening at an average rate of $+1.70 \pm 0.03 \text{ m yr}^{-1}$, while the area above 1050 m a.s.l. underwent a period of surface lowering averaging $-1.45 \pm 0.03 \text{ m yr}^{-1}$ (Figure 25a). Over this time, the highest rate of change was $+3.37 \pm 0.03 \text{ m yr}^{-1}$ in the 800-850 m a.s.l. hypsometric band. The dynamic balance line (DBL) is defined as the location where surface elevation change switches from positive to negative (i.e., where there is no net elevation change; Dolgoushin and Osipova, 1975), and was located at 19.9 km from the terminus for 1954-1972. The trends in ice volume change follow a similar pattern to the elevation change trends (Figure 25b). The area above 1050 m a.s.l. underwent a mean volume loss of $-0.02 \text{ km}^3 \text{ yr}^{-1}$ from 1954-1972, with the greatest loss ($-0.056 \text{ km}^3 \text{ yr}^{-1}$) at the 1250-1300 m a.s.l. hypsometric band.

At the terminus, the glacier margin elevations thickened at a mean rate of $2.63 \pm 0.12 \text{ m yr}^{-1}$ over 1954-1972, while the lower ablation area thickened by $8.00 \pm 0.19 \text{ m yr}^{-1}$ (total average elevation increase of $47 \pm 2.10 \text{ m}$ and $144 \pm 3.41 \text{ m}$, respectively) (Figure 26).

From 1972 to 1976, the historical DEM of difference shows that nearly all 50 m hypsometry bands had a negative surface elevation change (Figure 25a). The only positive area (900-950 m a.s.l.) had a mean rate of just $+0.73 \pm 0.32 \text{ m yr}^{-1}$. With that said, the elevation bands at higher elevations underwent the greatest thinning rates (Figure 25a), with the elevation bins from 450-950 m a.s.l. lowering by an average of $-3.10 \pm 0.32 \text{ m yr}^{-1}$, compared to lowering by an average $-5.79 \pm 0.32 \text{ m yr}^{-1}$ above 950 m a.s.l. (Figure 25a). This resulted in the greatest volume loss occurring in the upper parts of the glacier, at an average of $-0.072 \text{ km}^3 \text{ yr}^{-1}$ above 950 m a.s.l., compared to $-0.018 \text{ km}^3 \text{ yr}^{-1}$ below (Figure 25b).

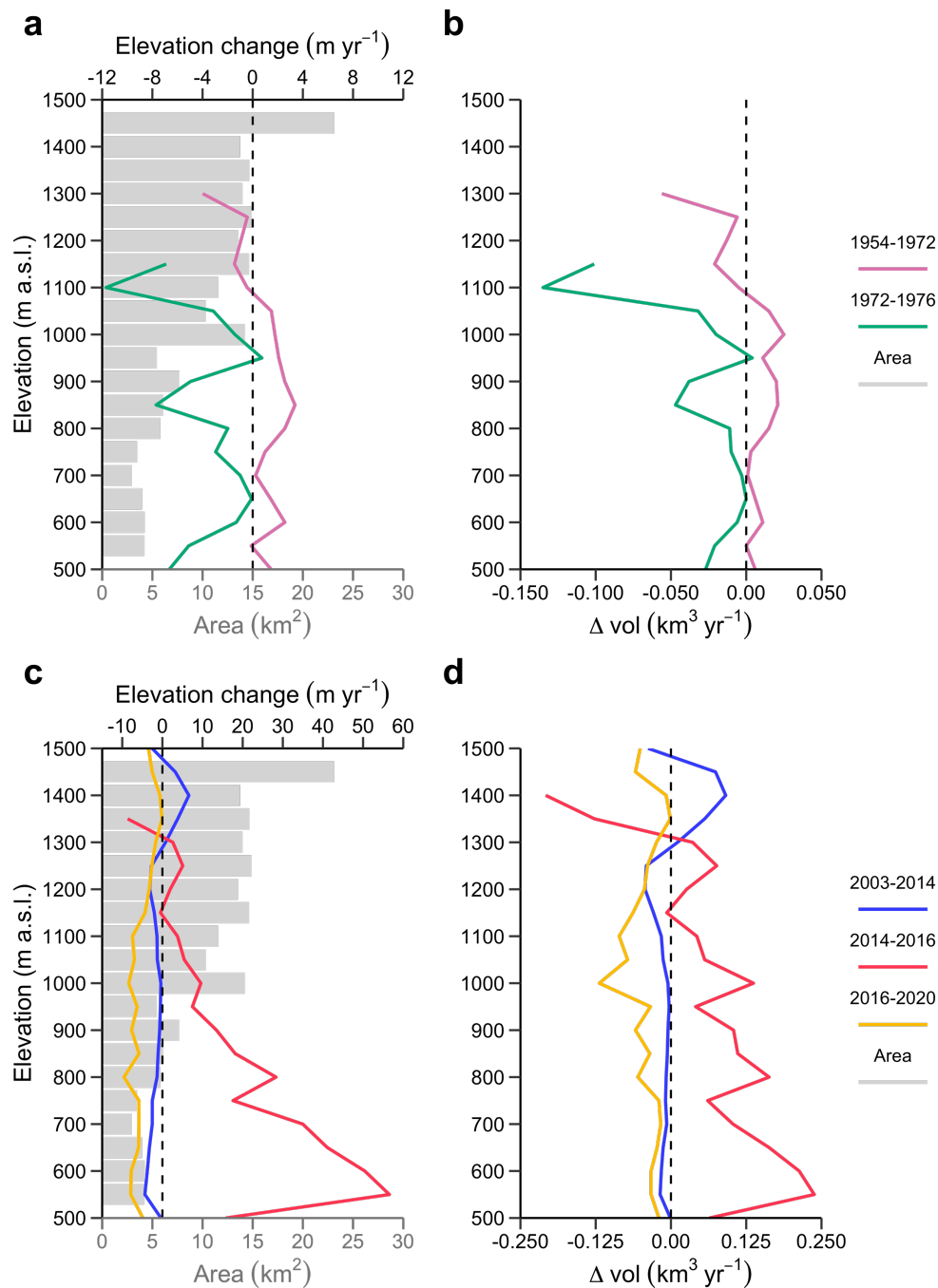


Figure 25. (a) Fisher Glacier hypsometry (grey bars) and average annual surface elevation change per 50 m elevation band for 1954-1972 (pink) and 1972-1976 (green). (b) Mean annual ice volume change per 50 m elevation band for the same periods as in a. (c) Fisher Glacier hypsometry (grey bars) and average annual surface elevation change per 50 m elevation band for 2003-2014 (blue), 2014-2016 (red), and 2016-2020 (orange). (d) Mean annual ice volume change per 50 m elevation band for the same periods as in c. Note the difference in scales between b and d.

The glacier margin elevations show that the ice in the lower ablation area thinned by an average of 50 ± 9.54 m between 1972 and 1976/77 (-11.0 ± 2.11 m yr⁻¹). Conversely, the elevation along the northern margin at the terminus did not change so much, with only a 4 ± 2.83 m surface lowering from 1972-1977 (-0.96 ± 0.63 m yr⁻¹) (Figure 26). Figure 26 shows a clear pattern of thickening from 1954-1972, particularly in the lower ablation area section (Figure 26b; light blue line far above dark blue line), followed by a surface lowering from 1972-1976 (grey line below light blue line).

The next available DEMs are the MMASTER DEMs covering 2003-2020. From 2003 to 2014, the DBL was located at 31.5 km upglacier from the terminus, with the area above this (at >1250 m a.s.l.) increasing in surface elevation at an average rate of $+2.37 \pm 0.27$ m yr⁻¹ (Figure 25c, Figure 27a). This translated to an average annual volume gain of 0.058 km³ yr⁻¹ (Figure 25d). The area below the DBL lowered at an average rate of -1.88 ± 0.27 m yr⁻¹ (Figure 25c), although the volume loss from 450-1000 m a.s.l. was only slightly negative (-0.008 km³ yr⁻¹) (Figure 25d).

Following this, the DEMs from 2014-2016 show a downglacier transfer of mass from above the DBL (1300 m a.s.l.; 31.5 km) to the area below it (compare Figures 27a and 27b). The area below 1300 m a.s.l. underwent an average surface elevation increase of 18 ± 5.58 m yr⁻¹, although this varied spatially. The terminus and lower ablation area (<700 m a.s.l.; 0-4.8 km from the terminus) raised by $+39.77 \pm 5.58$ m yr⁻¹, while the lower- to mid-section of the glacier (i.e., area between 700-1270 m a.s.l., or 4.8-23.7 km from the terminus) only raised by $+9.41 \pm 5.58$ m yr⁻¹ (Figure 25c). The surface elevation changes from the MMASTER difference DEMs suggests that the DBL moved upglacier by ~ 11.5 km (raised by ~ 250 m in elevation) between the historical (1954-1972) and recent (2003-2016) periods.

The northern margin ice elevations support these trends, with a thickening of 15.22 ± 1.45 m yr⁻¹ and 15.07 ± 0.53 in the lower ablation area and terminus, respectively, between 2010-2016 (89 ± 8.48 m and 88 ± 3.12 m higher in 2016 than in 2010) (Figure 26). Figure 26 illustrates this with the brown line (2010) being at a much lower elevation than the orange line (2016).

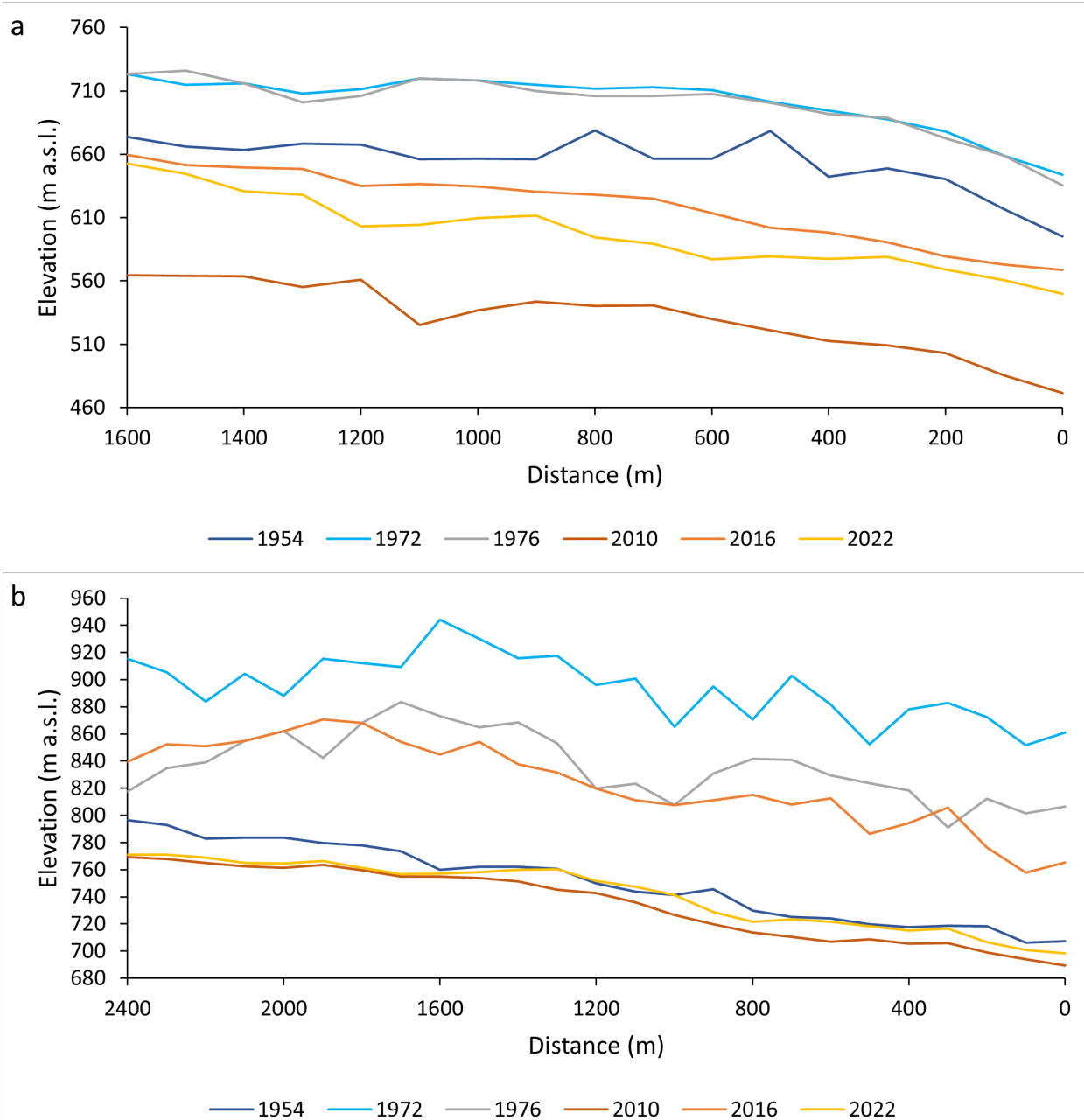


Figure 26. Ice surface elevation along the valley walls in two sections of the northern margin of Fisher Glacier: (a) The terminus section and (b) the lower ablation area section. The x axis labels represent the upglacier distance from the first transect at which velocities were extracted (see Figure 4 for reference).

From 2016 to 2020, Fisher Glacier started another period of widespread surface thinning, similar to what was observed after 1972 (compare the trend of the green line in Figure 25a and yellow line in Figure 25c). The glacier surface now displayed a negative rate of change over its entire length, averaging $-5.19 \pm 1.45 \text{ m yr}^{-1}$, with no evidence for mass gain in the reservoir zone.

The thinning rates were more pronounced in the receiving area (Figure 25c, Figure 27c), and the greatest volume loss occurred at 1000 m a.s.l. ($-0.119 \text{ km}^3 \text{ yr}^{-1}$; Figure 25d). The lower half of the glacier (up to 22.7 km from terminus; $<1100 \text{ m a.s.l.}$) thinned at an average rate of $-6.94 \pm 1.45 \text{ m yr}^{-1}$, while the upper half displayed an average rate of $-2.33 \pm 1.45 \text{ m yr}^{-1}$.

Along the northern ice margin from 2016 to 2022, the lower ablation area section thinned rapidly, lowering by a total of $82 \pm 8.58 \text{ m}$ ($-13.44 \pm 1.41 \text{ m yr}^{-1}$), essentially returning to its 2010 elevation (red and yellow lines in Figure 26b). The terminus area thinned at a more moderate rate of $-3.52 \pm 1.13 \text{ m yr}^{-1}$, resulting in a surface elevation change of $\sim 21 \pm 6.85 \text{ m}$ (Figure 26).

The overall surface elevation trend on the main trunk for 2003-2020 shows that the DBL was located at an average altitude of 990 m a.s.l., which corresponds to a distance of $\sim 16 \text{ km}$ from the terminus (Figure 27d) and is very similar to the DBL for 1954-1972. This means that about 70% of Fisher Glacier's area is located above the DBL. Based on the location of the accumulation zone determined by the ELA, the reservoir zone is shown to be located in the ablation area (it is below the ELA at $\sim 1400 \text{ m a.s.l.}$). Over the entire 2003-2020 period, Fisher Glacier's main trunk underwent a mean thickening of $1.69 \pm 0.34 \text{ m yr}^{-1}$ in the region below the DBL, and a mean surface thinning of $-2.40 \pm 0.34 \text{ m yr}^{-1}$ in the region above the DBL. This produced a glacier-wide net volume change of $-5.14 \pm 1.26 \text{ km}^3$ from 2003-2020. Assuming an ice volume to mass conversion factor of $850 \pm 60 \text{ kg m}^3$ (Huss, 2013), this translates to a net ice equivalent mass loss of $-4.37 \pm 1.38 \text{ Gt}$ ($-0.26 \pm 0.08 \text{ Gt yr}^{-1}$) over this period. Figure 27 also shows that, while most of the mass comes from the upper accumulation area of the main trunk, a few tributaries also contribute. Tributary 1 and the upper tributary both display elevated surface lowering rates compared to the other tributaries over this 17-year period (Figure 27d), and these were the same two that displayed the highest velocities from 2013-2016.

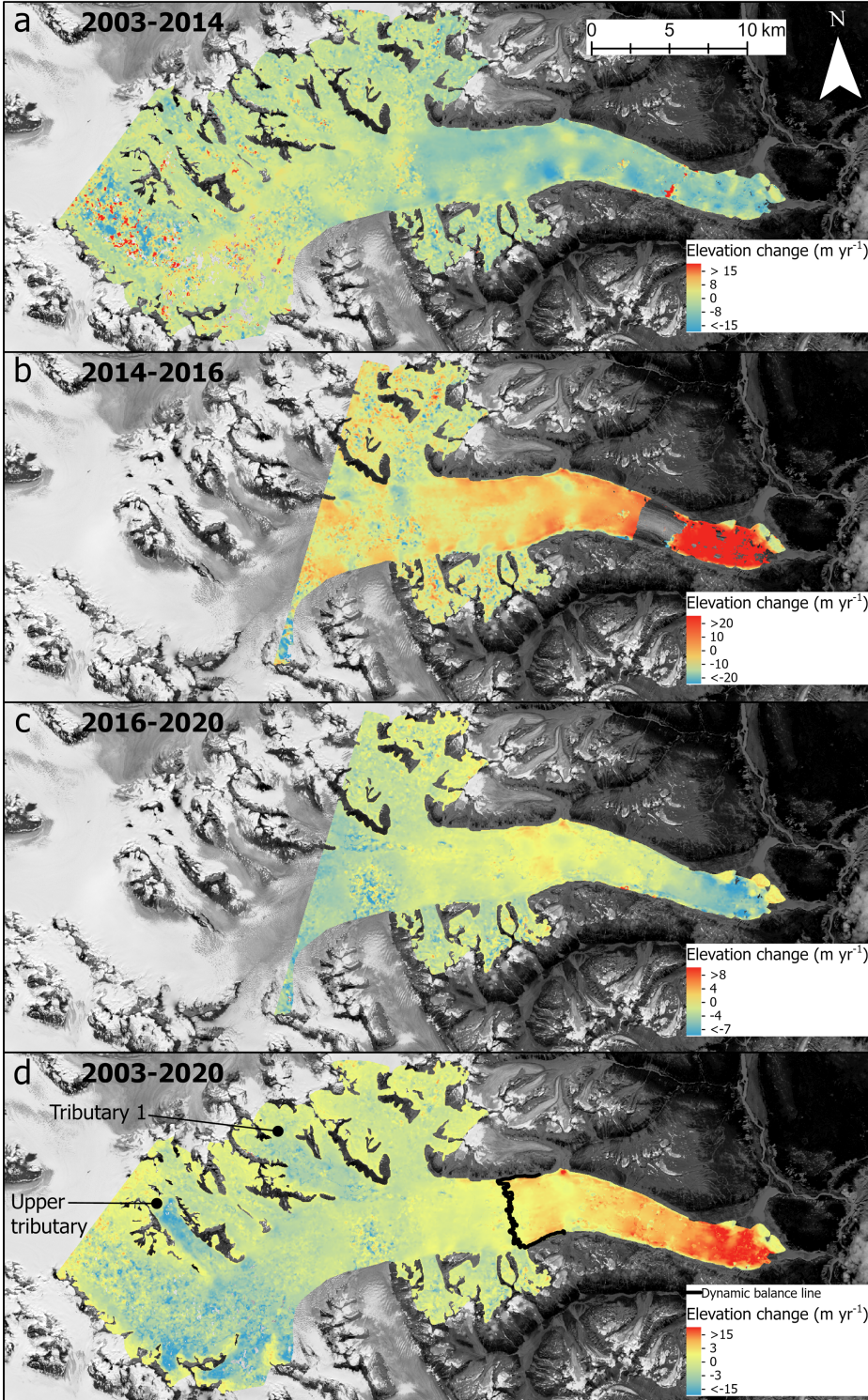


Figure 27. Annual rate of surface elevation change on Fisher Glacier from (a) 2003-2014, (b) 2014-2016, (c) 2016-2020, and (d) the entire period from 2003 to 2020. The 2003-2020 panel was derived from 9 MMASTER DEMs. The dynamic balance line for this period is shown as a black line. Base image: Sentinel-2, 2022-08-12.

2.5 Discussion

The results strongly suggest that Fisher Glacier underwent two surges between 1948 and 2022 that were separated by a ~40-year quiescent phase. In the following section we consolidate the results that support this conclusion, analyse the glacier's dynamics to infer the possible mechanisms and controls on these surges, compare Fisher Glacier's surges with others in the region, and finally put Fisher Glacier's surges in a broader global context. The most recent surge is discussed first as it is better constrained than the older one and can therefore provide insight into the changes that likely occurred in the older surge when less imagery and data is available (assuming that the surge dynamics were similar in both instances).

2.5.1 Evidence for a recent surge from 2013-2016

Fisher Glacier underwent significant change over the period 2013-2016, during which its velocities dramatically increased, mass was redistributed downglacier, the terminus underwent a major advance, and the ice surface displayed a number of changes including crevassing, serac development, and formation of new looped moraines.

Velocity changes were the first sign of change on Fisher Glacier, appearing much earlier than the other tracked changes. The ice velocity displayed a very long and gradual increase through the 1980s to 2000s, during which it increased both up and down glacier from the central part of the trunk, originating at a location around 24 km from the terminus. Although we refer to a velocity increase during this period, it should be emphasised that it was very gradual ($\sim 2 \text{ m yr}^{-1}$) and that the velocities remained very low ($< 50 \text{ m yr}^{-1}$) compared to peak values ($> 2000 \text{ m yr}^{-1}$), so we consider the glacier to be in its quiescent phase during this time. This is supported by the disappearance of the crevasses that were observed in 1972 and the development of potholes along the glacier's surface. As previously mentioned, crevasses often develop during a surge (Meier & Post, 1969), whereas potholes are frequently found on the glacier surface during quiescence (Sturm, 1987). Additionally, our results show that the terminus underwent sustained retreat averaging $-2058 \pm 8 \text{ m}$ ($-49 \pm 0.9 \text{ m yr}^{-1}$) across the terminus during this quiescent phase. Long periods of terminus retreat during quiescence are a common characteristic of many surge-type glaciers as the slow flow is insufficient to replace the negative mass balance of the ice at the terminus.

It was only from about 2008 onwards that the annual velocity began increasing at an enhanced rate (i.e., an annual rise of more than 2-3 m yr⁻¹). 2007/08 marked the start of the pre-surge buildup when a very localised region at the center of the glacier first began flowing at what we consider to be a moderate speed (mean centerline velocity of >70 m yr⁻¹). From 2008-2013 or so, most potholes that remained on the glacier surface drained or disappeared. The terminus, however, still showed consistent retreat with no sign of slowing. The slow velocity increase was mostly unnoticeable on the glacier surface as it was still smooth and unbroken. Additionally, the reservoir zone was still gaining elevation (mean of 33 ± 4 m from 2003-2014), while the receiving zone was losing elevation (mean of -26 ± 4 m from 2003-2014), typical behaviour for surge-type glaciers in their quiescent phase (Meier & Post, 1969).

In 2010, the moderate flow speeds of the pre-surge buildup phase finally spread to a greater portion of the glacier, increasing up- and downglacier from where they began in the central region. Previously published regional winter velocity maps covering the pre-surge buildup phase agree well with our results. Burgess et al.'s (2013) velocity map shows winter velocities in the 0.1-0.3 m d⁻¹ range (36.5-109.5 m yr⁻¹) from 2007-2011, which is in line with the increase that we report to have begun in ~2007. Van Wychen et al. (2018) report an increase in mean winter centerline velocities between winters 2007-08 and 2009-10, from ~40 m yr⁻¹ to ~60 m yr⁻¹ (± 8.5 -15.0 m yr⁻¹ uncertainty range), with highest values near the 20-25 km mark and gradual decreases both up- and downglacier from there, in line with our findings. Waechter et al.'s (2015) data continues this trend (i.e., increase in overall mean velocity, with the highest being reported near the mid-glacier), with mean centerline winter velocities of $\sim 73 \pm 16.6$ m yr⁻¹ in 2011 and 94 ± 12.7 m yr⁻¹ in 2012.

We consider the active phase of the surge to have begun in 2013 as the mean centerline velocities were >100 m yr⁻¹ with local flow speeds in excess of 200 m yr⁻¹ in the central region. However, the glacier surface remained visibly unchanged through 2013 and early 2014. The first visible signs of surging only became noticeable in late 2014, when velocities were rising at a faster rate (but not yet near their peak), and the terminus began advancing very slowly as the surge front seems to have finally reached it. The first significant changes in terminus length and area came in the early summer of 2015, and approximately coincided with the time when the rapid velocities reached the terminus area. Spring/early summer 2015 is also the time when the glacier surface became heavily crevassed, as shown in the *in situ* photographs from August 2015. In addition to this, the terminus began advancing at a faster rate as mass was being transferred to the lower

ablation area (the ice was higher along the valley walls in the lower ablation area in summer 2015 than in 2014).

All of these changes since mid-2014 align well with what is expected during a surge, with downglacier mass transfer causing extensional conditions in the middle and upper part of the glacier trunk, and compressive stress in the lower ablation area as the valley constricts and the slope becomes steeper. This is supported by the fact that from the head of the glacier at 45.5 km from the terminus, all the way down to at least 16 km from the terminus, the newly formed crevasses in 2015 were oriented in a transverse direction and relatively thin and closely spaced, indicative of accelerating ice and extending flow (Figure 28) (Hambrey & Alean, 2004; Colgan et al., 2016). Below the 16 km mark, the predominance of new crevasses near the glacier margins was likely caused by compressive strain exerted on the ice by the downglacier mass transfer (Figure 28), with the pattern of mass redistribution (i.e., the mean DBL for the entire 2003-2020 period) supporting this. Longitudinal crevasses (parallel to glacier flow) were observed from ~4 km down to the terminus where they splay outwards, suggesting that the ice is being thrust upwards at the terminus (Colgan et al., 2016). Seracs were also observed along the glacier margins and across the terminus in the lower ablation area in regions where the ice became convex, which is a clear sign of active surging (Meier & Post, 1969; Hambrey & Alean, 2004). The advancing ice and downglacier transfer of mass also created an ice cliff at the edge of the terminus, where the ice previously gradually entered the water without a pronounced cliff.

The spring of 2015 also saw the appearance of a number of supraglacial and ice-marginal lakes which subsequently drained through the summer at the same general time that velocities increased in the lower glacier. Although the spring season's enhanced melt rates regularly result in an increased production of supraglacial lakes even during quiescence, the timing of lake drainage and velocity increase during the surge suggests the possibility that the drainage of these supraglacial water features provided water to the glacier bed which increased basal sliding and the observed velocities.

The velocities and terminus advance both peaked during winter or early spring of 2016, ~2 years after the active phase initiated. Velocities peaked around 8-12 km from the terminus at $>5.5 \text{ m d}^{-1}$ and remained very high from early February to early June 2016. Correspondingly, the terminus underwent its greatest change during this period, with its area increasing at a rate of nearly $6 \text{ km}^2 \text{ yr}^{-1}$ and a terminus-wide rate of advance of 977 m yr^{-1} (2059 m yr^{-1} at the NT). Also,

the crevasses kept growing and spreading across the glacier's surface and the greatest increases in ice surface height along the margins occurred during this time. The location where the velocities peaked is also the point where the valley narrows, and the surface slope increases (see the hypsometric band area <1000 m a.s.l. in Figure 25), suggesting that geometric properties could play a role in increasing basal sliding rates. Our results are consistent with those of Altena et al. (2019), who described an increase in speed on Fisher Glacier from 2013-2018, and a down- and upglacier propagation from the same surge initiation point that we identified at ~22 km, although their peak reported velocities of $\sim 4 \text{ m d}^{-1}$ were lower than ours, likely because their results were more generalized at the glacier-wide scale.

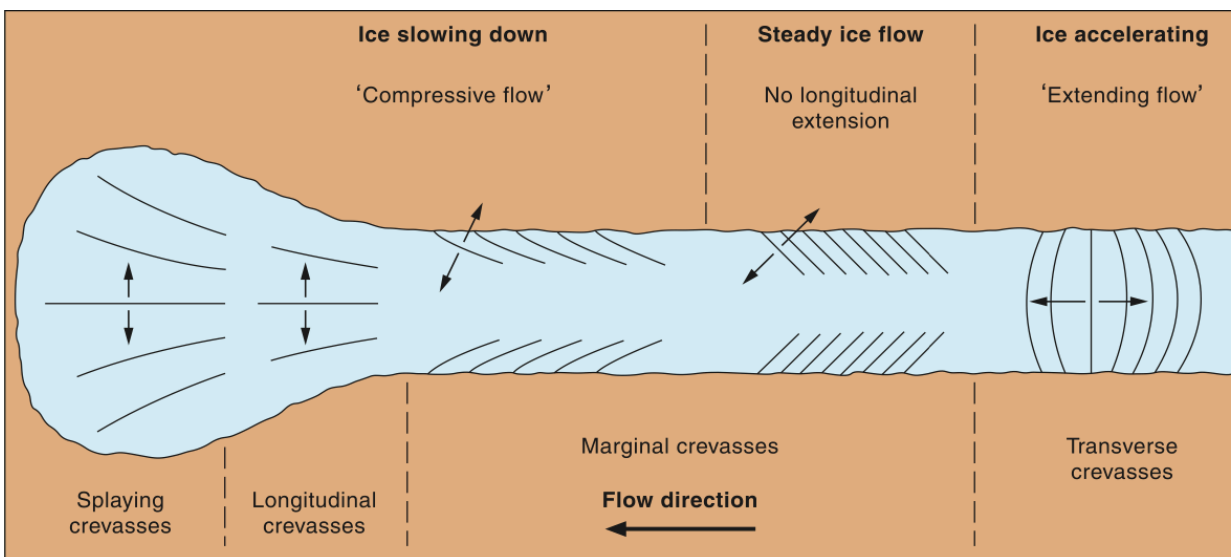


Figure 28. Plan view of the main types of crevasses found on valley glaciers. Arrows show the direction in which the ice is pulled apart. The type of flow involved is also indicated for each type of crevasse. Reproduced from Hambrey & Alean (2004, p.83).

The velocities then dropped rapidly at the end of summer 2016 within the space of only 1-2 months. The slowdown began at the same location where the surge had initiated (~20-25 km from the terminus) and, as with the initiation, it also spread up- and downglacier. SAR data from winter 2017 (i.e., the first winter after surge termination) shows velocities of $\sim 25 \text{ m yr}^{-1}$ throughout most of the glacier, except for the lower 5 km and upper 10 km of the glacier where velocities were still at $\sim 50 \text{ m yr}^{-1}$. This suggests that Fisher Glacier had reverted to its quiescent phase (i.e., very low velocities that are in the range observed during previous quiescent phases), but that the slowdown was still spreading from the center towards its extremities over the winter of 2016/17.

After the termination phase, the following years were characterized by slow velocities ($<50 \text{ m yr}^{-1}$), surface lowering, and terminus retreat. The crevasses took a few years to close completely but had mostly disappeared by 2021 and potholes had reappeared at the same location as before, in the pothole fields above the $\sim 15 \text{ km}$ mark, which supports the glacier's return to quiescence. Sturm (1987) postulates that potholes form from closed-up crevasses that fill with water. While that may certainly be true (the areas where potholes formed on Fisher Glacier were also crevassed during the surge), it does not explain the reappearance of potholes in certain specific locations but not in other locations that were also crevassed. Sturm (1987) mentions that potholes are generally found on or near tributary glaciers. This observation is corroborated by our observations, but we are unable to offer any possible hypotheses explaining this.

In summary, the sequence of events surrounding the recent surge can be summarized as: (i) quiescence from the mid-1970s to 2012, including a pre-surge buildup phase from 2010-2012 (with some ambiguity as to whether the pre-surge buildup began in 2007/8), (ii) active phase 2013-2016 (with 4-month peak from early February to early June 2016), (iii) quiescence 2017-present (2022). This amounts to a relatively long quiescent phase (compared to other glaciers in the region), a 3-5-year buildup, and a 3-year active phase. Note that we specify the start of the active phase based on a somewhat subjective criteria of when ice velocities became “significant” ($>70 \text{ m yr}^{-1}$ centerline average, as opposed to $\sim 45 \text{ m yr}^{-1}$ during quiescence, with local velocities at the surge nucleus being $\sim 200 \text{ m yr}^{-1}$) and the rate of velocity increase climbing significantly.

2.5.2 Evidence for a historic surge ~ 1970

The historic air photos and satellite imagery provide evidence for a surge that occurred shortly before 1972. This finding is supported by the observation of Clarke & Holdsworth (2002, page J312) that “*Fisher Glacier [...] surged around 1970, but little information about this event is available.*” The RCAF images from 1947, 1948, and 1954 all show a very smooth surface that is devoid of crevasses, and an ice margin that is in a state of downwasting. The 1954 orthophoto, whose coverage extends further upglacier than the 1947 and 1948 photos, also shows many large potholes covering the glacier surface above the 18-22 km mark, which typically form during the quiescent phase of the surge cycle (Sturm, 1987; Yde & Knudsen, 2005; Paul, 2015). Additionally, the terminus length and area data show that the terminus had retreated and covered a smaller area between 1948, 1954 and 1963. Furthermore, an oblique photograph taken by Austin Post in 1960

shows the ice extent at the southern terminus (Figure 29a), providing validation for the mapping in the low resolution 1963 KH-5 image (Figure 29b). The Austin Post photo also confirms that the ice surface was still smooth and free of crevasses and other such features in 1960. Thus, there is evidence that the glacier was in quiescence from <1947 to at least 1963.

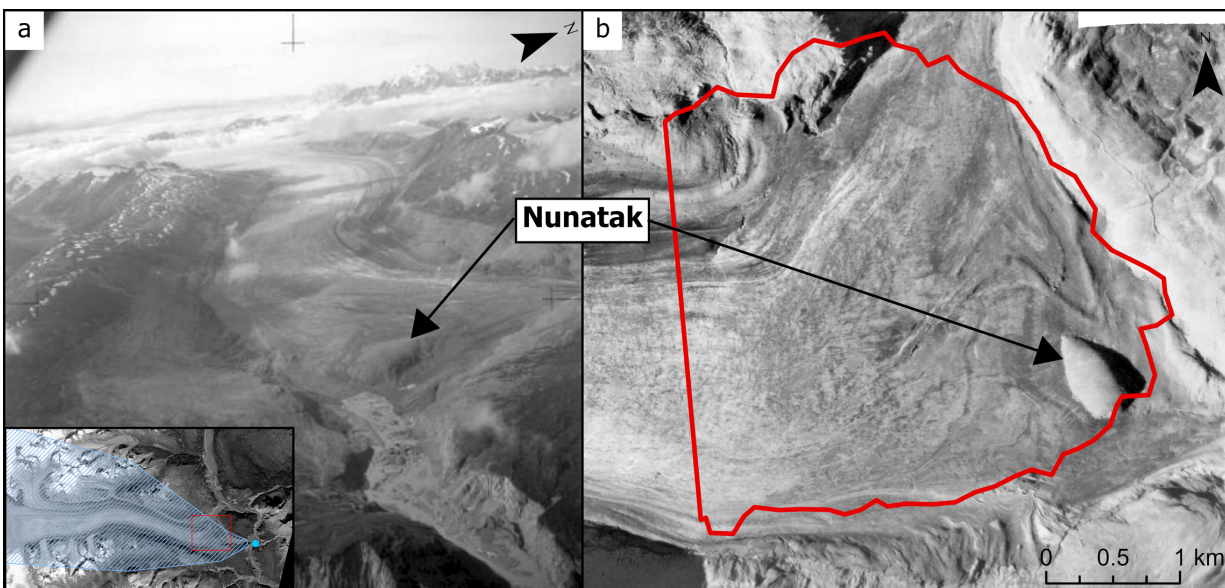


Figure 29. (a) Oblique photo of Fisher Glacier taken by Austin Post from an airplane flying down the Alsek River on 1960-08-26. The image looks West, upglacier from the southern terminus. The photo was obtained from the University of Alaska Fairbanks Geophysical Institute. Inset map shows the approximate location where the photo was taken (blue dot) and the view (blue hashmarks). (b) 1963 terminus outline traced from the KH-5 satellite image with ~140 m resolution. The nunatak is identified in both images for reference. The 1963 terminus outline is displayed on the 1954 RCAF image (A14459) instead of the 1963 image because the nunatak was not clearly distinguishable in the KH-5 image due to poor resolution.

In 1972, the air photo shows a surface that is crevassed from the terminus all the way up to >30 km upglacier, and a terminus that was 0.57 ± 0.14 km more advanced than in 1963. The nunatak that was visible at the terminus in the previous images (e.g., Figure 29b) has been completely covered by ice, indicating a significant increase in volume at the terminus. The glacier surface in 1972 greatly resembles that of 2016 in terms of crevasse distribution and extent, and like 2016, had almost no potholes and a small supraglacial lake that formed at the snout of Tributary 1. Much like 2016, the crevasses at the terminus were chaotic, being oriented both in transverse and longitudinal directions, splaying near the ST, and forming serac fields. The

appearance of a new looped moraine ~22 km upglacier also supports the occurrence of a surge during this time.

Based on the state of the glacier in 1972, and how it compares to the peak of the recent surge, it seems that a surge was already well under way in 1972. The point velocities support this and offer some insight into the possible end date of the surge. Velocities along the centerline between 1954 and 1972 show a mean of 233 m yr^{-1} , which is much higher than the quiescent velocities of $<50 \text{ m yr}^{-1}$ measured in the 1980s and 1990s. However, we know that the glacier was still in quiescence in 1963, meaning that velocities were likely much slower than reported from at least 1954-1963. If the historical surge dynamics were similar to those of the recent surge, the active phase likely only lasted ~3 years with perhaps a 5-year pre-surge buildup. This means that the velocities would have started increasing in the mid- to late-1960s, and then rose to much higher values around 1969 when the glacier entered the active phase of the surge. This sequence of events would corroborate the report by Clarke & Holdsworth (2002), as the surge would have been in full swing when it was observed in 1970. The changes in surface elevation also support this, with the difference DEM from 1954-1972 showing thinning in the reservoir zone ($>1100 \text{ m a.s.l.}$, or the 21 km mark) and thickening in the receiving zone (note that the location of the DBL during the historic surge was located around the same location as the surge nucleus in the recent surge and at approximately the same location as the 2003-2020 DBL).

Our results also give clues as to the possible end date of the historical surge. To start, the point velocities from surface feature tracking show a mean velocity of 22 m yr^{-1} from 1972 to 1976, in line with velocities observed from 2017-onwards during the return to quiescence after the recent surge, suggesting that the surge was near to its termination in 1972. Second, in the 1976 and 1977 imagery, the terminus had already begun retreating from its 1972 extent, the crevasses were mainly closed, and small potholes had started to reappear. In the recent surge, crevasses took several years to close after velocities reduced and potholes only reappeared in 2021, about 4 years after surge termination, providing further evidence that the 1972 image shows the glacier close to termination. Third, the pattern of elevation changes across Fisher Glacier's surface from 1972-1976 (i.e., widespread thinning) is very similar to the pattern that was observed from 2016-2020 during the most recent switch back to quiescence, which highlights the similarities between both surges and further supports a termination date around 1972.

To summarize, based on a combination of historical data and inferences made based on similarities with the recent surge, our reconstruction of the sequence of events surrounding the historical surge is: (i) quiescent phase from at least 1947 to ~1963 with a pre-surge buildup ~1964 to 1968, (ii) active phase from ~1969 to ~1972, and (iii) quiescence from 1972/73-2012. The time between the historical surge's termination and the recent surge's initiation suggests a quiescent phase duration of ~40 years.

2.5.3 Surge mechanisms of Fisher Glacier

Given the characteristics and timing of the two surges described for Fisher Glacier, we can make inferences concerning the likely mechanism(s) that caused them.

Some of the main characteristics typically seen in hydrologically controlled surges are short active phases lasting 1-3 years followed by short quiescent phases of about 10-40 years (Meier & Post, 1969; Kamb et al., 1985; Lawson, 1997), rapid surge initiation in the late autumn or winter (Kamb et al., 1985; Lawson, 1997), and an abrupt surge termination in the late summer (e.g., Kamb et al., 1985; Lawson, 1997; Murray et al., 2003; Bevington & Copland, 2014; Abe et al., 2016). Conversely, thermally controlled surges typically have longer active phases of 3-10+ years, longer quiescent phases of 50+ years, a gradual surge initiation and/or termination, and the surge initiation and termination timing is variable (i.e., no defined season) (Copland et al., 2003; Murray et al. 2003; Quincey et al., 2011).

The hydrologic mechanism that is responsible for the surge initiation of most glaciers in Yukon/Alaska could explain the velocity increases that occurred in winters 2015, and 2016. Since there is less water reaching the bed in the winter months, any subglacial drainage conduits that were formed during the summer (when there is more water flowing through the glacier) get overpowered by the ice overburden pressure, which collapses them and disperses the water into an inefficient/dispersed drainage system. This inefficient drainage system is highly pressurized and leads to increased basal sliding and, ultimately, increased velocities (Kamb et al., 1985). During Fisher Glacier's recent surge, the three winters from 2014-2016 were the periods with the most activity (greatest velocity changes, most significant terminus advances, appearance of crevasses), which is suggestive of a hydrological control on the surging. Furthermore, the appearance of short-lived supraglacial lakes in winter/spring that then drain through the summer provides further evidence that water availability affects surge dynamics. Pressurized water at the glacier bed during

the winter can exploit the fractures (crevasses) at the ice surface and create drainage pathways which facilitate subglacial water movement to the surface (Bennett et al., 2000), where it pools and creates the lakes that are observed each spring during the recent surge. Basal water pressures then lower throughout the summer and the water in the lakes is able to drain englacially. Fisher Glacier's rapid termination at the end of the summer also aligns with the hydrologic control, since increased water input from summer melt can create large drainage conduits, which decreases sliding speeds (Kamb et al., 1985). The opening of such conduits may be responsible for the rapid drainage of the large supraglacial lake/stream feature presented in Figure 22b-d, which occurred synchronously with the surge termination. If the two events are related, the drainage of this surface feature could be a visual representation of the switch to an efficient drainage system and would be evidence for a hydrologically triggered termination. Alternatively, pools of surface meltwater can also develop on the ice surface since surging causes normal drainage routes to close up as crevasse and moulin distribution changes.

Although the active phase length and the timing of surge initiation and termination show evidence of a hydrologically controlled surge, Fisher Glacier also displays characteristics that are more typical of thermally controlled surges. For one, its long quiescence phase of >40 years is reminiscent of a few surge-type glaciers in the Canadian High Arctic (although it is still within the upper limit of 'Alaskan-type' surges described by Meier & Post (1969)). Second, the pre-surge velocity buildup is a trait that is more typical of thermally controlled surges (Murray et al., 2003). Third, its active phase velocities are relatively low compared to most glaciers in Yukon/Alaska, but are similar to some glaciers in the Canadian High Arctic, Svalbard, and the Karakoram, which are suspected to be thermally controlled (discussed in detail in the following sections). It therefore seems that Fisher Glacier's surges may be the result of an interaction between hydrological and thermal controls, in which case they can be characterised by a hybrid mechanism.

Further research on the central region of the glacier at around 22 km upglacier could elucidate the mechanisms at play. This region seems to play a key role in Fisher Glacier's surges. It is the location where the recent surge initiated and terminated, and it is from here that the multidecadal velocity increase originated from. During the recent surge, this location is where the largest drainage feature formed (as seen in Figure 22b-d), and is also located just downglacier from Tributary 1 (i.e., the main tributary that seems to speed up and contribute to the downglacier mass transfer during a surge). A visual assessment of the height of the ice along the glacier margin

during the recent surge shows a clear thickening below this key location, but the increase is less obvious above it. It is also approximately the location of the DBL from 1954-1972. Additionally, during quiescence, the area of dense pothole fields is located just below this point. Although this section of the glacier is likely important, there is unfortunately no existing data that could give insights into the bed topography or ice thickness at this location. Therefore, this is a recommended area of investigation for future studies to validate whether the bed topography at this location acts as a pinning point, reverse slope (e.g., Flowers et al., 2011), or some sort of other control on Fisher Glacier's surges.

2.5.4 Implications for Fisher Glacier's future surges

Although the two surges broadly resembled each other in terms of general patterns, the maximum terminus extent reached in the ~1972 surge ($15.86 \pm 0.06 \text{ km}^2$) was greater than in 2016 ($7.88 \pm 0.04 \text{ km}^2$; both areas measured in relation to a common cross section near the glacier's terminus, as seen in Figure 3b), and the maximum height of the ice along the valley walls was much higher in the earlier surge. The northern terminus in the historical surge reached a maximum extent only 0.9 km away from the western bank of the Alsek River, whereas the recent surge's maximum extent was ~2.8 km away from reaching the Alsek River. Although the historical surge came relatively close to reaching the river at that time, the chances of Fisher Glacier damming it in the future are considered unlikely due to a few reasons. First, the 1972 extent was the furthest in documented history and under current and projected future climate conditions, Fisher Glacier is expected to maintain a negative mass balance, as it has for the past >40 years (Arendt et al., 2008; Berthier et al., 2010; Doumbia et al., 2020; Liu et al., 2022). Accordingly, we expect Fisher Glacier's future surges to reach successively smaller extents, as has happened at nearby Lowell Glacier (Bevington & Copland, 2014) and Donjek Glacier (Kochtitzky et al., 2019). Furthermore, if the ~40-year long quiescent phase is typical surge behaviour for this glacier the next surge is not expected to occur until ~2056, by which time the terminus will be even further retreated and the glacier will have lost more mass. Previous research has shown that surge-type glaciers which have undergone sustained negative mass balance can enter a climate-induced senescence (Dowdeswell et al., 1995; Striberger et al., 2011; Benn et al., 2019a). Topographic restrictions have also limited the change in extent of the central and southern parts of the terminus during the past two surges, meaning that the northern section would be the only one with potential for blocking the Alsek

River, but this is extremely unlikely under current trends due to the mass balance implications discussed above.

2.5.5 Comparison with other surge-type glaciers in the St. Elias Mountains

Yukon/Alaska as a whole is home to 322 surge-type glaciers (Sevestre & Benn, 2015), 151 of which are in the St. Elias Mountains (Clarke et al., 1986). As such, the region has been characterized as a surge supercluster with only Svalbard having more confirmed surge-type glaciers (Sevestre & Benn, 2015). The oldest reports of surging glaciers in Yukon/Alaska date from well before 1890 as they have been passed down through indigenous oral traditions for generations (Cruikshank, 1981). One such example is Lowell Glacier, immediately north of Fisher Glacier, whose Athapaskan name is Nàhùdäy, meaning “fish stop” due to its surges blocking fish migrations up the Alsek River (Cruikshank, 1981).

This section compares Fisher Glacier’s surge behaviour to that of other glaciers in Yukon/Alaska, particularly in the St. Elias Mountains. Table 8 shows a summary of the main characteristics for each glacier that is discussed. Consult Figure 1 to see the location of these glaciers.

Fisher Glacier displays local velocities up to 5.5 m d^{-1} at its peak, although average surge velocities over most of the glacier reached approximately 4 m d^{-1} . These velocities are relatively low for most hydrologically controlled surges in the Yukon (Table 8; Kochtitzky et al., 2019). For example, Lowell (Bevington & Copland, 2014), Variegated (Kamb et al., 1985; Lawson, 1997), and Tweedsmuir (Krimmel & Meier, 1975; Sharp, 2021) glaciers all reached peak velocities of $10+ \text{ m d}^{-1}$ during their surges, about double that recorded on Fisher Glacier. Even higher peak velocities have been recorded on Walsh (14 m d^{-1} ; Fu & Zhou, 2020; Meier & Post, 1969), Little Kluane (13 m d^{-1} ; Brittany Main, personal communication, 2023), and Turner (up to 25 m d^{-1} ; Nolan et al., 2021) glaciers. However, some large surge-type glaciers in the St. Elias Mountains do display similar speeds to Fisher Glacier, such as Klutlan Glacier which experienced velocities of $4+ \text{ m d}^{-1}$ in its 2014-2018 surge (Altena et al., 2019; Samsonov et al., 2021; Meier & Post, 1969), and Donjek Glacier which flows at $2.0\text{-}4.5 \text{ m d}^{-1}$, depending on the surge (Abe et al., 2016; Kochtitzky et al., 2019). Conversely, the few thermally controlled glacier surges in the St. Elias Mountains display much slower peak surge velocities of 0.12 m d^{-1} (Trapridge Glacier; Frappé & Clarke, 2007) and $0.05\text{-}0.10 \text{ m d}^{-1}$ (North Glacier; De Paoli & Flowers, 2009).

Quiescent phase velocities in the aforementioned glaciers are all $<0.6 \text{ m d}^{-1}$, except for Tweedsmuir Glacier which has a reported mean quiescent velocity of $\sim 1 \text{ m d}^{-1}$ (Clarke & Holdsworth, 2002). Considering the annual multidecadal speedup during quiescence, Fisher Glacier's quiescent velocities ranged from $0.09\text{-}0.4 \text{ m d}^{-1}$, comparable to most other surge-type glaciers in the area.

Fisher Glacier's 3-year active phase duration is comparable to that of the other glaciers in the St. Elias Mountains. Variegated, Lowell, Donjek, Tweedsmuir, Turner, and Little Kluane glaciers all have active phases that typically last less than 3 years (Meier & Post, 1969; Krimmel & Meier, 1975; Kamb et al., 1985; Lawson, 1997; Bevington & Copland, 2014; Abe et al., 2016; Kochtitzky et al., 2019; Nolan et al., 2021; Samsonov et al., 2021; Brittany Main, personal communication, 2023), whereas Walsh and Klutlan glaciers both exhibit active phases lasting up to 4 years (Fu & Zhou, 2020; Samsonov et al., 2021). Fisher Glacier's active phase duration is therefore in the middle of the typical range and is within the range initially described by Meier & Post (1969).

Fisher Glacier's surges initiate and terminate from the central section of the trunk, and the surge propagates both up- and downglacier, although the upglacier progression is a little more subdued (i.e., lower peak velocities than the downglacier progression). This is uncommon among glaciers in the St. Elias Mountains. Most glaciers have surges that initiate in the mid- to upper-glacier and progress downglacier from the initiation point. Notably, Walsh, Variegated, and Little Kluane glaciers all initiate in their upper region and propagate downglacier from the initiation location (Lawson, 1997; Samsonov et al., 2021; Brittany Main, personal communication, 2023). Like Fisher Glacier, Klutlan, Donjek, and Turner glaciers also initiate in the mid-trunk. However, Donjek Glacier's surges only affect the lower 20 km of the glacier as the surge front only progresses downglacier from the initiation point at 18-22 km from the terminus (Abe et al., 2016; Kochtitzky et al., 2019). Likewise, Turner Glacier has a similar downglacier surge progression from its initiation point in the mid-trunk (Nolan et al., 2021). Klutlan Glacier's surges are similar to Fisher Glacier's because it is one of the rare cases where the surge spreads both up- and downglacier, although its 2014 surge primarily affected the area downstream from the initiation point (Altena et al., 2019; Samsonov et al., 2021) and the upglacier progression of Klutlan Glacier's surge was limited in extent to its westernmost basin (Altena et al., 2019).

Table 8. Comparison of some of the main surge features on glaciers in the St. Elias Mountains based on the published literature. Glacier locations provided in Figure 1.

Glacier	Quiescence velocity (m d ⁻¹)	Peak surge velocity (m d ⁻¹)	Active phase duration (years)	Recurrence interval (years)	Surge nucleus	Surge initiation	Surge termination	Sources
Fisher	0.1-0.4	5.5	3	~40	Middle	Winter	Summer	-
Variegated	0.1-0.5	10 (1982)	1-2	12-26	Upper	Winter	Summer	a, b
Walsh	0.4	14	1.5-4	30-60	Upper	Winter/Spring	Summer	c, d
Klutlan		4+	3-4	30-40	Middle	Possibly winter	Summer	c, e, f
Lowell	0.1-0.5	10	1-2	12-20	Maybe terminus	Late summer/early fall	Summer	g
Donjek	0.5	2-4.5	1-2.15	12	Middle	Summer (mostly)		h, i
Tweedsmuir	1	10+	1	30+	Middle-Upper	Late summer/winter	Late-summer/early winter	j, k
Turner	< 1 (~0.6)	25	2	5	Middle	Winter/early spring		l
Little Kluane		13.7	~1.5		Upper	Summer	Summer	f, m
Trapridge	0.025	0.12	~20	~30				n
North Glacier		0.05-0.10		~30-40	Middle			o

a Kamb et al. (1985), b Lawson (1997), c Meier & Post (1969), d Fu & Zhou (2020), e Altna et al. (2019), f Samsonov et al. (2021), g Bevington & Copland (2014), h Abe et al. (2016), i Kochitzky et al. (2019), j Krimmel & Meier (1975), k Sharp (2021), l Nolan et al. (2021), m Main (pers. comm., 2023), n Frappé & Clarke (2007), o De Paoli & Flowers (2009)

The hydrologic surge mechanism suggests that surges are likely to initiate in late fall or winter and terminate in late summer when melt rates are at their highest (Kamb et al., 1985). In addition to Fisher Glacier, Variegated, Walsh, Turner, possibly Klutlan, and some of Tweedsmuir's past surges also initiated in the winter or early spring (Kamb et al., 1985; Lawson et al., 1997; Altena et al., 2019; Nolan et al., 2021; Samsonov et al., 2021; Sharp, 2021). Conversely, Lowell Glacier's surges typically initiate in late summer or early fall (Bevington & Copland, 2014), while Donjek is alone with a summer initiation time (Abe et al., 2016; Kochtitzky et al., 2019). Surges in the St. Elias Mountains show a clear pattern of active phase termination during the summer (Table 8).

This regional comparison of surge characteristics suggests that Fisher Glacier generally behaves as would be expected from a temperate glacier controlled by a hydrological switch mechanism, but that it also shows some characteristics which are not easily explained by the hydrological mechanism. Characteristics which suggest a hydrological control include its short active phase that initiates in the winter and terminates in late summer, and that the termination occurs more rapidly than the surge initiation. It also has many surface drainage features, which is suggestive of high amounts of water transport in the glacier system. Furthermore, its quiescence speeds are relatively rapid. Conversely, its relatively long quiescent phase, and decades-long velocity increase followed by a pre-surge buildup phase are unusual for surge-type glaciers in the Yukon but resemble some characteristics of polythermal glacier surges, as seen in the Canadian High Arctic and Svalbard (Murray et al., 2003), as discussed below. A particularly unusual characteristic is the propagation of the surge front both upglacier and downglacier from the initiation point.

2.5.6 Comparison with surge-type glaciers globally

The following section will compare Fisher Glacier's surges to surging in other regions around the globe. Table 9 shows some surge characteristics of glaciers in various regions.

In eastern Greenland, Sortebræ has been shown to have high velocities (peak $\sim 24 \text{ m d}^{-1}$) and a short active phase (~ 2.5 years), much like glaciers in Yukon/Alaska (Pritchard et al., 2005). Its active phase length is therefore comparable to Fisher Glacier, although its peak velocities are much higher. Storstrømmen and L. Bistrup Bræ on the other hand, are more similar to 'Svalbard-

type' surges with long active phases, long quiescent phases and slow initiation and termination phases (Mouginot et al., 2018). Both of these glaciers have active phases lasting 10 years while Storstrømmen has a quiescent phase of 70 years and L. Bistrup Bræ's is 30-50 years (Mouginot et al., 2018).

In Svalbard, Monacobreen displays similar peak velocities to Fisher Glacier (4.5 m d^{-1} and $4.5\text{-}5.5 \text{ m d}^{-1}$, respectively), but has a long active phase length of 6 years and a shorter quiescent phase than Fisher (Murray et al., 2003; Mansell et al., 2012; Banerjee et al., 2022). Morsnevbreen is another glacier which contrasts significantly with Fisher Glacier as it displays very high peak flow velocities of 15 m d^{-1} and a very long active phase (9 years) and quiescent phase (75 years) (Benn et al., 2019b). These studies show that while peak surge velocities in Svalbard can be quite elevated, they typically have longer initiation and termination phases than Fisher Glacier and, in particular, differ in the amount of time it takes for the velocities to transition back to quiescence (a gradual process that can take years for Svalbard glaciers as opposed to just months for Fisher Glacier).

In the Canadian High Arctic, glaciers typically have long active and quiescent phases, similar to Svalbard (Copland et al., 2003). For example, Iceberg Glacier has a 22-year active phase with peak velocities of 6.3 m d^{-1} followed by a long slowdown phase reaching very low velocities of 0.032 m d^{-1} (Lauzon et al., 2023). Recent studies have shown that some glaciers in the CAA that were previously thought to be surge-type may in fact have dynamic instabilities caused by other factors unrelated to surging (Van Wychen et al., 2016; Medrzycka et al., 2019; Lauzon, 2022). For example, several glaciers (e.g., Tuborg, Cañon, and Parrish glaciers) that had been classified as surge-type by Copland et al. (2003) were recently reclassified as pulse-type as they do not display sufficient evidence of surging (Van Wychen et al. 2016). Still others, notably Good Friday Glacier and Airdrop Glacier, have displayed multi-decadal periods of high velocities and terminus advance which may be due to a delayed response to positive mass balance conditions during the Little Ice Age, rather than a classic surge (Medrzycka et al., 2019; Lauzon, 2022, respectively). Slow surges have also been observed in the CAA, such as at Split Lake Glacier which has been surging for 50+ years with lower velocities than what is typically considered to be a surge (i.e., velocities of less than one order of magnitude above quiescence speeds), but with most other typical surge characteristics (i.e., terminus advance, downglacier mass redistribution, and velocity changes out of line with surface mass balance) (Van Wychen et al., 2022). These

examples of long-term dynamic instabilities are therefore very different from what is observed on Fisher Glacier, and highlight the vast spectrum of behaviours that surge-type glaciers have in the Canadian High Arctic.

In the Karakoram, surge-type glaciers seem to be predominantly controlled by a thermal mechanism (especially the timing of surge initiation and termination), but also display characteristics typical of the hydrological control (e.g., short active phases, rapid buildup and slowdown, etc.) and display a lot of inter-glacier variability (Quincey et al., 2011; 2015). This characterisation is therefore similar to our assessment of Fisher Glacier's surges. Peak velocities measured on Kunyang Glacier are similar to those on Fisher Glacier (Quincey et al., 2011; 2015). Additionally, both of these glaciers' surges resulted in a terminus advance (which is not always the case in the Karakoram), and both terminated during the summer. However, unlike Fisher Glacier, Kunyang Glacier's surge initiated in the late summer (Quincey et al., 2011), which is inconsistent with hydrologically controlled surges, thus suggesting a predominant thermal control. Also similar to Fisher Glacier, Khurdopin Glacier has a relatively short active phase of 3-5 years, similar velocities, possibly a winter initiation and a summer termination, and its last surge had no pronounced surge front (like what we observed on Fisher Glacier). However, Khurdopin Glacier's surge differs from Fisher because it did not result in a terminus advance, and it has a quiescence period length about half that of Fisher Glacier.

In sum, we can draw many similarities between Fisher Glacier's surges and other surges observed around the world. However, comparing surge dynamics between glaciers on the basis of either the thermal or hydrological theories proves to be difficult because many glaciers show characteristics of both. Instead of using this binary classification, surging might be better explained by some other underlying processes such as basal enthalpy production (Benn et al., 2019a). Benn et al. (2023) demonstrate that this theory can explain surges observed on both polythermal and temperate glaciers and that both 'Alaskan-type' and 'Svalbard-type' surges occur due to the same underlying processes. A unifying theory such as this is likely more appropriate for assessing surging as it allows for more variability in observed dynamics and does not limit the glaciers to a binary classification. Finally, this comparison also showed that Fisher Glacier is one of the few glaciers with a surge that propagates both up- and down-glacier from a central location.

Table 9. Global-scale comparison of some of the principal surge characteristics on glaciers in the main surging hotspots, based on the published literature.

Glacier	Quiescence velocity (m d ⁻¹)	Peak surge velocity (m d ⁻¹)	Active phase duration (years)	Recurrence interval (years)	Surge nucleus	Surge initiation	Surge termination	Sources
Fisher	0.1-0.4	5.5	3	~40	Middle	Winter	Summer	-
Sortebræ (Greenland)	0.9	24	~2.5		Middle	Fall/winter	Spring/summer	a
L. Bistrup Bræ (Greenland)	~0.27	3.7	10	30-50				b
Storstrømmen (Greenland)	~0.55	8.2	10	70				b
Morsnevbreen (Svalbard)	<0.25	15	9	74	Middle			c
Monacobreen (Svalbard)	0.6	4.5	6	20	Terminus	Late fall/winter	Winter (initial slowdown)	d, e, f
Split Lake Glacier (CAA)		1.5	50+					g
Iceberg Glacier (CAA)	<0.055				Terminus			h, i
Mittie Glacier (CAA)		2.7	12					h, j, k,
Otto Glacier (CAA)	0.55-0.83	~2.2-7.7	14 (recent surge)	~30-35	Terminus			g, h, l
Middle Glacier (CAA)		>0.45	>10					h, k
Chapman Glacier (CAA)	0.014-0.14	0.55		Shorter than usual for CAA (i.e., not decades long)	Upper			h
Khurdopin Glacier (Karakoram)	0.027 (1999)-0.14 (1979)	2.74 -13.70	3-5	20	Middle	Possibly winter	Summer	m, n
North Gasherbrum Glacier (Karakoram)	0.13	1.34			Middle	Fall	Possibly summer	m
Kunyang Glacier (Karakoram)	0.055	5.48			Middle	Late summer	Summer	m

a Pritchard et al. (2005), b Mouginot et al. (2018), c Benn et al. (2019b), d Murray et al. (2003), e Mansell et al. (2012), f Banerjee et al. (2022), g Van Wychen et al. (2022), h Van Wychen et al. (2016), i Lauzon et al. (2023), j Copland et al. (2003), k Van Wychen et al. (2021), l Hattersley-Smith (1969), m Quincey et al. (2011), n Quincey & Luckman (2014)

2.6 Conclusions

This study demonstrates that Fisher Glacier has undergone two full surges over the past ~70 years. Its surge history can be broken into 5 distinct periods:

- (i) Quiescent phase 1 (1947-1968/9): 21+ year long period during which the terminus progressively retreated, and ice thinned in the lower receiving zone and thickened in the upper reservoir zone. The ice surface was characterized by potholes and there were few crevasses. Assuming similar surge dynamics to its recent history, we suggest that a pre-surge buildup phase would have occurred from ~1964-1968/9.
- (ii) Active phase 1 (historic surge; ~1969-1972): The terminus underwent a significant advance of several hundred meters and thickened by tens of meters as mass was transferred down-glacier, and the glacier surface became heavily crevassed, particularly along its margins and in the terminus region. Fisher Glacier reached its maximum documented extent of ~0.9 km from Alsek River during this surge.
- (iii) Quiescent phase 2 (1972/3-2012): 40-year period characterized by low velocities, but which increased gradually (by $\sim 2 \text{ m yr}^{-1}$) on an annual basis since at least the mid-1980s, spreading both up- and down-glacier from the central region of the trunk at ~22 km upglacier from the terminus. The terminus underwent sustained retreat during this time ($-2058 \pm 8 \text{ m}$), and the ice surface was generally smooth with only potholes and moulins visible in certain areas. The reservoir zone was subject to thickening, and the terminus thinned, both in the tens of meters. A pre-surge buildup began around 2008-10, during which velocities in the central region of the glacier increased at an accelerated rate for 3-5 years before reaching active phase velocities of $>200 \text{ m yr}^{-1}$.
- (iv) Active phase 2 (recent surge; 2013-2016): The velocities increased significantly (to $\sim 1500 \text{ m yr}^{-1}$) in the middle of the glacier and propagated both up- and down-glacier. Although there was no clearly defined initiation time or surge front, the greatest velocity increases and terminus advances occurred each winter during the three surge years. Velocities peaked at $>2100 \text{ m yr}^{-1}$ in winter 2016 at ~12 km from the terminus and dropped very rapidly along most of the glacier's length over a period of 1-2 months in late summer 2016. The surge termination occurred first in the middle part of the glacier, near to where it initiated, and spread upglacier and downglacier from there. During this active phase, the receiving zone thickened by a mean of $32 \pm 10 \text{ m}$ while the reservoir zone thinned by a mean of 16 ± 10

m, the terminus advanced by a terminus-wide mean of 868 ± 8 m, and crevasses developed across the glacier's surface.

- (v) Quiescent phase 3 (2017-Present (2022)): By early 2017, velocities returned to quiescent speeds (47 ± 28 m yr⁻¹) along the entire length of the glacier and the regular seasonal velocity pattern (i.e., mid-summer speedup) returned. Additionally, the crevasses and seracs closed up and potholes started developing in the pothole fields ~4 years after surge termination.

Fisher Glacier's surge dynamics seem to be generally hydrologically controlled, although some of its characteristics are better explained by the thermal control, suggesting that both processes may be interacting. The timing of surge velocity peaks in winter, and the summer termination, supports the hydrological control, as does the abrupt termination and the short active phase length. However, its relatively long quiescent phase of ~40 years is more typical of thermally-controlled ('Svalbard-type') surges, but still fits in the upper limit of the range provided by Meier & Post (1969) for hydrologically-controlled 'Alaskan-type' surging glaciers. On the other hand, the pre-surge buildup phase and the lack of a distinct surge bulge is reminiscent of thermally controlled surges. Its unique attribute is the double surge front which progresses both up- and down-glacier from the initiation point in the middle of the glacier. Finally, although not necessarily unusual for surge-type glaciers at higher latitudes, its pre-surge buildup phase is uncommon for surge-type glaciers in the St. Elias Mountains.

There remain several questions that would help elucidate the surge mechanisms and triggers of surging on Fisher Glacier. For one, obtaining bed topography and ice thickness measurements would help to determine the importance of the central part of the glacier, which is the key location where most of the surge activity occurs (velocity increase/slowdown, surface elevation changes, pothole field location, etc.). These measurements could potentially confirm the presence of a reverse bed slope, pinning point, or some other basal geometrical property of Fisher Glacier that could explain the observed patterns in that section of the trunk. Second, ice temperature measurements would contribute to understanding if/where the ice is frozen to the bed, which would provide valuable information on the possible controls on surging. Furthermore, modelling accumulation rates in the reservoir zone could provide insight into whether a critical accumulation threshold needs to be attained in order for a surge to initiate, and a more detailed

climate analysis of the region could yield information on water inputs to the glacier system around the surge initiation and termination dates (e.g., if it initiates after a period of significant rainfall). Finally, testing the applicability of Benn et al.'s (2019a) Enthalpy Balance Theory could provide additional insight into factors affecting its surges and might provide a better way to compare its surges to other glaciers.

To conclude, this study provides a reconstruction of Fisher Glacier's previous surges and a base from which we can carry out more detailed studies that focus on particular aspects of its surge dynamics. The present research also helps to define the variability of surge behaviours in Yukon/Alaska and on a global scale. It is highly likely that future surges of Fisher Glacier will not dam the Alsek River, although water inputs from the glacier could contribute to changing river water levels. Future work to determine this glacier's dynamics would benefit from more *in situ* measurements and ice penetrating radar data to further explore the mechanisms at play and the physical properties of the glacier that contribute to surging.

References

- Abe, T., Furuya, M., & Sakakibara, D. (2016). Brief Communication: Twelve-year cyclic surging episodes at Donjek Glacier in Yukon, Canada. *The Cryosphere*, 10, 1427-1432. doi:10.5194/tc-10-1427-2016.
- Allaby, M. (2020). *A Dictionary of Geology and Earth Sciences*. Oxford University Press. eISBN: 9780191874901.
- Altena, B., Scambos, T., Fahnestock, M., & Kääb, A. (2019). Extracting recent short-term glacier velocity evolution over southern Alaska and the Yukon from a large collection of Landsat data. *The Cryosphere*, 13(3), 795-814. <https://doi.org/10.5194/tc-13-795-2019>.
- Arendt, A. A., Luthcke, S. B., Larsen, C. F., Abdalati, W., Krabill, W. B., & Beedle, M. J. (2008). Validation of high-resolution GRACE mascon estimates of glacier mass changes in the St Elias Mountains, Alaska, USA, using aircraft laser altimetry. *Journal of Glaciology*, 54(188), 778-787. doi:10.3189/002214308787780067.
- Banerjee, D., Garg, V., & Thakur, P. K. (2022). Geospatial investigation on transitional (quiescence to surge initiation) phase dynamics of Monacobreen tidewater glacier, Svalbard. *Advances in space research*, 69(4), 1813-1839. <https://doi.org/10.1016/j.asr.2021.08.020>.
- Benn, D. I., Fowler, A. C., Hewitt, I., & Sevestre, H. (2019a). A general theory of glacier surges. *Journal of Glaciology*, 65(253), 701-716. DOI: <https://doi.org/10.1017/jog.2019.62>.
- Benn, D. I., Hewitt I. J., & Luckman A.J. (2023). Enthalpy balance theory unifies diverse glacier surge behaviour. *Annals of Glaciology*, 1–7. doi:10.1017/aog.2023.23.
- Benn, D. I., Jones, R. L., Luckman, A., Fürst, J. J., Hewitt, I., & Sommer, C. (2019b). Mass and enthalpy budget evolution during the surge of a polythermal glacier: a test of theory. *Journal of Glaciology*, 65(253), 717-731. DOI: <https://doi.org/10.1017/jog.2019.63>.
- Bennett, M. R., Huddart, D., & Waller, R. I. (2000). Glaciofluvial crevasse and conduit fills as indicators of supraglacial dewatering during a surge, Skeiðarárjökull, Iceland. *Journal of Glaciology*, 46(152), 25-34. DOI: 10.3189/172756500781833232.
- Berthier, E., Schiefer, E., Clarke, G.K.C., Menounos, B., & Rémy, F. (2010). Contribution of Alaskan glaciers to sea level rise derived from satellite imagery. *Nature Geoscience*, 3(2), 92-95. 10.1038/ngeo737.
- Bevington, A. & Copland, L. (2014). Characteristics of the last five surges of Lowell Glacier, Yukon, Canada, since 1948. *Journal of Glaciology*, 60(219), 113-123. doi: 10.3189/2014JoG13J134.
- Box, J. E., Colgan, W. T., Wouters, B., Burgess, D. O., O'Neel, S., Thomson, L. I., & Mernild, S. H. (2018). Global sea-level contribution from Arctic land ice: 1971-2017. *Environmental Research Letters*, 13(12), 125012. doi: <https://doi.org/10.1088/1748-9326/aaf2ed>.
- Burgess, E. W., Forster, R. R., & Larsen, C. F. (2013). Flow velocities of Alaskan glaciers. *Nature communications*, 4(1), 2146-2146. DOI: 10.1038/ncomms3146.
- Ciraci, E., Velicogna, I., & Swenson, S. (2020). Continuity of the Mass Loss of the World's Glaciers and Ice Caps From the GRACE and GRACE Follow-On Missions. *Geophysical Research Letters*, 47(9), e2019GL086926. Doi: <https://doi.org/10.1029/2019GL086926>.
- Clague, J. J. & Rampton, V. N. (1982). Neoglacial Lake Alsek. *Canadian Journal of Earth Sciences*, 19(1), 94-117. <https://doi.org/10.1139/e82-008>.
- Clarke, G. K. C. (1976). Thermal regulation of glacier surging. *Journal of Glaciology*, 16(74), 231-250. DOI: <https://doi.org/10.3189/S0022143000031567>.

- Clarke, G. K. C. & Holdsworth, G. (2002). Glaciers of the St Elias Mountains. In R.S. Williams, Jr., and J.G. Ferrigno (Eds.), *Satellite image atlas of glaciers of the world: Glaciers of North America—Glaciers of Canada*. (J301–J328). US Geological Survey. <https://pubs.usgs.gov/pp/p1386j/>.
- Clarke, G. K. C., Schmok, J. P., Ommanney, C. S. L., & Collins, S. G. (1986). Characteristics of surge-type glaciers. *Journal of Geophysical Research: Solid Earth*, 91(B7), 7165-7180.
- Colgan, W., Rajaram, H., Abdalati, W., McCutchan, C., Mottram, R., Moussavi, M. S., & Grigsby, S. (2016). Glacier crevasses: Observations, models, and mass balance implications. *Reviews of Geophysics*, 54, 119–161, doi:10.1002/2015RG000504.
- Copland, L., Sharp, M. J., & Dowdeswell, J. A. (2003). The distribution and flow characteristics of surge-type glaciers in the Canadian High Arctic. *Annals of Glaciology*, 36, 73-81. doi:10.3189/172756403781816301
- Copland, L., Sylvestre, T., Bishop, M. P., Shroder, J. F., Seong, Y. B., Owen, L. A., Bush, A. & Kamp, U. (2011). Expanded and recently increased glacier surging in the Karakoram. *Arctic, Antarctic, and Alpine Research*, 43(4), 503-516. DOI: 10.1657/1938-4246-43.4.503.
- Crompton, J. W., Flowers, G.E., Stead, D. (2018). Bedrock Fracture Characteristics as a Possible Control on the Distribution of Surge-Type Glaciers. *Journal of Geophysical Research Earth Surface*, 123(5), 853-873. DOI: 10.1002/2017JF004505.
- Cruikshank, J. (1981). Legend and landscape: Convergence of oral and scientific traditions in the Yukon Territory. *Arctic Anthropology*, 18(2), 67-93.
- Cuffey, K. M. & Paterson, W. S. B. (2010). *The Physics of Glaciers*. Oxford: Butterworth-Heinemann - Elsevier.
- Curley, A., Kochtitzky, W., Edwards, B., & Copland, L. (2021). Glacier changes over the past 144 years at Alexandra Fiord, Ellesmere Island, Canada. *Journal of Glaciology*, 67(263), 511-522. doi:10.1017/jog.2021.4.
- De Paoli, L. & Flowers, G. (2009). Dynamics of a small surge-type glacier using one-dimensional geophysical inversion. *Journal of Glaciology*, 55(194), 1101-1112. doi:10.3189/002214309790794850.
- Dolgoushin, L. D. & Osipova, G. B. (1975). Glacier surges and the problem of their forecasting. *IAHS-AISH Publication*, 104, 292–304.
- Doumbia, C., Castellazzi, P., Rousseau, A. N., & Amaya, M. (2020). High Resolution Mapping of Ice Mass Loss in the Gulf of Alaska From Constrained Forward Modeling of GRACE Data. *Frontiers in Earth Science*, 7. doi:10.3389/feart.2019.00360.
- Dowdeswell, J. A., Hamilton, G. S., & Hagen, J. O. (1991). The duration of the active phase on surge-type glaciers: contrasts between Svalbard and other regions. *Journal of Glaciology*, 37(127), 388-400. DOI: <https://doi.org/10.3189/S0022143000005827>.
- Dowdeswell, J. A., Hodgkins, R., Nuttall, A. M., Hagen, J. O., & Hamilton, G. S. (1995). Mass balance change as a control on the frequency and occurrence of glacier surges in Svalbard, Norwegian High Arctic. *Geophysical Research Letters*, 22(21), 2909-2912. DOI: <https://doi-org.proxy.bib.uottawa.ca/10.1029/95GL02821>.
- Eisen, O., Harrison, W. D., & Raymond, C. F. (2001). The surges of Variegated Glacier, Alaska, USA, and their connection to climate and mass balance. *Journal of Glaciology*, 47(158), 351-358. DOI: <https://doi.org/10.3189/172756501781832179>.

- Eisen, O., Harrison, W. D., Raymond, C. F., Echelmeyer, K. A., Bender, G. A., & Gorda, J. L. D. (2005) Variegated Glacier, Alaska, USA: a century of surges. *Journal of Glaciology*, 51(174), 399–406. doi: 10.3189/172756505781829250.
- Flowers, G.E., Roux, N., Pimentel, S., Schoof, C. (2011). Present dynamics and future prognosis of a slowly surging glacier. *The Cryosphere*, 5(1), 299–213. doi:10.5194/tc-5-299-2011.
- Fowler, A.C. (1987). A theory of glacier surges. *Journal of Geophysical Research*, 92(B9), 9111–9120. doi:10.1029/JB092iB09p09111.
- Fowler, A. C., Murray, T., & Ng, F. S. L. (2001). Thermally controlled glacier surging. *Journal of Glaciology*, 47(159), 527–538. DOI: <https://doi.org/10.3189/172756501781831792>.
- Foy, N., Copland, L., Zdanowicz, C., Demuth, M., & Hopkinson, C. (2011). Recent volume and area changes of Kaskawulsh Glacier, Yukon, Canada. *Journal of Glaciology*, 57(203), 515–525. doi:10.3189/002214311796905596.
- Frappé, T.-P. & Clarke, G. K. C. (2007). Slow surge of Trapridge Glacier, Yukon Territory, Canada. *Journal of Geophysical Research*, 112(F03S32). doi:10.1029/2006JF000607.
- Fu, X. & Zhou, J. (2020). Recent Surge Behavior of Walsh Glacier Revealed by Remote Sensing Data. *Sensors*, 20(3), p.716. doi:10.3390/s20030716.
- Gardner, A. S., Fahnestock, M. A., & Scambos, T. A. (2019). ITS_LIVE Regional Glacier and Ice Sheet Surface Velocities. Data archived at National Snow and Ice Data Center. doi:10.5067/6II6VW8LLWJ7.
- Gardner, A. S., Moholdt, G., Scambos, T., Fahnestock, M., Ligtenberg, S., van den Broeke, M., & Nilsson, J. (2018). Increased West Antarctic and unchanged East Antarctic ice discharge over the last 7 years. *Cryosphere*, 12(2), 521–547. doi:10.5194/tc-12-521-2018.
- Girod, L., Nuth, C., Kääb, A., McNabb, R., & Galland, O. (2017). MMASTER: improved ASTER DEMs for elevation change monitoring. *Remote Sensing*, 9(7), 704. doi:10.3390/rs9070704.
- Hambrey, M. & Alean, J. (2004). *Ice on the move*. In *Glaciers* (pp. 69–100). Cambridge: Cambridge University Press. doi:10.1017/CBO9780511807602.006.
- Hugonnet, R., McNabb, R., Berthier, E., Menounos, B., Nuth, C., Girod, L., Farinotti, D., Huss, M., Dussaillant, I., Brun, F., & Kääb, A. (2021). Accelerated global glacier mass loss in the early twenty-first century. *Nature*, 592(7856), 726–731. DOI: <https://doi.org/10.1038/s41586-021-03436-z>.
- Huss, M. (2013). Density assumptions for converting geodetic glacier volume change to mass change. *The Cryosphere*, 7(3), 877–887. <https://doi.org/10.5194/tc-7-877-2013>.
- Jiskoot, H., Boyle, P., & Murray, T. (1998). The incidence of glacier surging in Svalbard: evidence from multivariate statistics. *Computers & Geosciences*, 24(4), 387–399. DOI: [https://doi.org/10.1016/S0098-3004\(98\)00033-8](https://doi.org/10.1016/S0098-3004(98)00033-8).
- Jiskoot, H., Murray, T., & Boyle, P. (2000). Controls on the distribution of surge-type glaciers in Svalbard. *Journal of Glaciology*, 46(154), 412–422. DOI: <https://doi.org/10.3189/172756500781833115>.
- Kamb, B., Raymond, C. F., Harrison, W. D., Engelhardt, H., Echelmeyer, K. A., Humphrey, N., Brugman, M.M., & Pfeffer, T. (1985). Glacier surge mechanism: 1982–1983 surge of Variegated Glacier, Alaska. *Science*, 227(4686), 469–479. DOI: 10.1126/science.227.4686.469.
- Kochtitzky, W., Copland, L., Painter, M. & Dow, C. (2020a). Draining and filling of ice dammed lakes at the terminus of surge-type Dañ Zhùr (Donjek) Glacier, Yukon, Canada. *Canadian Journal of Earth Sciences*, 57, 1337–1348. doi: 10.1139/cjes-2019-0233.

- Kochtitzky, W., Jiskoot, H., Copland, L., Enderlin, E., McNabb, R., Kreutz, K., & Main, B. (2019). Terminus advance, kinematics and mass redistribution during eight surges of Donjek Glacier, St. Elias Range, Canada, 1935 to 2016. *Journal of Glaciology*, 65(252), 565-579. doi: 10.1017/jog.2019.34.
- Kochtitzky, W., Winski, D., McConnell, E., Kreutz, K., Campbell, S., Enderlin, E., Copland, L., Williamson, S., Main, B. & Jiskoot, H. (2020b). Climate and surging of Donjek Glacier, Yukon, Canada. *Arctic, Antarctic, and Alpine Research*, 52(1), 264-280. doi: 10.1080/15230430.2020.1744397.
- Krimmel, R. M. & Meier, M. F. (1975). Glacier Applications of ERTS Images. *Journal of Glaciology*, 15(73), 391-402. doi:10.3189/S002214300003450X.
- Krumwiede, B. S., Kamp, U., Leonard, G. J., Kargel, J. S., Dashtseren, A., & Walther, M. (2014). Recent glacier changes in the Mongolian Altai Mountains: case studies from Munkh Khaikhan and Tavan Bogd. In J. Kargel, G. Leonard, M. Bishop and A. Kääb (Eds.), *Global land ice measurements from space* (pp. 481–508). Springer, Berlin. doi:10.1007/978-3-540-79818-7_22.
- Lauzon, B. (2022). *Glacier Surge Dynamics on Western Axel Heiberg Island, Nunavut* [Unpublished master's thesis]. University of Ottawa, Ottawa, ON. <http://hdl.handle.net/10393/43984>.
- Lauzon, B., Copland, L., Van Wychen, W., Kochtitzky, W., McNabb, R., & Dahl-Jensen, D. (2023). Dynamics throughout a complete surge of Iceberg Glacier on western Axel Heiberg Island, Canadian High Arctic. *Journal of Glaciology*, 1-18. doi:10.1017/jog.2023.20
- Lawson, W. (1997). Spatial, temporal and kinematic characteristics of surges of Variegated Glacier, Alaska. *Annals of Glaciology*, 24, 95-101. DOI: <https://doi.org/10.3189/S026030550001199X>.
- Lewińska, P., Głowacki, O., Moskalik, M., & Smith, W.A.P. (2021). Evaluation of structure-from-motion for analysis of small-scale glacier dynamics. *Measurement: journal of the International Measurement Confederation*, 168, 108327. <https://doi.org/10.1016/j.measurement.2020.108327>.
- Liu, B., Zou, X., Yi, S., Sneeuw, N., Li, J., & Cai, J. Reconstructing GRACE-like time series of high mountain glacier mass anomalies. *Remote Sensing of Environment*, 280, 113177. <https://doi.org/10.1016/j.rse.2022.113177>.
- Mansell, D., Luckman, A., & Murray, T. (2012). Dynamics of tidewater surge-type glaciers in northwest Svalbard. *Journal of Glaciology*, 58(207), 110-118. doi: 10.3189/2012JoG11J058.
- Marcus, M. G. & Ragle, R. H. (1970). Snow Accumulation in the Icefield Ranges, St. Elias Mountains, Yukon. *Arctic and Alpine Research*, 2(4), 277-292. <https://doi.org/10.2307/1550241>.
- Medrzycka D., Copland L., & Noël B. (2023). Rapid demise and committed loss of Bowman Glacier, northern Ellesmere Island, Arctic Canada. *Journal of Glaciology*, 1–14. <https://doi.org/10.1017/jog.2022.119>
- Medrzycka, D. Copland, L., Van Wychen, W., & Burgess, D. (2019). Seven decades of uninterrupted advance of Good Friday Glacier, Axel Heiberg Island, Arctic Canada. *Journal of Glaciology*, 65(251), 440–452. doi: 10.1017/jog.2019.21.
- Meier, M. F., & Post, A. (1969). What are glacier surges? *Canadian Journal of Earth Sciences*, 6(4), 807-817. <https://doi.org/10.1139/e69-081>.

- Mouginot, J., Bjørk, A. A., Millan, R., Scheuchl, B., & Rignot, E. (2018). Insights on the surge behavior of Storstrømmen and L. Bistrup Bræ, northeast Greenland, over the last century. *Geophysical Research Letters*, 45, 11,197–11,205. doi: 10.1029/2018GL079052.
- Murray, T., Strozzi, T., Luckman, A., Jiskoot, H., & Christakos, P. (2003). Is there a single surge mechanism? Contrasts in dynamics between glacier surges in Svalbard and other regions. *Journal of Geophysical Research: Solid Earth*, 108(B5). doi:10.1029/2002JB001906.
- Nolan, A., Kochtitzky, W., Enderlin, E., McNabb, R., & Kreutz, K. (2021). Kinematics of the exceptionally-short surge cycles of Sít' Kusá (Turner Glacier), Alaska, from 1983 to 2013. *Journal of Glaciology*, 67(264), 744-758. doi:10.1017/jog.2021.29.
- Nuth, C., & Kääb, A. (2011). Co-registration and bias corrections of satellite elevation data sets for quantifying glacier thickness change. *The Cryosphere*, 5(1), 271-290. doi:10.5194/tc-5-271-2011.
- Paul, F. (2015). Revealing glacier flow and surge dynamics from animated satellite image sequences: examples from the Karakoram. *The Cryosphere*, 9(6), 2201–2214. doi:10.5194/tc-9-2201-2015.
- Planet Team (2023). Planet Application Program Interface: In Space for Life on Earth. San Francisco, CA. <https://api.planet.com>.
- Post, A. & Mayo, L.R. (1971). *Glacier dammed lakes and outburst floods in Alaska*, United States Geological Survey. doi: 10.3133/ha455.
- Pritchard, H., Murray, T., Luckman, A., Strozzi, T., & Barr, S. (2005). Glacier surge dynamics of Sortebræ, east Greenland, from synthetic aperture radar feature tracking. *Journal of Geophysical Research*, 110(F3), F03005-n/a. doi: 10.1029/2004JF000233.
- Quincey, D. J., Braun, M., Glasser, N. F., Bishop, M. P., Hewitt, K., & Luckman, A. (2011). Karakoram glacier surge dynamics. *Geophysical Research Letters*, 38, L18504. doi:10.1029/2011GL049004.
- Quincey, D. J., Glasser, N. F., Cook, S. J., & Luckman, A. (2015). Heterogeneity in Karakoram glacier surges, *Journal of Geophysical Research: Earth Surface*, 120, 1288–1300. doi:10.1002/2015JF003515.
- Quincey, D. J. & Luckman, A. (2014). Brief Communication: On the magnitude and frequency of Khurdopin glacier surge events. *The Cryosphere*, 8(2), 571-574. doi:10.5194/tc-8-571-2014.
- RGI Consortium. (2017). Randolph Glacier Inventory - A Dataset of Global Glacier Outlines, Version 6. [Region 1: Alaska]. Boulder, Colorado USA. NSIDC: National Snow and Ice Data Center. doi: <https://doi.org/10.7265/4m1f-gd79>.
- Samsonov, S., Tiampo, K., & Cassotto, R. (2021). Measuring the state and temporal evolution of glaciers in Alaska and Yukon using synthetic-aperture-radar-derived (SAR-derived) 3D time series of glacier surface flow. *The Cryosphere*, 15(9), 4221-4239. <https://doi.org/10.5194/tc-15-4221-2021>.
- Sevestre, H. & Benn, D. I. (2015). Climatic and geometric controls on the global distribution of surge-type glaciers: implications for a unifying model of surging. *Journal of Glaciology*, 61(228), 646-662. doi: 10.3189/2015JoG14J136.
- Sharp, M. (2021). *Amplification of surface topography during surges of Tweedsmuir Glacier* [Unpublished master's thesis]. University of Calgary, Calgary, AB. <http://hdl.handle.net/1880/113988>.
- Shean, D. E., Alexandrov, O., Moratto, Z. M., Smith, B. E., Joughin, I. R., Porter, C., & Morin, P. (2016). An automated, open-source pipeline for mass production of digital elevation

- models (DEMs) from very-high-resolution commercial stereo satellite imagery. *ISPRS Journal of Photogrammetry and Remote Sensing*, 116, 101-117. <http://dx.doi.org/10.1016/j.isprsjprs.2016.03.012>.
- Shugar, D. H., Clague, J. J., Best, J. L., Schoof, C., Willis, M. J., Copland, L. & Roe, G. H. (2017). River piracy and drainage basin reorganization led by climate-driven glacier retreat. *Nature Geoscience*, 10, 370-375. doi: 10.1038/ngeo2932.
- Striberger, J., Björck, S., Benediktsson, Í. Ö., Snowball, I., Uvo, C. B., Ingólfsson, Ó., & Kjær, K. H. (2011). Climatic control of the surge periodicity of an Icelandic outlet glacier. *Journal of quaternary science*, 26(6), 561-565. DOI: 10.1002/jqs.1527.
- Sturm, M. (1987). Observations on the distribution and characteristics of potholes on surging glaciers. *Journal of Geophysical Research*, 92(B9), 9015-9022.
- Thomson, L. & Copland, L. (2016) White Glacier 2014, Axel Heiberg Island, Nunavut: mapped using Structure from Motion methods. *Journal of Maps*, 12(5), 1063-1071. DOI: 10.1080/17445647.2015.1124057
- Urbanski, J. A. (2018). A GIS tool for two-dimensional glacier-terminus change tracking. *Computers & geosciences*, 111, 97-104. <https://doi.org/10.1016/j.cageo.2017.11.004>.
- Van Wychen, W., Burgess, D., Kochtitzky, W., Nikolic, N., Copland, L., & Gray, L. (2021). RADARSAT-2 derived glacier velocities and dynamic discharge estimates for the Canadian High Arctic: 2015-2020. *Canadian Journal of Remote Sensing*, 46(6), 695–714. doi:10.1080/07038992.2020.1859359.
- Van Wychen, W., Copland, L., Jiskoot, H., Gray, L., Sharp, M. & Burgess, D. (2018). Surface velocities of glaciers in western Canada from speckle-tracking of ALOS PALSAR and RADARSAT-2 data. *Canadian Journal of Remote Sensing*, 44(1), 57-66. doi: 10.1080/07038992.2018.1433529.
- Van Wychen, W., Davis, J., Burgess, D. O., Copland, L., Gray, L., Sharp, M., & Mortimer, C. (2016). Characterizing interannual variability of glacier dynamics and dynamic discharge (1999–2015) for the ice masses of Ellesmere and Axel Heiberg Islands, Nunavut, Canada, *Journal of Geophysical Research: Earth Surface*, 121, 39–63. doi:10.1002/2015JF003708.
- Van Wychen, W., Hallé, D. A. M., Copland, L., & Gray, L. (2022). Anomalous surface elevation, velocity, and area changes of Split Lake Glacier, western Prince of Wales Icefield, Canadian High Arctic. *Arctic Science*, 8(4), 1288–1304. [dx.doi.org/10.1139/AS-2021-0039](https://doi.org/10.1139/AS-2021-0039).
- Vargo, L. J., Anderson, B. M., Horgan, H. J., Mackintosh, A. N., Lorrey, A. M., & Thornton, M. (2017). Using structure from motion photogrammetry to measure past glacier changes from historic aerial photographs. *Journal of Glaciology*, 63(242), 1105-1118. doi: 10.1017/jog.2017.79.
- Waechter, A., Copland, L., & Herdes, E. (2015). Modern glacier velocities across the Icefield Ranges, St. Elias Mountains, and variability at selected glaciers from 1959 to 2012. *Journal of Glaciology*, 61(228), 624-634. doi: 10.3189/2015JoG14J147.
- Williamson, S. N., Zdanowicz, C., Anslow, F. S., Clarke, G. K. C., Copland, L., Danby, R., Flowers, G. E., Holdsworth, G., Jarosch, A. H. & Hik, D. S. (2020). Evidence for elevation-dependent warming in the St. Elias Mountains, Yukon, Canada. *Journal of Climate*, 33, 3253-3269. doi: 10.1175/JCLI-D-19-0405.1.
- Wood, W. A. (1963). The Icefield Ranges Research Project. *Geographical Review*, 53(2), 163-184. <https://doi.org/10.2307/212508>.

- Yde, J.C. & Knudsen, N.T. (2005). Glaciological Features in the Initial Quiescent Phase of Kuannersuit Glacier, Greenland. *Geografiska Annaler. Series A, Physical Geography*, 87(3), 473-485. <https://www.jstor.org/stable/3554290>.
- Young, E. M. Flowers, G. E., Berthier, E., & Lato, R. (2021). An imbalancing act: the delayed dynamic response of the Kaskawulsh Glacier to sustained mass loss. *Journal of Glaciology*, 67(262), 313-330. doi: <https://doi.org/10.1017/jog.2020.107>.