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Testing the potential for using Structure from Motion photogrammetry methods to estimate seasonal mass balance on lower Easton Glacier, Mount Baker, WA

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Testing the potential for using Structure from Motion photogrammetry methods to estimate seasonal mass balance on lower Easton Glacier, Mount Baker, WA

By

Elizabeth A. Kimberly

Accepted in Partial Completion of the Requirements for the Degree Master of Science

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Elizabeth Kimberly

April 17th, 2020
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A Thesis
Presented to
The Faculty of
Western Washington University

In Partial Fulfillment
Of the Requirements for the Degree
Master of Science

by
Elizabeth A. Kimberly
April 2020
Abstract

The traditional glaciological method of measuring glacier mass balance is labor-intensive and relies on broad extrapolation of sparse ablation stake data collected in the field to assess mass change across the glacier. In contrast, digital elevation models (DEMs) obtained from unmanned aerial vehicle (UAV) imagery and Structure-from-Motion (SfM) photogrammetry resolve a spatially distributed data set of surface elevation change. In this study, I compare seasonal mass balance estimated by field-based glaciological methods and UAV-SfM methods during summer 2018 on the Easton Glacier, Mount Baker, WA. Total snow and ice surface melt was measured at five ablation stakes between May 20th and September 17th, 2018. A research team at the University of Washington conducted UAV surveys on October 5th, 2017, June 6th, 2018, and October 6th, 2018 and used SfM software to generate DEMs of the Easton glacier’s surface for each date. Change detection maps were created by differencing the DEMs, and surface elevation changes across each differenced DEM were used as proxies for winter accumulation and summer ablation. I used a positive degree day model to adjust the glaciological data to span the SfM survey interval for summer ablation (June 6th - October 6th, 2018). Glacier thickness changes estimated by both methods were converted to water equivalent based on the density of the material lost (snow versus ice). Altitudinal swaths centered on each ablation stake were modified to fit the extent of the UAV imagery, and mass balance was estimated by extrapolating the discrete stake measurements and SfM averages across their respective swaths. SfM methods yield 4.3% less volume loss and 11.4% less mass loss in volume of water across the study area compared to the glaciological method. This discrepancy is likely explained by vertical ice flow related to emergence velocity during the study interval. After adjusting for emergence, SfM estimates overestimate mass balance, likely because of upper limit estimates of the emergence velocity. Uncertainties related to mass balance in crevasses, challenges with horizontal ice flux, and density assumptions are discussed. My study concludes that the influence of secondary processes, particularly emergence/submergence, must be more thoroughly constrained and integrated before SfM-UAV techniques can altogether replace the glaciological method.
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**List of Acronyms**

ALS: Airborne Laser Scanning  
ASTER: Advanced Spaceborne Thermal Emission and Reflection Radiometer  
DDF: Degree Day Factor  
DEM: Digital Elevation Model  
GCP: Ground Control Point  
GNSS: Global Navigation Satellite System  
MFN: Middle Fork Nooksack  
NCNP: North Cascades National Park  
NOCA: North Cascades National Park Service Complex  
NPS: National Park Service  
NRCS: National Resources Conservation Service  
PDD: Positive Degree Day  
PRISM: Parameter-elevation Regressions on Independent Slopes Model  
SfM: Structure from Motion  
SMB: Surface Mass Balance  
Snotel: Snow Telemetry  
SWE: Snow Water Equivalent  
TLS: Terrestrial Laser Scanning  
UAV: Unmanned Aerial Vehicles  
USGS: United States Geological Survey  
UW: University of Washington  
USDA: United States Department of Agriculture

**Terminology (from Cogley et al., 2011)**

\[ \Delta h = \text{change in thickness from snow or ice ablation measured by the glaciological method} \]

\[ \Delta z = \text{change in surface elevation measured by SfM differencing} \]

\[ b_w = \text{winter mass balance} \]

\[ b_s = \text{summer mass balance} \]

\[ S = \text{area} \]

\[ \Delta V = \text{change in volume between successive surveys} \]

\[ k = \text{swath number} \]

\[ s = \text{snow} \]

\[ i = \text{ice} \]
$\Delta Q = \text{flux divergence term}$
$\rho = \text{density}$
1.0 Introduction

The North Cascade Range in Washington is the most glacierized region in the contiguous United States and its glaciers have significant influences on related ecological, hydrologic, geologic, and socio-economic systems (Bidlake et al., 2010; Riedel and Larrabee, 2011a; Grah and Beaulieu, 2013). For example, glacial dynamics affect the magnitude and timing of down-valley streamflow, which in turn have major implications for the biotic and abiotic communities that depend on glacial meltwater (Grah and Beaulieu, 2013). Glaciers also release fine-grained sediment into their meltwater, which impacts nutrient concentrations and aquatic habitat (Riedel and Larrabee, 2011). Glaciers are proxies for natural and anthropogenic climate change as they respond to perturbations in temperature and precipitation on intra-annual to decadal timescales (Vargo et al., 2017). Because mountain glaciers exist in remote regions and at high altitudes, their mass balances can describe climate trends in regions where other climatic variables and observations are difficult to measure (Josberger et al., 2007). Warmer temperatures and lower snowfalls in the 21st century are causing these glaciers to thin, retreat, and in some cases, disappear altogether (Pelto and Brown, 2012; Menounos et al., 2019).

Glacier mass balance studies have been conducted globally because mass balance measurements are the primary means to directly measure the response of glaciers to climate change; there are many long-term monitoring projects in the Alps, Antarctica, Scandinavia, the Andes, Alaska, and in the North Cascades (Zemp et al., 2013). These studies predominantly rely on the glaciological method developed in the 1950s to quantify glacial change and the data almost unanimously show that glaciers worldwide are losing mass (Zemp et al., 2013). However, the glaciological method is labor-intensive and expensive, and continuous, long-term surveys are available for only a handful of glaciers. Thus, one or two glaciers are often used to represent mass balance for hundreds of glaciers in a region (Cox and March, 2004). Cox and March (2004) also suggested that the inaccuracy of the mass balance records on these glaciers is greater than the inaccuracies caused by the region-wide extrapolation of a single mass balance record due to inherent errors within the glaciological method.

Several studies have measured glacier mass balance in the North Cascades but these efforts are limited in their spatial coverage and resolution because of the time-consuming, labor-intensive nature of the glaciological method, as well as challenges with access (Riedel and Larrabee, 2011). The glaciological method records measurements at a small number of discrete
locations, then relies on interpolation and extrapolation to construct a spatially distributed assessment of mass gain and loss across an entire glacier. Relying on point measurements to represent an entire glacier involves significant assumptions about spatial and temporal homogeneity of snow/ice accumulation and loss.

Methods to more accurately, precisely, and efficiently measure glacier mass balance have improved with the advancement of geospatial and remote sensing techniques, but data collection is still challenging in remote, alpine environments (Westoby et al., 2012). For example, many glaciers exist in regions where satellite coverage is poor and where airborne surveys are limited by the complex terrain of mountainous landscapes, which make lines of sight difficult to maintain (Westoby et al., 2012). Although airborne lidar can produce highly accurate and relatively precise digital elevation models (DEMs), it is prohibitively expensive to collect lidar imagery at frequent intervals, particularly in remote regions (Ryan et al., 2015). Additionally, the steep, rugged terrain akin to glacial landscapes makes it hard to acquire data with the more precise ground-based methods like total station surveys or terrestrial laser scanning (TLS; Westoby et al., 2012).

Recently developed methodologies using aerial imagery combined with Structure-from-Motion (SfM) software have the potential to quantify glacial change at fine spatial resolutions and temporal frequencies, and likely at much lower costs than other digital survey methods (Nolan et al., 2015). Employing cheaper, lightweight, and more autonomous surveying devices, like unmanned aerial vehicles (UAVs), also eliminates the need to rely on traditional airborne surveys for image acquisition, and increases the accessibility of remote regions. More precise and spatially extensive measurements of net glacial melt may provide more accurate data for monitoring alpine glacier health and tuning hydrological models that predict streamflow responses to glacial melt. It can also help us better understand glacier energy balance, spatial variability, and secondary accumulation by drift and avalanche.

My research focuses on the ablation zone of the Easton Glacier on the southwest slopes of Mount Baker in northwest Washington (Figure 1). I test the following questions: can glacier mass balance be determined accurately from SfM analyses of UAV-derived aerial images and can it improve upon traditional glaciological mass balance field methods? Having spatially complete mass balance data at a much higher resolution, potentially collected more efficiently, could offer new evidence of the rates and severity of climate change and better inform policy and
management decisions. The results of this study indicate that although repeat SfM surveys hold great promise for streamlining mass balance measurements, there are a number of crucial problems with converting SfM data to true mass balance data.

2.0 Background and Previous Research

2.1 Glaciological Method

2.1.1 Concept, Process, and Limitations

To gather mass balance data using traditional glaciological methods, researchers transport bulky equipment to a glacier during periods of peak net snow accumulation (typically late spring in the western Cordillera), drill and install a series of ablation stakes into the snow and ice along the length of a glacier, and return to the sites multiple times throughout the melt season to record snow or ice accumulation and loss. These measurements are then converted to water equivalent and interpolated and/or extrapolated across altitudinal swaths between each stake to estimate the total mass loss throughout a melt season. Any remaining snow at the end of the ablation season is classified as firn.

Altitude is the dominant control on glacial accumulation and ablation because it is highly correlated with temperature, snowfall amounts, and wind (Fountain and Vecchia, 1999). Thus, survey stakes are typically placed in a longitudinal transect up the centerline of a glacier to ensure a spatial representation of each altitudinal swath. However, because this method involves interpolation and extrapolation from point measurements, it requires assumptions of spatial homogeneity that introduce error. For example, glaciers may experience enhanced accumulation along their edges from avalanches and rock-fall. Wind drifting, topographic shading, and avalanches can also lead to locally high areas of accumulation. The magnitude of this spatial variability can be significant and can occur over small areas: Braithwaite and Olsen found variations of 0.23 meters of water equivalent (m w.e.) in one year among three stakes that were less than five meters apart (Braithwaite and Olesen, 1989). Ablation may vary spatially because of changes in albedo, effects of aspect, and proximity to crevasses. Fountain and Vecchia (1999) found that error significantly increases as the number of observation sites decreases, especially on small alpine glaciers where the aforementioned accumulation/ablation effects are more profound. This could be addressed by installing spatially distributed stakes across glaciers or in
secondary accumulation zones, but it demands more fieldwork as well as knowledge of a glacier’s locations of secondary accumulation (Riedel et al. 2008).

In addition to spatial heterogeneities of accumulation and ablation, there can be practical limitations to installing evenly spaced ablation stakes across a glacier. Crevasses, ice-falls, steep slopes, and financial restrictions can limit where and how many stakes can be installed, thereby decreasing the sample size and increasing the uncertainties with measuring mass balance from the stakes. Additionally, financial restrictions and the logistics of transporting installation equipment with snowmobiles or helicopters can prevent stakes from being installed at the time of maximum accumulation. To compensate for this, scientists often use a temperature-based technique, like a degree-day model, to quantify the ablation that takes place before stake installation (or after stake removal at the end of the season) – further increasing error (Ohmura, 2001; Cox and March, 2004; Rasmussen and Wenger, 2009).

Additionally, accumulation and ablation seasons do not always correspond to winter and summer seasons. On low-latitude glaciers, ablation and accumulation can occur at any point in the year and accumulation and ablation often occur simultaneously in the Himalaya (Fountain and Vecchia, 1999). This problem may become more widespread as climate change induces anomalous weather (i.e., significant mid-winter ablation and/or mid-summer accumulation in the Cascades). Furthermore, stakes can be damaged or lost during surveys, resulting in permanent loss of the data associated with that stake for a given ablation season. Ablation stakes have also been known to sink into the snowpack (Cox and March, 2004; Riedel and Larrabee, 2011). Additionally, although surface ablation dominates glacier mass balance (about 90%, according to Mayo et al., 1972), glaciological methods do not account for internal and basal accumulation and ablation, such as mass loss from geothermal heat (Kaser et al., 2006; Bidlake et al., 2010). The United States Geological Survey (USGS) incorporates an estimated absolute value of 0.05 m w.e. a\(^{-1}\) into their annual mass balances to account for internal ablation (Cox and March, 2004).

Lastly, the glaciological method relies on density measurements to convert snow and ice ablation to water equivalent volume. Density can be measured in the field by extracting snow from snow pits or from snow density cores, and weighing the mass of the contents. Spring and summer snowpack densities on the South Cascade Glacier vary consistently with snowpack depth and altitude from year to year, so the USGS uses snow density schemes to make estimations (Krimmel, 1999). Similarly, when the National Park Service (NPS) has not made in
measurements of snow density, they use the average density of the spring snowpack since 1993, which is 0.5 ± .08 g/ml (Riedel et al., 2008). These estimates are corroborated by density data at Snow Telemetry (Snotel) sites and they reduce labor intensity and time demands, but they introduce more uncertainty.

2.1.2. Glacial monitoring in the North Cascades

The USGS and the North Cascades National Park Service (NCNP) have been conducting glacial monitoring studies in the North Cascades for many decades (Harper, 1993; Bidlake et al., 2010). These studies have documented glacial trends and improved our understanding of climate change effects on glaciers and the links among glacial retreat, water resources, hydrological hazards, and aquatic and terrestrial ecosystems (Bidlake et al., 2010; Riedel and Larrabee, 2011). The South Cascade Glacier, the longest continuously monitored glacier in North America, has been termed one of five benchmark glaciers by the USGS; researchers began to annually monitor its mass balance and related hydrological and meteorological variables in 1957 (Bidlake et al., 2010). The USGS also installed temperature and precipitation instruments to account for meteorological conditions.

The NCNP has used the glaciological method to monitor four glaciers in the North Cascades National Park Service Complex (NOCA): Silver Creek, North Klawatti, and Noisy Creek (since 1993), and Sandalee Glacier (since 1995). These glaciers were selected because they feed meltwater into four different watersheds and are located at a variety of elevations and aspects. The North Cascade glaciers have followed the global trend of cumulative mass loss throughout that time (Riedel and Larrabee, 2016). For example, the average annual melt rate for the four NPS glaciers increased by about 10% (1 m w.e.) between 1993 and 2009 (Riedel and Larrabee, 2011).

2.2 Geodetic Methods

2.2.1. Concept and Process

To remedy the challenges and limitations of the glaciological method, scientists have tested geodetic methods to measure glacier mass balance. Geodetic methods use DEMs or topographic maps to measure the surface elevation of a geomorphic surface. By differencing DEMs created from repeat surveys (known as “change detection”), ongoing changes can be monitored through time (Whitehead et al., 2013).

The potential for change detection with digital elevation surveys to assess changes in
glacier mass balance is only beginning to be tested rigorously (Sold et al., 2013; Belart et al., 2017; Klug et al., 2018; Shean et al., 2020). Some studies apply the geodetic method over a single accumulation season (Sold et al., 2013; Belart et al., 2017) while others measure annual balance over multiple years (Klug et al., 2018). Because this method covers the entire glacier surface, it accounts for spatial variability in accumulation and ablation at a high resolution. In order to accurately extract accumulation and ablation from changes in a glacier’s surface elevation across portions of the glacier, corrections for density and glacier flow between surveys must be applied (Van Beusekom et al., 2010; Cox and March, 2004; Sold et al., 2013).

Numerous studies have used photogrammetry techniques on aerial photographs (Beedle et al., 2014), topographic map reconstructions (Geck et al., 2013), or satellite or airborne stereo imagery (e.g., DigitalGlobe WorldView, Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER), Pleiades satellite imagery) (Berthier et al., 2010; Belart et al., 2017; Shean et al., 2020) to determine surface elevation changes between subsequent dates. Others have used repeat altimetry techniques, like lidar and laser altimeters, to directly measure and compare surface elevation at different times (Bamber and Rivera, 2007; Sold et al., 2013; Das et al., 2014; Pelto et al., 2019). However, these techniques generate elevation data with relatively coarse resolution and vertical error on the order of tens of meters (Immerzeel et al., 2014).

Previous studies have compared mass balance data estimated by geodetic photogrammetric data and the glaciological method. Cox and March (2004) found a strong correlation between mass balance measurements derived from the two methods on the Gulkana Glacier in Alaska. Conversely, Van Beusekom et al. (2010) found conflicting mass balance trends for the Wolverine Glacier in Alaska; the field data collected by the glaciological method measured net melt whereas the geodetic methods calculated a positive cumulative balance. They attributed this discrepancy to error in the glaciological method: the ablation stakes were located in areas with local accumulation anomalies due to large wind drifts or avalanches, which significantly biased their mass balance results.

### 2.2.2 Structure-from-Motion

Structure-from-Motion (SfM) is a digital photogrammetric imaging technique that applies the geodetic concept. SfM combines repeat, overlapping two-dimensional photographs taken by planes, unmanned aerial vehicles (UAVs), balloons, or kites into SfM software to create DEMs
As the flight device flies in a predetermined grid pattern over an area of interest, it photographs each spot many times and from different angles so that the site is fully covered by many stereo image combinations (Whitehead et al., 2013). Ground control points (GCPs), typically distinctive targets located with survey-grade Global Navigation Satellite System (GNSS), are placed within the surveyed landscape to allow co-registration of the imagery. SfM software then automatically compares and aligns the images to produce point clouds of the aerial imagery, which are used to develop high-resolution DEMs of a surface with resolutions as high as centimeters depending on camera resolution and altitude of flight (Alfredsen et al., 2018).

Imagery compatible with SfM can be collected by fixed-wing aircraft, helicopters, or satellites, but in the past, the costliness of these flights has required that researchers rely on imagery collected for government purposes, like map updates (Whitehead et al., 2013). The autonomy and relative affordability of UAVs, however, allow researchers to collect similar data on a more timely and flexible agenda, and on an ongoing basis (Whitehead et al., 2013). Advances with SfM software have made image processing more widely accessible than high resolution satellite imagery and more affordable than lidar collection (Whitehead et al., 2013).

Although the value of differencing SfM surveys to assess rates and styles of active geological landscapes has been well documented in other geologic applications, (e.g., Mancini et al., 2013; Lucieer et al., 2014; Piermattei et al., 2015) studies have only recently begun to apply SfM to the cryosphere. Whitehead et al. (2013) measured surface motion and elevation changes on the order of meters on the Fountain Glacier using UAVs and a photogrammetry imaging software called Inpho. Nolan et al. (2015) demonstrated that SfM could quantify changes in snow depth across Alaska at unprecedented resolutions. Vargo et al. (2017) applied SfM to measure variations in equilibrium line altitudes on several New Zealand glaciers using historic photographs. Alfredsen et al. (2018) removed the need to access a remote and dangerous field site by mapping river ice thickness and spatial distribution with drones and SfM. Ryan et al. (2015) used similar methods to quantify calving dynamics, ice flow, crevasse patterns, and terminus thickness of a tidewater glacier in Greenland over a 19-hour and 52-day interval. Through the employment of SfM and change detection, all of these studies allowed researchers to obtain data from remote areas at higher accuracy and more frequent intervals than previously used techniques.
Few studies have applied UAV-SfM methods to quantify surface melt or mass balance on a glacier’s surface between sequential surveys (Whitehead et al., 2013; Immerzeel et al., 2014; Bash et al., 2018). Furthermore, to the best of my knowledge, there are no studies currently published that explicitly test and compare mass balance data collected from ablation stake methods to those derived by UAVs and SfM. Bash et al. (2018) compared surface melt estimated by UAV-SfM methods and in situ ablation stake measurements over short (1-3 day) time intervals. They found good agreement between the two methods, and attribute differences to errors with SfM reconstruction and uncertainties in their GCP placement. They did not translate their surface melt measurements to mass balances, and their study was conducted over short time-scales (Bash et al., 2018).

2.2.3 Geodetic Limitations

Geodetic methods have a number of limitations. For example, snow-covered areas may not provide sufficient photogrammetric contrast to accurately register and difference DEMs from two different times. Crevasses, surface debris, and exposed ice typically provide enough definition, but such prominent features are usually snow-covered in the accumulation season (Cox and March, 2004).

The geodetic method also requires density values to convert snow and ice to water volume, which present a significant challenge, especially because there is often minimal correlation between snow density and elevation or depth (Huss, 2013; Pelto et al., 2019). Studies have used the density of ice (0.9 g ml) as a constant density value to convert all volume loss to mass balance, but they thereby overestimate the mass change when there is substantial snow and/or firn loss (Elsberg et al., 2001; Geck et al., 2013). To account for loss of lower density material, Huss (2013) suggests a glacier-wide value of 0.85 ± 0.6 g ml. This assumption produces an acceptable uncertainty in mass balance for geodetic studies over multiple years to decades, but not for shorter time scales like seasonal balances (Pelto et al., 2019). Seasonal geodetic balances, however, could track the amount of snow left over at the end of the melt season, and thereby estimate firn ablation the following summer (density of 0.70 g/ml).

To better constrain density variability, Pelto et al. (2019) used satellite imagery to classify surface material as spring snow, late summer snow, firn, or ice, and then applied corresponding densities to account for varying surface density and to calculate a seasonal balance (Klug et al., 2018; Pelto et al., 2019). In their study, they found that regional
observations of late summer snow density were consistent for glaciers across the Pacific Northwest (0.53 to 0.63 g/ml), and they assigned this value to all pixels on surfaces classified as late summer snow. Annual glacier monitoring studies by the USGS corroborate this density assumption; spring snow density on South Cascade Glacier between 1,618 and 1,660 meters averaged 0.502 ± 0.022 g/ml from 1986 to 2003 (Riedel et al., 2008). Similarly, spring snow density on North Cascades Glaciers monitored by the NPS ranged from 0.41 to 0.53 g/ml from 1993 to 2003 (Riedel et al., 2008). Although snow density varies with altitude on some glaciers, they suggest that 0.5 g/ml is representative of the entire glacier (Riedel et al., 2008). Errors introduced by these assumptions are further complicated by factors such as firn densification and high spatial variability in density (Huss, 2013). However, density has less spatial variability than snow depth (Pelto et al., 2019).

Unlike the glaciological method, which can measure thickness change of the same point on a glacier through time (i.e., within a Lagrangian frame of reference – the stake is moving with the glacier flow), the geodetic method operates in a fixed Eulerian frame of reference. Thus, change detection surveys of two DEMs can yield a change in surface elevation at a single point in space but not on the same parcel of glacier ice because of the flow of ice between surveys (Cox and March, 2004). Modern UAVs have limited flight distances and battery life in cold conditions and at altitude which make it difficult (and sometimes impossible) to survey entire glaciers (Ryan et al., 2015). For example, to survey the entire length of the 16 km Fountain Glacier, scientists would need to hike to the upper glacier and conduct multiple flights to maintain a line of site while taking off, landing, and flying the UAV (Whitehead et al., 2013).

2.2.4 Vertical Ice Flow

Glaciers convey ice from locations of net snowfall excess to locations of net loss through horizontal and vertical ice fluxes. Thus, in addition to uncertainties related to horizontal ice flow (e.g., Hodge, 1974), geodetic change-detection surveys must consider vertical ice movement as a result of flow redistribution. Vertical flow velocity is negative/downward in the accumulation area (submergence) and positive/upward in the ablation area (emergence) (Cuffey and Paterson, 2010; Sold et al., 2013). Because the velocity approaches zero at the terminus, the glacier adjusts to increased summer velocity by compressing in the ablation zone, causing the ice to bulge upward and emerge vertically. This can be visible in the winter when surface ablation is
essentially zero, and time-lapse photos show glaciers “puffing up” near the terminus (e.g., Sólheimajökull Glacier, Iceland, https://vimeo.com/6039933). Emergence is not as apparent in the summer because it is typically matched or outpaced by ablation.

Emergence and submergence are equal to the upward or downward flow of ice relative to a local plane of the glacier surface (Cuffey and Paterson, 2010). Any change in the surface elevation at a point that is not caused by accumulation or ablation, and is not from the vertical component/elevation loss of down-glacier flow, can be ascribed to a glacier’s emergence or submergence velocity. Thus, surface elevation change measured by differenced DEMs incorporates 1) surface snow and/or ice accumulation or ablation, 2) advection of topography from horizontal flow, and 3) the vertical displacement of the ice surface as a result of glacier flow (Sold et al., 2013). The vertical component of ice flow (emergence/submergence, $\Delta Q$), is calculated by the following equation:

$$\Delta h = \frac{b}{\rho} - \Delta Q$$  \[1\]

where $\Delta h$ is the rate of thickness change, $b$ is the specific surface mass balance rate, $\rho$ is density of the material, and $\Delta Q$ is a flux-divergence term that quantifies the emergence or submergence velocity (e.g., Rasmussen and Krimmel, 1999; Cuffey and Paterson, 2010).

Vertical velocity is considered negligible when calculating glacier-wide balance because it is simply a redistribution of mass within the glacier (Cuffey and Paterson, 2010; Immerzeel et al., 2014; Klug et al., 2018). Mass balance at specific points or across swaths of a glacier cannot be accurately derived from SfM change detection without a detailed assessment of the vertical component of ice velocity (Sold et al., 2013; Beedle et al., 2014, 2015; Belart et al., 2017). When estimating volume change and mass balance from geodetic methods, it is necessary to estimate the magnitude of this ice motion to account for surface height changes related to emergence/submergence. Additionally, emergence/submergence rates vary seasonally and spatially, so inter-annual and spatial variation in their magnitudes should be considered as well (Hodge, 1974; Sold et al., 2013).

3.0 Study Site

The Easton Glacier (2018 surface area: 232,200 m$^2$) is one of eleven major glaciers on Mount Baker in the North Cascades of Northwest Washington, about 50 km northeast of Bellingham, WA (Figure 1) (Harper, 1993; Pelto and Brown, 2012). The summit of Mt. Baker is
at 3286 m, and the mountain receives ~7-12 meters of snow annually due to the region’s highland climate and maritime influence (Riedel and Larrabee, 2016). The 30-year normal average precipitation on Mount Baker is ~5.5 meters according to the Parameter-elevation Regressions on Independent Slopes Model (PRISM Climate Group, 2020). The Easton flows down the southwest face of Mount Baker from its head at ~2,950 meters to its terminus at ~1,700 meters. Easton’s two prominent lateral moraines, the Railroad Grade and Metcalf moraines, record the thickness and width of its Little Ice Age extent. The Easton Glacier is a major contributor of meltwater to Baker Lake and the Puget Sound’s largest watershed, the Skagit watershed (Riedel and Larrabee, 2016).

Since the end of the Little Ice Age, the Easton Glacier has broadly followed the trends of glacial retreat and advance across the North Cascades. A substantial increase in temperatures between the 1880s and 1940s led to rapid retreat for North Cascade glaciers, including the Easton. Mapping based on aerial photographic data compiled by Pelto and Brown (2012) suggest the Easton Glacier’s terminus retreated 2,420 meters between its Little Ice Age Maximum and 1950. Its terminus then advanced ~680 meters from 1960-1989 in response to wetter and cooler conditions (Harper, 1993; Pelto and Brown, 2012). Following this advance, the Easton Glacier retreated 300 meters and its surface elevation lowered an average of 13 meters between 1990 and 2009 (Pelto and Brown, 2012).

The Easton Glacier is well suited to conduct this study because of its accessibility, historical mass balance data, and environmental significance. The Scott-Paul/Railroad Grade hiking trails in the Mount Baker-Snoqualmie National Forest provide ready access to the glacier about three miles from the Schreiber Meadows Trailhead. Also, unlike other glaciers on Mount Baker, it is in a National Recreation Area (rather than wilderness) so the use of powered vehicles is permitted. This allowed my research team to use snowmobiles to transport equipment to install the ablation stakes. Additionally, Dr. David Shean (University of Washington) had already begun collecting photogrammetric data on the Easton Glacier and has solved some of the difficulties that have arisen with the UAV and the image collection process. Due to limited UAV flight coverage, however, my study focuses on the ablation zone of the Easton Glacier (Figure 2). By focusing on the ablation zone, I eliminate the need to account for the confounding influence of firn densification on surface elevation change, which is discussed later.
4.0 Methods

Here I summarize the methods used to: 1) quantify mass change across the lower portion of the Easton Glacier between June 6th and October 6th, 2018 with the glaciological method, and 2) estimate mass change from UAV-derived aerial imagery and SfM processing for the same time interval and survey area. Figure 3 shows the series of steps I follow to estimate total ablation volume for each method.

4.1 Glaciological Method

4.1.1 Field Measurements

Following protocols established by the NPS on nearby glaciers in NCNP, the glaciological method was used to measure surface mass balance (Riedel et al., 2008). I installed five ablation stakes along a longitudinal transect up the Easton Glacier (Figure 2). Stakes 4 and 5 were installed in the accumulation zone on May 5th, 2018 and stakes 3, 2, and 1 were installed in the ablation zone on May 20th, 2018. Accessible, crevasse-free zones were located on lidar to approximate suitable locations for the stake sites before going into the field. The Northwest Cruisers Snowmobile Club transported the necessary installation gear to the ablation stake sites. Using a portable propane steam drill borrowed from NCNP, we melted holes in the snow and ice to insert the ablation stakes; each stake was comprised of 4-6 1.5 meter segments of 2.54 cm diameter PVC pipe connected with wooden dowels and duct tape. I perforated the lower end of each ablation stake to limit the potential for floating ablation stakes – this can be problematic when the drill holes fill with melt water (Riedel et al., 2008). Before drilling, we probed each stake site with 10-mm-diameter steel probes to evaluate snow depth variability along contour, and to ensure we did not install a stake into a buried crevasse.

To estimate snow depths at each stake site, we probed to the previous summer surface 3-5 times along transverse profiles, and averaged these values to represent total winter accumulation (Appendix I, Table 2). We hit an impenetrable firn or interstitial ice layer at stakes 4 and 5 and did not get accurate snow depths there. A Trimble GeoXH 6000 mapping grade GNSS unit recorded the GNSS position coordinates and elevations at each stake (Appendix I, Table 1). Post-processing corrections of the data typically result in horizontal uncertainties of 5-10 cm and vertical uncertainties of 10-20 cm. We revisited each ablation stake site 1-5 times throughout the ablation season to record surface snow and/or ice melt (and any accumulation) between subsequent field visits (Appendix I, Table 2). The variation in field visits reflect the technical
difficulty of visiting the site, and limitations due to weather and poor air quality (wildfire smoke). We were unable to access stake 4 after July 24th, and stake 5 fell into a crevasse, so I discarded data from these sites.

4.1.2. Meteorological Constraints

To constrain meteorological conditions on the glacier during the ablation season, I installed a Campbell Scientific research-grade WxPro weather station directly adjacent to the glacier. The weather station was established near the highest point on the left-lateral Metcalfe moraine at 5576 m and was deployed between July 20th and October 6th, 2018. The weather station recorded temperature, wind speed and direction, total precipitation, and solar radiation at 30 minute intervals. The temperature data were used to establish a local lapse rate relative to the nearby Middle Fork Nooksack (MFN) Snotel site and to evaluate other climatic variables affecting the glacier.

4.1.3 Glaciological Data Analyses

4.1.3.1. Ablation Stake Data

I measured the surface melt between field visits by differencing the stake height above the surface at the time of visit relative to its height above the surface at the last visit, including any removed sections (Figure 11). Total surface melt encompasses thickness of snow melt combined with any additional ice melt at each stake. Since I did not measure in situ densities for snow, firn, and ice in the field, I used well-established average snow-water equivalents (SWE) for snow and ice on alpine glaciers in the North Cascades to convert surface melt to m w.e. (Riedel and Larrabee, 2011).

4.1.3.2. Altitudinal Swaths

To represent spatially distributed estimates of glacier ablation, I extrapolated the measurements at each ablation stake location across altitudinal swaths of the glacier (Figure 2). Because accumulation and ablation are largely influenced by altitude, topographic contours derived from DEMs determined the outlines of these swaths (Fountain and Vecchia, 1999). The swath boundaries follow the elevation contour equidistant between two stakes so that each stake was located at the median altitude of a swath. The swath boundaries were further modified to fit the extent of David Shean’s aerial imagery to ensure that there was SfM elevation data for the entire swath extent (Figure 2).
4.1.3.3. Positive Degree-Day Model

Completing a UAV survey and visiting multiple ablation stakes in a single day would involve substantial logistical coordination, but it would allow a direct comparison to be made without any data extrapolation. Because it was logistically difficult, we did not visit the ablation stake sites on the dates that we collected the UAV imagery (June 6th and October 6th, 2018). In order to compare the two data-sets, I used a positive degree-day (PDD) model to adjust the stake measurements to span the UAV survey interval. In doing this, I reduced the glaciological winter balance (melt from May 20th to June 6th) and increased the glaciological summer balance (September 17th to October 6th). A positive degree-day model assumes a positive empirical relationship between temperature and snow/ice melt: it establishes that a certain depth of snow/ice – known as the degree-day factor – melts for every 1°C above 0°C (Ohmura, 2001). Climactic variables like wind speed and sensible heat flux control the magnitude of the degree-day factor, causing them to vary among glaciers (Rasmussen and Wenger, 2009). Although PDD models are widely used and have proven to be effective, they introduce some amount of uncertainty.

Because the weather station was not deployed until July 17th, I used air temperature from a nearby Snotel site: the MFN Snotel (48°, 49’N, 121°, 56’ W). The MFN is ~180 m lower than the weather station, so a lapse rate was calculated to convert the Snotel temperatures to air temperatures at the weather station and at each stake. Using a linear regression (y = 1.1092x – 2.3195 and R² = 0.99) between daily average temperatures at the MFN Snotel site and the weather station, I adjusted the Snotel data to weather station temperatures for June 6th to October 6th, 2018 (Figure 4). Since only positive temperatures are meaningful for the PDD model, all negative temperatures were set to 0. Next, I calculated the average difference in daily average temperatures at the weather station and MFN for the survey interval. Given an elevation difference of ~180 m between MFN and the weather station, and a daily average temperature 1.085 °C higher at the MFN Snotel, I calculated a temperature lapse rate of -0.595 °C per 100 meter elevation gain. This lapse rate was used to extrapolate temperatures from the weather station to each ablation stake. The following positive degree-day equation was used to find the total snow and/or ice melt at each stake for time intervals that lack ablation stake data (Braithwaite and Olesen, 1989):

\[ M = K_I(PDD_I) + K_S(PDD_S) \]
where \( M = \) depth of snow/ice melted (m), \( K_I = \) the degree-day factor for ice (m °C\(^{-1}\) day\(^{-1}\)), \( K_S = \) the degree-day factor for snow (m °C\(^{-1}\) day\(^{-1}\)), \( PDD_I = \) the total number of positive degree days melting ice (°C), \( PDD_S = \) the total number of positive degree days melting snow (°C).

The degree-day factor (DDF) for snow (\( K_S \)) and ice (\( K_I \)) was calculated from field measurements of snow and ice melt at each stake and total positive degree days for the different intervals (Appendix I, Tables 3 and 4). For example, field measurements recorded 1.28 meters of snow melt at stake 1 between May 21\(^{st}\) and June 17\(^{th}\), 2018. During this same time interval, the adjusted weather station data records ~148 positive degree-days. Thus, \( K_S \) at stake 1 during the time interval is 0.0086 m °C\(^{-1}\) day\(^{-1}\). I calculated and applied \( K_I \) to estimate ablation that occurred between the last stake measurement on September 17\(^{th}\) and the UAV survey on October 6\(^{th}\), when there was entirely ice melt in the survey area (Appendix I, Table 4).

4.2 Structure-from-Motion Methods

For the SfM portion of the study, I collaborated with Dr. David Shean at the University of Washington (UW) who has been conducting repeat SfM surveys of lower Easton Glacier using UAVs and stereo satellite imagery over the past five years. David and his research team flew a Fixed-Wing eBee RTK mounted with a Sony S110 RGB camera to collect overlapping aerial photographs on October 5\(^{th}\), 2017, June 6\(^{th}\), 2018, and October 6\(^{th}\), 2018. Prior to the flights, we used a Trimble R10 and GNSS base station to survey GCPs near the terminus of the glacier and on the adjacent moraines. Using SfM Agisoft Photoscan Pro software, David Shean triangulated the positions of points that were photographed many times, and produced point clouds from each survey (Nolan et al., 2015). Each point in a point cloud has a X, Y, and Z coordinate representing the glacier’s surface. From the point clouds, David generated DEMs of the surveyed area on the three survey dates with the “point2dem tool” in Ames Stereo Pipeline – a suite of tools developed by NASA to process stereo imagery (Shean et al., 2016). I differenced the DEMs to create change-detection maps of the glacier between successive surveys.

4.2.1 SfM - Winter Accumulation

To estimate the winter balance with SfM methods, I differenced DEMs obtained from UAV flights on October 5\(^{th}\), 2017 and on June 6\(^{th}\), 2018. The resulting map, dDEMW, quantifies the change in surface elevation across swaths 1, 2, and 3 for the 2017-2018 winter accumulation season. These provide a rough constraint of June 6\(^{th}\) snow depths, without adjusting for horizontal or vertical ice flow. Because I cannot yield point measurements from dDEMW, I
extracted the average surface elevation change ($\Delta z$) across each swath to represent the height of snow accumulation. Following Klug et al. (2018), the total volume change ($\Delta V$) and approximation for winter balance ($b_{\text{wSfM}}$) for swaths 1, 2, and 3 was calculated with the following expressions:

$$\Delta V = \sum_{k=1}^{3} \Delta z_k \times S \quad [3]$$

$$b_{\text{wSfM}} = \sum_{k=1}^{3} (0.5 \times \Delta z_k^e) \times S \quad [4]$$

where $\Delta V$ = volume change between surveys (m$^3$), $k$ = swath number, $b_{\text{wSfM}}$ = winter mass balance estimated by SfM between surveys (m$^3$ w.e.), $\Delta z_k^e$ = average elevation change due to snow accumulation across dDEMW for swath $k$ (m), and $S$ = swath area (m$^2$).

According to the record of average daily SWE listed on the United States Department of Agriculture (USDA) Natural Resources Conservation Service (NRCS) Washington website, the MFN Snotel recorded a peak SWE of ~2.2 m w.e. on April 29th, 2018. Thus, dDEMW likely does not record the true maximum winter accumulation as there was likely significant surface melt between April 29th and the UAV survey on June 6th. This is not important for the purpose of this study, which is to compare measurements derived from the glaciological method and SfM change detection over the same time interval. However, this means the balance estimates likely do not capture the full winter or summer balances.

Snow-bridge formation near crevasses and propagation of crevasses into new cells between surveys present an additional potential source of uncertainty; crevassed regions record as unreasonably high areas of accumulation, up to 33 m. To account for this, and to make the SfM data comparable to the glaciological method data (which assumes constant mass gain across crevasses when it extrapolates), I used the raster calculator in ArcGIS to truncate maximum accumulation at 8.5 m. I also truncated negative values at 0 m. I assess this issue further in the discussion.

4.2.2 SfM - Summer Ablation

The end of summer drone survey on October 6th, 2018 marked the end of the survey interval. Because of equipment malfunctions and poor weather conditions, Shean was unable to collect imagery for the upper mid-glacier section near and above stake 3. Thus, there was only DEM data to apply change detection across swaths 1 and 2. By differencing the June 6th, 2018 and October 6th, 2018 DEMs, I produced dDEMs, which quantifies the change in surface elevation across swaths 1 and 2 through the summer ablation season. As with the accumulation
season data, crevasses introduce a likely bias; combinations of ice flow and snow-bridge/sérac collapse record as unreasonably high zones of ablation in crevassed regions. To reduce this bias, the raster calculator in ArcGIS was used to truncate maximum elevation loss at -10 m. Negative values were truncated at 0 m. I address the justification for these limits in the discussion.

Equation 3 was used to calculate the volume change between surveys. Average changes in surface elevation ($\Delta z$) were extracted from swaths 1 and 2 to represent the average ablation across each swath. These were converted to water equivalent on the basis of the density of the material lost or gained, which I determined based on snow depths in dDEMW (Klug et al., 2018). I assumed the snow depth averages on dDEMW for swaths 1 and 2 estimate the total snow melt ($\Delta h^s_k$) between UAV surveys on June 6th and October 6th, 2018, and I attributed additional elevation loss between surveys to ice melt ($\Delta z^i_k$) (Figure 5). Snow and density assumptions were used to convert meters of surface melt to meters of water equivalent. The summer balance for swaths 1 and 2 were determined by summing these with the following expression.

$$b_{sSfM} = \sum_{k=1}^{2} [(0.5 \times \Delta z^s_k) + (0.9 \times \Delta z^i_k)] \times S$$

where $b_{sSfM}$ = summer mass balance estimated by SfM between surveys (m$^3$ w.e.), $k$ = swath number, $\Delta z^s_k$=average elevation change due to snow melt across dDEM$^s$ for swath k (m), $\Delta z^i_k$ = average elevation change due to ice melt across dDEM$^i$ for swath k (m), and $S$ = swath area (m$^2$).

5.0 Results

5.1 Glaciological Method

Between the stake installation date on May 20th and stake removal on September 17th, 2018, I measured -8.93 (-5.02 m snow and -3.91 m ice), 7.48 (-6.05 m snow and -1.43 m ice), and -5.43 (-5.4 m snow and 0 m ice) meters of surficial snow and/or ice loss at stakes 1, 2, and 3, respectively, following the traditional glaciological method (Figure 6). Based on density values for snow (0.5 ± 0.05 g/m$^3$) and ice (0.9 g/m$^3$) on North Cascades glaciers (Riedel and Larrabee, 2011), these values equate to -6.03 ± 0.3 m w.e. at stake 1, -4.31 ± 0.3 m w.e. at stake 2, and -2.71 ± 0.3 m w.e at stake 3 (Figure 6). These measurements have a high negative correlation ($R^2 = 0.995$) with stake elevation (Figure 6).

After using the PDD model to estimate June 6th snow depths from in situ measurements on the stake installation date (May 20th, 2018) to June 6th, I estimate 4.2 m at stake 1 and 5.47 m
at stake 2 (Table 1). These snow depths mark the beginning of the ablation season and I use them as a reference for my summer mass balance estimates.

Incorporating the PDD model changes the ablation values only slightly, to -9.03 m (-4.21 m snow, -4.83 m ice), -7.18 m (-5.89 m snow, -1.19 m ice), and -5.26 m (-5.26 m snow, 0 m ice) of surficial snow/ice loss and -6.45, -4.28, and -2.63 m w.e. at Stakes 1, 2, and 3 respectively between June 6th and October 6th, 2018 (Figure 6). As with the raw measurements, these values have perfect negative correlations with elevation \((R^2 = 1)\) (Figure 6). These differences are within the uncertainties of the PDD model.

Extrapolating the ablation estimates across the swath areas estimates apparent changes in volume: -766,664 m\(^3\) and -547,330 m\(^3\) w.e. across swath 1, -1,832,006 m\(^3\) and -1,091,184 m\(^3\) w.e. across swath 2, combined to a total volume loss of -2,598,669 m\(^3\) and -547,330 m\(^3\) w.e across swaths 1 and 2 (Figures 7 and 8).

5.1.1 Glaciological Method Error

As summarized in section 2.1.1, the glaciological method is subject to an assortment of random and systematic errors (Zemp et al., 2013; Beedle et al., 2014; Riedel and Larrabee, 2008). According to Zemp et al. (2013), there are three primary sources of error: 1) errors in field measurements at the ablation stakes, 2) the inability for a limited number of stakes to adequately capture the spatial variability of mass balance, and 3) changes in glacier hypsometry through time. For example, interstitial layers in the snow and decimeter-scale variability in the previous summer’s surface may introduce error in the snow depth measurements at the end of the accumulation season. Imprecision in stake height measurements, oblique probing to the snow/ice interface, and enhanced ablation around stakes or sinking of stakes all impart potential errors as well (Zemp et al., 2013; Beedle et al., 2014). Errors and uncertainties related to density measurements and/or assumptions must be considered too. Furthermore, there is uncertainty in the local representativeness of point measurements for the areas where they are extrapolated, and the potential for under-sampling of inaccessible glacier regions such as those with crevasses or steep slopes. The glaciological method relies on its altitudinal swaths to extrapolate stake measurements, but a glacier’s geometry may change over the course of a survey interval - requiring that swath extents be regularly redefined. Lastly, the glaciological method does not capture basal or internal mass balance (Klug et al., 2018).
Zemp et al. (2013) assumed that the three primary sources of error in the glaciological method are uncorrelated, and they used the law of error propagation to estimate a total error of ±0.34 m w.e. a⁻¹ on average for twelve glaciers monitored by the World Glacier Monitoring Service (Zemp et al., 2013). Similarly, Fountain and Vecchia (1999) approximated cumulative annual errors of ±0.1 to ±0.3 m w.e. a⁻¹, depending on the number of ablation stakes. Others estimated errors within the glaciological method ranging from ±0.2 to ±0.4 m w.e. a⁻¹ (Cogle and Adams, 1998 and Cox and March, 2004, as cited in Thibert et al. 2008). North Cascades National Park estimates error in their glaciological field measurements at each stake and on each glacier at an annual basis (Riedel et al., 2008). They estimated 0.22 m w.e. error for their summer balance, or 6.2% of the total summer balance (-3.62 m w.e.) on North Klawatti Glacier in 2002 (Riedel et al., 2008). This estimate closely matches the estimated measurement error of the other three NPS-monitored glaciers, which had an average error of ±0.29 between 1992 and 2010 (Riedel and Larrabee, 2016). Based on the consistency of these estimates in the literature, I assign a conservative average error of ±0.3 m w.e. for the summer ablation in my study. The NPS also estimates a winter balance error in their annual mass balance studies.

5.2 Structure-from-Motion Change Detection

In contrast to the high-precision, low spatial resolution of point measurements in the glaciological method, results from the SfM models provide a spatially distributed pattern of surface elevation gain during the accumulation season (October 2017-June 2018; dDEMW) and surface elevation loss during the ablation season (Figure 9). Values of surface elevation change in grid-cells across dDEMW (May 2018 DEM minus October 2017 DEM) are representative of winter snow accumulation, and surface elevation changes across dDEMs (October 2018 DEM minus June 2018 DEM) are representative of summer snow and ice ablation.

The average gains in surface elevation through the accumulation season (dDEMW) are 5.51 ± 1.23 m for swath 1 and 5.89 ± 1.08 m for swath 2, with an average of 5.79 ± 1.13 m for swaths 1 and 2. Conversely, there is greater elevation loss across swath 1 than swath 2 during the summer (dDEMs): -8.05 ± 0.94 m for swath 1 and -7.08 ± 0.95 m for swath 2 (-7.32 ± 1.04 m average). Both dDEMs indicate apparent enhanced accumulation and ablation in crevassed regions, up to 33 m and 24 m respectively.

Uncertainties related to ice flow between surveys, particularly emergence/submergence (e.g., Cuffey and Paterson, 2010), do not allow measurements of elevation change for a parcel of
ice from the dDEMs; unlike the glaciological method, which allows us to make measurements of the same point on the glacier because the stake flows with the ice in a Lagragian frame of reference. Instead, I calculate average changes in surface elevation across each swath. Extrapolating these averages across the swath areas yields apparent changes in swath volume. For the ablation interval, there is -686,600 m³ of snow and ice loss across swath 1, -1,805,700 m³ across swath 2, and a total volume loss of -2,489,300 m³ across swaths 1 and 2.

As with the glaciological method, accepted regional values of snow and ice density are used to convert volumes to mass estimates (Riedel and Larrabee, 2011). There was ice ablation in addition to seasonal snow melt at stakes 1 and 2. I infer that the total snow melt during the survey interval is equal to the June 6th snow depth, and any additional surface elevation loss relates to ice ablation. The results indicate mass loss of -5.04 m w.e. for swath 1, -4.02 m w.e. for swath 2, and an average of -4.27 m w.e. for combined swath 1 and 2 between June 6th and October 6th, 2018. These translate to a total of -1,452,400 m³ w.e. of snow and ice mass loss across swaths 1 and 2 (Table 1).

Although I cannot extract point measurements from the SfM model across the entire glacier, I compared point measurements on dDEMW at the stakes’ GNSS positions during three of our field visits (t = 1, 2, and 3) to swath averages. Point measurements along these trajectories remain within 3.7% for the swath 1 average and 10.7% for the swath 2 average (Stake 1: 7.85, 8.05, 7.75 and Stake 2: 6.32, 6.44, 6.34). This consistency suggests that because sites selected for the glaciological method have a relatively gentle (~10-15°), constant slope with no crevasses, they appear to represent the swath as a whole reasonably well.

5.2.1 SfM Error

Structure-from-Motion surveys incorporate the following sources of error: 1) insufficient contrast in the aerial imagery (especially in the spring survey, when there are fewer crevasses and no exposed ice), 2) low resolution/quality of the aerial photos (blurred, under/over-exposed), 3) inaccuracies in the horizontal and vertical measurements of GCPs due to GNSS Trimble imprecision, 4) co-registration uncertainties between sequential surveys Beedle et al., 2015; David Shean, pers. comm., 2019). Assumptions related to snow and ice density incorporate additional error.

A common way to evaluate the inherent systematic uncertainty in stereo models is to approximate the standard deviation of elevation change on “static”, stable bedrock features (Cox
and March, 2004; Beedle et al., 2015; Pelto et al., 2019; Shean et al., 2020). Because bedrock surfaces remain fixed in space through time, unlike points on the glacier, their geographic locations should not change between sequential surveys. Any discrepancy in their position can be attributed to residual error in the stereo models, and averaging the residuals can be used to correct elevation measurements on the glacier (Beedle et al., 2015; Pelto et al., 2019).

In a recent study, Shean et al. (2020) generated DEMs from commercial satellite stereo imagery (i.e., DigitalGlobe WorldView) to estimate glacier mass balance across High Mountain Asia. They used the standard deviation in stable bedrock features to estimate the background noise and error in their measurements of glacier elevation change on individual glaciers. They also incorporated error from uncertainty in the accuracy of their glacier polygon digitization, changes in glacier extent during the study interval, and signals related to ice dynamics. Their assumptions related to density incorporated the third error component (Shean et al., 2020). They assumed these errors were independent and uncorrelated and propagated them into the total mass balance error for each individual glacier in their study.

According to David Shean, the positions of the GCPs and the photograph positions (taken by the RTK on board the eBee) are all accurate to a few centimeters in my study. Additionally, because the snowpack had matured by June 6th, it had sufficient texture for feature matching, and the co-registration between the October 2017 and June 2018 DEMs (dDEMW) was excellent. The exposed crevasses and ice in late October facilitated an accurate co-registration for the June 2018 and October 2018 DEMs (dDEMs) as well. Surface elevations are accurate to <5-10 cm (David Shean, pers. comm., 2019).

5.3 Glaciological versus SfM Data

The average surface elevation change for each swath in the differenced DEMs is used to compare the stake measurements of surface melt measured by the glaciological method with elevation change measured by SfM change detection. Surface melt measured by in situ stake surveys and surface elevation change measured by SfM change detection are hereafter referred to as glacier thickness change.

The SfM swath averages estimate greater apparent June 6th snow depths at stakes 1 and 2 compared to the apparent snow depths estimated by the in situ measurements and PDD model via the glaciological method. Snow depth estimates are 23.7% (1.3 m) and 7.2% (0.4 m) higher for swaths 1 and 2 from SfM compared to glaciological method depths on June 6th (SfM average
elevation change on dDEMW is 5.51 ± 1.23 m and 5.89 ± 1.08 m for swaths 1 and 2; the glaciological method estimates 4.2 and 5.5 m at stakes 1 and 2). The snow depth at stake 2 falls within the standard deviation of the swath 1 average, but stake 2 does not.

Data from both methods reveal greater glacier thickness change, and thereby greater volume and mass loss, across swath 1 than swath 2. Average surface elevation change measured by the differenced DEMs is lower than the stake measurements of surface ablation (10.8% lower at stake 1, and 1.4% lower at stake 2; Figure 7). These differences increased in magnitude when converted to meters of water equivalent (21.8% lower at stake 1, and 6.1% lower at stake 2; Figure 7). Swath 2 estimates have smaller discrepancies between results yielded from glaciological and SfM data.

Extrapolating the measurements across each swath and combining them to estimate a total volume for swaths 1 and 2 indicates that total apparent volume in cubic meters estimated by SfM methods was 4.3% lower than glaciological methods. Likewise, the total apparent volume in cubic meters of water equivalent estimated by SfM is 11.4% lower than glaciological method measurements (Figure 7). These comparisons do not take into account surface elevation change from horizontal or vertical ice flow.

Table 1. June 6th snow depth, total thickness change, and total snow versus ice melt between June 6th and October 6th, 2018 at stakes 1 and 2 estimated by the glaciological method and the SfM swath averages. For the glaciological method, a positive degree-day model was used to estimate June 6th snow depths based on the in situ snow depths measured in the field on May 20th. For SfM, the average snow depth was extracted from swaths 1 and 2 on dDEMW. Any ablation that exceeds the June 6th snow depths is attributed to ice ablation (Figure 5).

<table>
<thead>
<tr>
<th>Stake</th>
<th>Glaciological method</th>
<th>SfM</th>
<th>Glaciological method</th>
<th>SfM</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Snow depth (06/06/18) (m)</td>
<td>Thickness change (06/06-10/06) (m)</td>
<td>Snow melt (m)</td>
<td>Ice melt (m)</td>
</tr>
<tr>
<td>Stake 1</td>
<td>4.2</td>
<td>9.03</td>
<td>4.2</td>
<td>4.83</td>
</tr>
<tr>
<td>SfM</td>
<td>5.51 (+23.8%)</td>
<td>8.05</td>
<td>5.51</td>
<td>2.54</td>
</tr>
<tr>
<td>Stake 2</td>
<td>5.47</td>
<td>7.18</td>
<td>5.47</td>
<td>1.72</td>
</tr>
<tr>
<td>SfM</td>
<td>5.89 (+7.7%)</td>
<td>7.08</td>
<td>5.89</td>
<td>1.19</td>
</tr>
</tbody>
</table>
6.0 Discussion

My results affirm there are advantages to and limitations of both the glaciological method and UAV-SfM techniques, and that SfM cannot yet replace the glaciological method without constraints on some confounding factors. Both methods record mass loss in volume of water across swaths 1 and 2 of the Easton Glacier between June 6th and October 6th, 2018, and reveal a higher total loss across swath 1 compared to swath 2. Both methods also yield a strong correlation between ablation and altitude during the study period. Several variables must be addressed and incorporated into the SfM mass balance measurements in order to make an accurate comparison of the two methods.

Mass balances estimated by the two methods cannot be directly compared without correcting for: 1) different time intervals between the UAV surveys and in situ ablation stake measurements, and 2) the vertical component of ice flow in the SfM models (i.e., submergence/emergence; Cox and March, 2004; Sold et al., 2013; Beedle et al., 2014; Belart et al., 2017). Additionally, because of the Eulerian frame of reference of the geodetic method, the differenced DEMs cannot yield point measurements of thickness change at the same point on the glacier without accounting for horizontal (down-glacier) ice flux. In the following sections, I discuss these variables and their implications for measuring mass balance, my attempts to account for them, and possible considerations for future studies. I also evaluate the uncertainties in my study from difficulties with measuring mass balance in crevasses.

Lastly, my study did not include the accumulation area, where firn densification may result in overestimates of thickness change during the ablation season as snow at the snow-ice interface compacts and evolves into ice (Belart et al., 2017). For example, SfM models could record changes in glacier volume without any mass change solely from firn densification (Huss, 2013). Studies that survey the accumulation area should consider the impact of firn densification on SfM-derived mass balance totals (Sold et al., 2013; Belart et al., 2017).

6.1 Horizontal Ice Flux

Horizontal ice flux displaces points on the glacier surface through time, so SfM differencing cannot yield a measurement of true accumulation or true ablation on a single parcel of ice (Beedle et al. 2014). In some places, stakes moved 14 meters down-glacier between SfM surveys. Since small-scale spatial variability in accumulation and ablation is significant, I cannot assume thickness change at both positions is the same. Additionally, the propagation of surface
features like crevasses or séracs has a significant effect on surface elevation changes between SfM surveys. I address this problem in more detail below. In order to measure true surface elevation change at the same point on the glacier, the second DEM would need to be shifted backwards along flow trajectories to ensure the same point on the glacier surface is being compared. Flow vectors vary spatially with ice thickness and surface slope, so this method would require glacier-wide analysis of horizontal surface ice flux (Whitehead et al., 2013).

The problem of flow vector variability can be mitigated substantially by using SfM swath averages. Although portions of the Easton glacier surface from swath 2 flow into swath 1 throughout the ablation season, this advective movement of points is negligible when extracting my swath averages because of the scale of the swaths. In their study on Castle Creek Glacier, Beedle et al. (2014) did not attempt to quantify the errors introduced by advection of topography, but they suggest that this omission may increase the geodetic errors of surface mass balance.

6.2 Vertical Ice Flux

Some of the discrepancy between the glaciological and SfM data relates to difficulties in estimating vertical ice flow. Vertical changes indicated by differencing SfM surveys reflect a combination of accumulation/ablation and submergence/emergence due to ice dynamics; thus, mass balance cannot be fully quantified without also constraining submergence/emergence (Figure 10). Because emergence dominates in the ablation zone, differenced SfM models should overestimate surface elevation change in the ablation zone during the accumulation season, and underestimate surface elevation change in the ablation zone during the ablation season.

Emergence effects are evidenced in my SfM results. Emergence during the 8-month interval between fall and spring SfM surveys (the winter accumulation season) likely explains why SfM swath averages estimate greater apparent snow depths on June 6th compared to glaciological method in situ depths that were adjusted to the same date (Figure 10). Likewise, emergence during the ablation season probably explains the substantially greater glacier thickness change estimated by SfM (dDEM5) compared to the glaciological method between June and October surveys (Figures 7 and 8).

Field measurements near the stakes record lower June 6th snow depths than SfM models, and so they estimate a higher proportion of ice to snow melt compared to SfM models (Figure 5). Because additional ablation that exceeds the June 6th snow depth is characterized as higher
density ice melt, the discrepancy in summer ablation data is exacerbated when I adjust the total volumes for density (Figures 5, 7, and 8).

Additionally, discrepancies in the data for both the apparent June 6th snow depth and glacier thickness change during the ablation season were substantially greater for swath 1 (for example, SfM models record 23.8% higher snow depth on June 6th than the glaciological method as compared to 7.7% for stake 2) (Table 1). I attribute this to higher magnitudes of emergence across swath 1; greater magnitudes of ablation increase horizontal and vertical flow velocities near the glacier terminus, and cause emergence velocities to decrease as altitude increases (Gudmundsson and Bauder, 1999).

6.2.1. Emergence Velocity Estimates

To address the influence of emergence velocity on the surface elevation changes, I estimate emergence at stakes 1 and 2 between June 6th and October 6th by combining data from my in situ stake measurements, GNSS positions, and SfM DEMs (Figures 12 and 13). Because I have field data on May 20th and September 17th, I calculated total emergence between these two dates, determined the average daily emergence during that interval, and adjusted the rate to span the June 6th – October 6th study interval.

To estimate total emergence between May 20th and September 17th, I compared the in situ GNSS elevation of the ice surface at the stake on September 17th with a projected elevation based solely on down-glacier flow (i.e., zero emergence and zero ice melt). The ice surface slope, which I extracted from the October DEM, is used to determine the projected elevation. The difference in these elevations is attributed to the magnitude of emergence. I illustrate the process for stake 1 in Figures 12 and 13 and I simplify the concept by using data from May 20th to June 17th, when there isn’t any ice melt. To isolate the influence of ice flow (excluding surface melt), and because the ice surface is most representative of the glacier’s surface motion, I focus my calculations at the snow/ice interface (Klug et al., 2018).

Based on GNSS X-Y position coordinates of the stakes recorded during the May 20th (T1) and September 17th, 2018 (T2) field visits, I calculated the horizontal distance traveled between surveys (x). I also recorded the GNSS surface elevations of the snow at stakes T1 and T2 during these field visits. From these elevations, I subtracted the field measured May 20th snow depth (5.0 m) to estimate the elevation of the snow/ice interface at P1 on that date (1,695.9 m asl).
I then projected the May 20th (P1) and September 17th (P2) GNSS X-Y position coordinates onto the October 6th, 2018 SfM DEM to obtain their elevations and estimate the surface slope of the ice between stake positions. Because it was mostly snow free in the ablation zone in October, I assume the surface slope on the October DEM more accurately reflects the actual ice surface slope (and thereby, the driving stress of the glacier) than does the snow-covered June DEM. I also assume that the surface slope remains relatively constant along this transect through the summer.

Assuming surface-parallel flow, I then determined what the change in surface elevation of the snow/ice interface would be at P2 on T2, if there were no emergence. Given the DEM-determined ice-surface slope (α) and the GNSS-determined horizontal distance (x) traveled down-glacier between the two field visits, I estimated an expected elevation of the snow/ice interface on September 17th (T2), disregarding any influence of emergence. To determine the actual elevation of the ice surface on September 17th, I have to account for the ice melt between the dates by adding 3.9 m of ice melt to the GNSS elevation on September 17th. This elevation represents the ice surface elevation if there were no ice melt and no emergence. I attribute the difference in my expected and the actual elevation of the snow/ice interface at P2 to the magnitude of emergence between May 20th and September 17th.

I calculated the average rate of emergence per day between May 20th and September 17th, and extrapolate it forward and backward to estimate the total emergence during the study interval - June 6th to October 6th, 2018. Because differencing DEMs spanning the ablation season underestimates elevation change related to emergence, I added the emergence estimates from the above calculations to the swath averages in dDEMs and thereby increase the net surface elevation change between June 6th and October 6th.

After this adjustment for emergence, the SfM estimates are higher than the glaciological measurements (Figures 14 and 15). For example, with the emergence estimate corrections, SfM methods estimate 13.3% higher volume change in cubic meters and 13.6% higher volume change in cubic meters of water equivalent compared to the glaciological measurements, as compared to SfM estimates that are 4.2% and 11.3% lower than glaciological measurements before emergence corrections (Figures 14 and 15). Between June 6th and October 6th, 2018, the calculated emergence at stake 1 (2.88 m) is greater than at stake 2 (0.82 m), due to higher emergence velocities near the terminus.
Although these emergence estimates provide a means to constrain the influence of vertical ice flow on surface elevation change, the calculations are sensitive to small changes in inputs (e.g., surface slope, horizontal distance traveled, and uncertainties in GNSS-derived and DEM-derived positions). Additionally, the estimates do not account for varying velocities across each swath. They are based on measurements located along the centerline of the glacier, where horizontal and vertical flow are typically highest (Whitehead et al., 2013). Thus, these estimates likely provide an upper limit for emergence, and thereby overestimate total volume loss when applied to the SfM data (Figures 14 and 15). Floating ablation stakes could have confounded my emergence calculations, though I perforated the lower ends of each stake to limit this (Riedel et al., 2008). I offer recommendations for future studies to more accurately quantify vertical velocities in order to extract robust ablation values from SfM dDEMs.

Lastly, this problem may be an even greater issue for making accumulation-season estimates from SfM DEMs; slower flow velocities during the winter accumulation season are likely counterbalanced by the longer time span between SfM surveys. Without accounting for emergence during the accumulation season, the SfM-derived June 6th snow depths likely overestimate the actual snow depths, introducing additional unquantified error.

6.2.2. Previous Efforts to Estimate Emergence Velocities

Studies have considered other means to account for the influence of vertical ice flow when attempting to measure volume change or mass balance from differenced DEMs. Nolan et al. (2015) suggested making two DEMs during an interval of time where there is no melt or snowfall, then all changes in surface volume can be attributed to vertical ice flow, and the magnitude of flow at points across the glacier surface can be quantified. This quantity can then be subtracted from differenced maps, and the resulting surface elevations should just represent changes in surface mass balance (Nolan et al., 2015). However, this technique relies on finding an interval of time when there is no surface ablation or accumulation, which is impossible.

Other mass balance studies have adjusted for emergence by estimating its magnitude with field measurement or models (e.g., Rasmussen and Wenger, 2009; Sold et al., 2013; Beedle et al., 2014; Belart et al., 2017) while others have ignored its influence altogether (e.g., Pelto and Menounos, 2019). Here I summarize their attempts, as well as the associated assumptions and uncertainties. I also highlight the elements of these papers that seem most relevant to SfM change detection in the future.
Emergence can also be estimated with three-dimensional flow modeling. The full-Stokes ice flow model, for example, inputs a bedrock DEM, surface DEM, and in situ GPS velocities to calculate a 3-D velocity field (Belart et al., 2017). There are limitations to this method because of challenges with generating a bedrock DEM. Belart et al. (2017) estimated the winter balance of an Icelandic ice cap by differencing DEMs generated from satellite stereo-images, and by taking field measurements with the in situ glaciological method. They found agreement in their mass balance estimates after using the full-stokes model to correct for vertical ice flow and after making adjustments for firn densification and the time difference between the satellite and in situ surveys.

Hamilton and Whillans (2000) installed survey markers tens of meters into firn on the Greenland Ice Sheet. The markers were connected to steel cables reaching the surface, which allowed them to record horizontal and vertical positions of the markers at the surface through repeated GPS surveys. After correcting for horizontal along-slope motion and firn densification, they derived vertical velocities within uncertainties of ~0.01 m (one-sigma). To further decrease uncertainty, they took measurements at several sites and with five markers installed at each site. In a mass balance study at the Castle Creek Glacier, Beedle et al. (2014) compared two methods of estimating vertical velocity: one, using a Lagrangian frame of reference and transient ablation stakes and a second using an Eulerian frame of reference with fixed GPS measurements. In the Lagrangian frame of reference, they combined glaciological and GPS measurements on a network of ablation stakes to measure surface ablation between subsequent visits with ice velocity oriented along the flow. The glacier surface slope was derived from a DEM. They used the following equation to estimate vertical velocity:

$$w_s = b + \mu s \tan \alpha$$

where $w_s$ = vertical velocity, $b$ = surface ablation at stake moving with ice (from in situ stake measurements), $\mu s$ = ice velocity oriented along the flow (from GPS), and $\alpha$ = slope of glacier surface (from DEM).

In the Eulerian frame of reference, they measured surface ablation on the same array of ablation stakes. They used GPS measurements to estimate thickness change on a series of fixed points which were located at the initial position of each stake. They ignored advective movement by assuming that ablation at a transient stake doesn’t significantly change between surveys, even
though their stakes traveled 5-20 meters down-glacier through the course of their study. They then used the following equation to estimate vertical velocity:

\[ w_s = h - \frac{b}{\rho} \]

where \( w_s \) = vertical velocity, \( b \) = surface ablation at stake moving with ice (from \textit{in situ} stake measurements), \( h \) = thickness change (from GPS on fixed coordinates), and \( \rho \) = density of surface material.

Using the method in the Eulerian frame of reference they estimated higher emergence velocities and attribute the discrepancy to advection of surface topography (Beedle et al., 2014). In addition, neither estimate takes into account the influence of firn densification, basal and internal mass balance, isostatic displacement, and erosion of the bed surface. Lastly, their methods are field-intensive and do not significantly reduce the logistics compared to the glaciological method.

6.2.3. Considerations for Future Studies

I recommend that future studies build on the techniques employed by Beedle et al. (2014) to constrain emergence velocities in the ablation zone. Using equation 7 to take measurements in the Lagrangian frame of reference avoids errors introduced by horizontal ice flux. Installing an array of stakes across swaths can account for spatial variability in emergence and reduce uncertainty. Concentrating stakes along the centerline of the glacier, where flow is fastest, as well as along the sides of each swath would most adequately capture the high latitudinal variability in emergence velocity.

Estimates can then be averaged across each swath, and added to the elevation change resolved by the differenced SfM models. While these techniques are field-intensive at the front end, emergence velocities remain relatively constant for multiple mass balance years, and measurements would not have to be taken annually (Sold et al., 2013). Additionally, Belart et al. (2017) assumed that the slower emergence velocities during the winter season are negligible, but I suggest future studies constrain emergence during the winter accumulation season before accepting this assumption.

Alternatively, future studies could estimate the emergence velocity by combining measurements of ice thickness with surface flow vectors (Immerzeel et al., 2014). Ice thickness can be derived from ground penetrating radar measurements, and surface velocity can be obtained from feature tracking of imagery (Immerzeel et al., 2014; pers. comm. David Shean,
Constraints on the proportion of basal sliding vs. internal deformation for observed surface velocities are necessary - making it a currently unsolved challenge (pers. comm., David Shean, 2019).

**6.3 Mass Balance in Crevasses**

Crevasses introduce a potentially large source of uncertainty for SfM surveys from two sources of noise: 1) snow bridging between surveys, and 2) advection of crevasses into other grid cells on subsequent DEMs. These problems are more significant on glaciers where crevasses make up a large proportion of the surface area. For example, 2015 lidar imagery suggest that heavily crevassed zones (i.e. ice falls) represent as much as 50% of the surface area on some glaciers on Mount Baker. As a result, the differenced DEMs record significantly greater surface elevation change in crevassed regions of my study area. While some of the variability in surface elevation may be due to wind redistribution of snow on the glacier surface, I attribute the larger anomalies to crevasses.

As crevasses get bridged with snow through the winter, straight-differenced DEMs show that they are entirely filled with snow and thus overestimate the amount of snow accumulation in those regions. Crevasses with depths of 20 meters in October record an anomalously high total accumulation of 20+ meters in June when the crevasse was presumably only bridged with 8-9 meters of snow accumulation. This could also be explained by the advection of crevasses into new grid cells: DEM pixels with crevasses that are tens of meters deep in October may have non-crevassed glacier surface flow into them through the winter season, thereby recording >20 meters of accumulation.

Few studies have examined surface mass balance in crevasses, and it is not fully understood. Crevasses likely do get some excess snow accumulation compared to non-crevassed surfaces because of a minimum depth that is needed for snow bridges to form, so a simple interpolation across them likely underestimates true accumulation. Additionally, studies suggest that heat capture in crevasses and/or the increased albedo of glacier ice exposed to the sun in crevasses may enhance ablation by ~15% (Colgan et al., 2016). Others suggest decreased ablation in crevasses due to shading from crevasse walls (Krimmel, 1999).

To remedy these uncertainties, I looked for maximum surface elevation change values in adjacent, non-crevassed cells and truncate the differenced data at upper limits in each dDEM. For example, few cells outside of crevassed regions exceeded 8.5 meters of snow accumulation in
dDEMw so I use the conditional tool in ArcGIS to truncate the raster at 8.5 meters so that any cells greater than 8.5 were set to 8.5. Using the same assessment for the ablation dDEMs, I truncated the differenced data at 10 meters for both swaths.

Since it was difficult to precisely determine the highest ablation in cells adjacent to but not in crevasses, I test the sensitivity of changing the truncation upper limit. The truncation limits do not dramatically change the results for swath 1, swath 2, and combined swath 1 and 2 averages (Table 1). For example, decreasing the upper limit from 10 to 9 m for the ablation dDEMs decreases the average surface elevation by -1.45% for swath 1, -0.08% for swath 2, and -0.1% for swaths 1 and 2. Increasing the upper limit from 10 to 11 m increases the average surface elevation change by 0.08% for swath 1, 0.06% for swath 2, and 0.06% for swaths 1 and 2.

These modifications were corroborated with a simple analysis of histogram data from each swath: 7.15% (swath 1) and 3.78% (swath 2) of cells had accumulation values greater than 8.5 m (Figure 16). I also observe that 8.7% (swath 1) and 0.09% (swath 2) of cells record ablation values greater than 10 m. I infer that the tallest bin in each histogram represents the dominant smooth, non-crevassed surface across each swath, and I observe that the swath averages fall within this mode.

In addition, some cells record apparent mass loss in dDEMw. This reflects that although crevasse fields are stable year to year, individual crevasses and séracs are not. For example, when a large sérac or unbridged crevasse changes position through the winter, a point that occupied a sérac in October may be crevassed in June, resulting in a negative change in absolute surface elevation at that point. Because this is an inaccurate record of accumulation, I set all points that record negative accumulations to 0 m for both dDEMs. The actual accumulation at these points is almost certainly greater than 0 m, but this truncation eliminates data that is identifiably incorrect and significantly reduces error.

Previous studies masked crevasses because of unreliable data in these regions, which seems reasonable for glaciers with few crevasses (Sold et al., 2013; Pelto et al., 2019). Crevasses make up a large proportion of the study area and I determine that nulling them would substantially skew the data. For example, setting all cells with values greater than 10 m to null decreases swath 1’s ablation average by 32.6%. I conclude that this analysis would benefit from further research into mass balance processes in crevasses.
Table 2. Results of sensitivity test on ablation dDEM (dDEMs) to address bridging on crevasses.

<table>
<thead>
<tr>
<th>Truncation Range</th>
<th>Swath 1 Average (m)</th>
<th>Swath 2 Average (m)</th>
<th>Swath 1 and 2 Average (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-9 m</td>
<td>7.935 (-1.45%)</td>
<td>7.021 (-0.08%)</td>
<td>7.25 (-0.1%)</td>
</tr>
<tr>
<td>0-10 m</td>
<td>8.052</td>
<td>7.08</td>
<td>7.323</td>
</tr>
<tr>
<td>0-11 m</td>
<td>8.117 (+0.08%)</td>
<td>7.122 (+0.06%)</td>
<td>7.369 (+0.06%)</td>
</tr>
<tr>
<td>No truncation</td>
<td>7.19 (-10.7%)</td>
<td>8.22 (+16.1%)</td>
<td>7.45 (1.7%)</td>
</tr>
</tbody>
</table>

6.4 Validating mass balance measurements with the hydrological method

Previous studies were not able to measure mass balance using the hydrological method because of significant error within run-off estimates (e.g., Tangborn et al., 1975). Given the limitations of both methods, future studies could attempt to validate mass balance by quantifying meltwater discharge at the Easton Glacier. Such measurements, however, are difficult to obtain for this glacier and there is currently no USGS stream gauge downstream of the Easton between its terminus and its inflow into Baker Lake. I attempted to monitor stream discharge by installing stream gauges in the Easton’s two main meltwater streams through summer 2018 (Appendix II). Because the drainage network near the terminus is highly dynamic, a high discharge event or debris flow destroyed the equipment. I was also unable to yield a statistically significant relationship between stream stage and discharge with a rating curve – likely due to aggradation and erosion near the stream gauges, which decreased the accuracy of the stage readings. This could be attempted again in a better constrained and less dynamic stretch of the meltwater creek. For example, the Water Resources Program with the Nooksack Indian Tribe has modeled continuous discharge through their stream monitoring efforts in Sholes Glacier Creek on Mount Baker. Sholes Creek is a more ideal site for streamflow measurements: there are few debris flows, the flow is relatively steady and uniform, and the channel is straight and confined (Jezra Beaulieu, pers. comm., 2017).

7.0 Conclusions and Summary

Structure-from-Motion methods have the potential to substantially increase the number of glaciers that can be monitored globally. Because of their high resolution and spatial extent, SfM models provide elevation data equivalent to what would be yielded from millions of ablation stakes (Immerzeel
et al., 2014). Additionally, by generating elevation models from UAV-derived imagery, they can collect data from otherwise inaccessible regions and without the substantial fieldwork associated with the glaciological method.

The challenges with adjusting for secondary processes that affect surface elevation (e.g., horizontal and vertical ice flux, firn compaction) must be more thoroughly addressed and integrated before UAV-SfM techniques can altogether replace the glaciological method. For now, estimating emergence requires field measurements at several ablation stakes organized in arrays across each swath. However, if velocities and therefore emergence remain relatively constant over roughly 5-year time scales, measurements of emergence would similarly only be needed on a several-year time frame (e.g., Belart et al., 2017).

The uncertainties related to density assumptions and with mass balance in crevassed regions necessitate more rigorous analysis as well. This study could have been improved by more closely aligning the dates of the SfM surveys and stake visits, and eliminating the error incorporated by the positive degree-day model.

SfM change detection requires fieldwork on the front end because GCPs are needed to validate the SfM models. However, because they can be placed in accessible regions of the glacier and on moraines, GCP placement demands substantially less time-intensive fieldwork than installing ablation stakes. Additionally, fieldwork will not be necessary when GCPs already exist (Beedle et al., 2014).

There is potential for WorldView satellite imagery to complete the gap in elevation data across the accumulation zone of the Easton glacier (David Shean, pers. comm., 2019). Because emergence and submergence reflect a distribution of mass along the entire glacier, they accrue to net zero vertical elevation change when totaled across the glacier (Whitehead et al., 2010). Thus, having elevation data for the entire glacier surface would eliminate the need to account for emergence when differencing the DEMs (Sold et al., 2013). However, studies that include the accumulation zone would have to address the effect of firn densification on surface height change - potentially a bigger challenge than constraining emergence. My findings suggest that future studies continue to implement SfM surveys alongside the glaciological method to further understand each method’s advantages and constrain these limitations.

The UAV-SfM method to measure glacier volume change and mass balance is only beginning to be tested in the field of glaciology, and this study demonstrates its potential as a
higher resolution and more widely applicable alternative to the glaciological method. Its spatially distributed model of surface elevation change improves upon the extrapolated point measurements of the glaciological method and the coarse resolution of geodetic techniques, and its remote sensing capabilities eliminate the rigorous fieldwork of the glaciological method.

8.0 References Cited


Whitehead, K., Moorman, B., and Wainstein, P., 2010, The Use of Ground-Based Photogrammetry and SAR Interferometry to Characterize the Surface Motion of an Arctic:

Figure 1. Location of the Easton Glacier on Mount Baker in northwest Washington.
Figure 2. Location of the five ablation stakes on the Easton glacier on Mount Baker. The study area encompasses the portion of the ablation zone highlighted by swaths 1 and 2.
Figure 3. Diagram showing the flow from raw glaciological method data and SfM data to ablation volume across swath 1.
Figure 4. (A) Highly significant correlation between daily average temperatures recorded at the Middle Fork Nooksack Snotel station and the weather station on the Metcalfe Moraine between July 20th and October 5th, 2018, (B) Weather station deployed on the Metcalfe Moraine adjacent to the Easton Glacier
Figure 5. Accumulation and ablation* estimated by the glaciological and SfM methods. Diagram highlights the different June 6th snow depths and their influence on the amount of snow versus ice melt during the ablation season. The glaciological method estimates a lower June 6th snow depth and a larger percentage of total ablation is thus attributed to higher density ice melt. This increases the discrepancy between the methods’ estimates of ablation: despite only 10.8% greater surface elevation loss estimated by the glaciological method compared to SfM, the glaciological method estimates 21.8% more volume loss in meters of water equivalent. This diagram does not include the influence of emergence.

*Reference figure 3 for further detail on determining ablation estimates for each method.
Figure 6. (A) Surface ablation versus elevation for stakes 1, 2, and 3 between May 20th and September 17th, 2018, based on in-situ measurements. Blue dashed line shows surface ablation in meters, and orange dashed line shows surface ablation converted to meters of water equivalent; (B) Surface ablation versus elevation for stakes 1, 2, and 3 between June 6th and October 6th, 2018 after using a positive degree-day model to adjust measurements.
Figure 7. (A) Ablation estimated by the glaciological method and positive degree-day model at stakes 1 and 2 compared to surface elevation loss estimated by the SfM averages for swaths 1 and 2 between June 6th and October 6th, 2018. (B) Water equivalent loss estimated by the glaciological method and positive degree-day model at stakes 1 and 2 compared to water equivalent loss estimated by the SfM averages for swaths 1 and 2 between June 6th and October 6th, 2018. Values are corrected for density.

Figure 8. (A) Ablation volume (m³) across swaths 1 and 2 between June 6th and October 6th, 2018, as measured by glaciological method and SfM swath averages. Values are not adjusted for snow or ice density. (B) Ablation volume corrected for density (m³ w.e.) across swaths 1 and 2 between June 6th and October 6th, 2018, as measured by the glaciological method and SfM.
Figure 9. (continued on next page) Elevation change across swaths 1 and 2 of the Easton Glacier between UAV surveys on October 5th, 2017 and June 6th, 2018 (dDEM<sub>W</sub>), and June 6th, 2018 and October 6th, 2018 (dDEM<sub>S</sub>). A. Hillshade generated from 10/05/17 UAV survey. B. Hillshade generated from 06/06/18 UAV survey. C. 06/06/18 DEM minus 10/05/17 DEM, used to represent winter accumulation. Values are truncated to 0 and 8.5 m to account for unreliable data in crevassed regions. D. Hillshade generated from 06/06/18 UAV survey. E. Hillshade generated from 10/06/18 UAV survey. F. 06/06/18 DEM minus 10/06/18 DEM, used to represent summer ablation. Values are truncated at 0 and 10 m.
A. October 5th, 2017

B. June 6th, 2018

C. Elevation Gain:
October to June (dDEMw)

D. June 6th, 2018

E. October 6th, 2018

F. Elevation Loss:
June to October (dDEMs)
Schematic showing the different factors that affect surface elevation in the ablation zone during the three UAV survey dates. Blue is snow fallen in winter, grey is preexisting glacier ice and firn, orange is ice ablation in summer, and green is emergence (dh\textsuperscript{WE}: surface elevation change from winter emergence and dh\textsuperscript{SE}: surface elevation change from summer emergence). SfM-UAV methods overestimate accumulation in the winter and underestimate ablation in the summer because of the Eulerian frame of reference and the influence of emergence. Figure adapted from Belart et al., 2017.

Schematic showing snow and ice melt relative to ablation stake 1 between its installation on May 20th, 2018 and the final visit on September 17th, 2018. Emergence doesn’t affect stake measurements taken with the glaciological method because of the LaGrangian frame of reference. Sections of the stake were removed over the course of the summer, but the entire stake length is shown here for clarity. With an initial snow depth of 5.00 m on May 18th, and 8.93 m of ablation relative to the stake through Sept. 17th, I infer that 3.91 meters of glacial ice melted.
Δx: Horizontal component of stake movement from down-glacier flow relative to bedrock
Δy: Vertical component of stake movement from down-glacier flow relative to bedrock
µ: Vertical emergence
Δh: Net change in surface elevation from emergence and surface melt, relative to the glacier ice surface
α: Surface slope of the snow/ice interface yielded from the October DEM
A.: Elevation of snow/ice interface at T_{G1}, estimated by subtracting the in situ snow depth from the GNSS surface elevation
B.: Actual elevation of snow/ice interface at T_{G2}, estimated by subtracting the in situ snow depth from the GNSS surface elevation
C.: Expected elevation of the snow/ice interface, calculated from α and Δx; without emergence (see below)

Figure 12. Sketch of the horizontal and vertical movement of ablation stake 1 and emergence of the snow/ice interface between T_{G1} and T_{G2}. Sketch highlights the expected elevation of the snow/ice interface at T_{G2} without emergence, and the actual elevation of the snow/ice interface, given emergence. Figure adapted from MIT OCW.

Figure 13. Expected elevation of the snow/ice interface at T_{G2} given Δx•tan(α) = Δh, when horizontal stake movement relative to the bedrock Δx = 2.1 m, and surface slope α= 11.8°.
Figure 14. (A) Ablation estimated by the glaciological method and degree-day model at stakes 1 and 2 compared to ablation estimated by the SfM averages for swaths 1 and 2 between June 6th and October 6th, 2018, after applying emergence adjustments. (B) Total ablation estimated by the glaciological method and degree-day model at stakes 1 and 2 compared to total surface elevation change estimated by the SfM averages for swaths 1 and 2 between June 6th and October 6th, 2018. Values are corrected for density.

Figure 15. (A) Total ablation volume across swaths 1 and 2 between June 6th and October 6th, 2018, as measured by the glaciological method and SfM swath averages with and without emergence adjustments. Values are not adjusted for snow or ice density. (B) Total ablation mass across swaths 1 and 2 between June 6th and October 6th, 2018, as measured by the glaciological method and SfM before and after emergence adjustments. Values are corrected for density.
Figure 16. (A) Histogram distribution of surface elevation change (m) across swaths 1 and 2 between October and June 6th, 2018, before truncating the lower and upper limits. Grey columns indicate cells that were set equal to 8.5 m to account for unreliable data in crevasses. (B) Histogram distribution of surface elevation change (m) across swaths 1 and 2 between June 6th and October 6th, 2018, before truncating the lower and upper limits. Grey columns indicate cells that were set equal to 10 m to account for unreliable data in crevasses.
Appendix I.

I conducted the glaciological method on three ablation stakes installed along a longitudinal transect up the lower Easton Glacier. Using a Trimble GeoXH, I recorded the GNSS position and elevation of each stake on the installation date on May 20\textsuperscript{th}, 2018 (Table 1). I also probed to the ice (stakes 1 and 2) and firn (stake 3) surface to estimate snow depths on the installation date (Table 1). I returned to each stake throughout the summer and measured the depth of snow or ice melt relative to each stake (Table 2). Weather and wildfire smoke impeded me from visiting each stake on every field visit. The dates of my glaciological measurements did not align with the UAV surveys, so I used a positive degree day model to adjust the glaciological measurements span the same interval of time. With the PDD model, I estimated the amount of snow melt between stake installation date on May 20\textsuperscript{th} and the first UAV survey, on June 6\textsuperscript{th}. I also estimated the amount of snow and ice melt that occurred between my final stake visit on September 17\textsuperscript{th}, and the second UAV survey on October 6\textsuperscript{th} (Tables 3 and 4).

**Table 1.** Position and elevation data recorded with the GNSS Trimble in the field for stakes 1, 2, and 3 on May 20\textsuperscript{th}, 2018.

<table>
<thead>
<tr>
<th>GNSS Northing</th>
<th>GNSS Easting</th>
<th>Elevation (m)</th>
<th>05/20 in situ snow depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stake 1</td>
<td>5398875.19</td>
<td>585468.64</td>
<td>1703.35 ± 0.1</td>
</tr>
<tr>
<td>Stake 2</td>
<td>5399385.75</td>
<td>585561.83</td>
<td>1865.43 ± 0.2</td>
</tr>
<tr>
<td>Stake 3</td>
<td>5399888.74</td>
<td>585786.43</td>
<td>1998.38 ± 1.3</td>
</tr>
</tbody>
</table>

**Table 2.** Surface snow and ice ablation measured relative to ablation stakes during field visits on June 17\textsuperscript{th}, July 24\textsuperscript{th}, and September 17\textsuperscript{th}, 2018. Dashes indicate dates we could not access the stake site.

<table>
<thead>
<tr>
<th></th>
<th>05/20/18 – 06/17/18</th>
<th>05/20/18 – 07/24/18</th>
<th>05/20/18 – 09/17/18</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stake 1</td>
<td>1.28 m</td>
<td>-</td>
<td>8.93 m</td>
</tr>
<tr>
<td>Stake 2</td>
<td>-</td>
<td>3.86 m</td>
<td>7.48 m</td>
</tr>
<tr>
<td>Stake 3</td>
<td>0.97 m</td>
<td>1.86 m</td>
<td>5.43 m</td>
</tr>
</tbody>
</table>
Table 3. Positive degree-day model calculations used to estimate total snow ablation at stake 1 between 04/20/18 and 05/20/18. The degree day factor for snow \((K_S)\) is calculated from snow melt measured between 05/21 and 06/17 and the number of positive degree days \((PDD_s)\) occurring at stake 1 during this time interval. The established \(K_S\) \((0.009 \text{ m}^\circ\text{C})\) is inserted into equation 2 to calculate the total snow melt for 04/29-05/20 and 06/21-06/06, given 148.23 positive degree days. This total, \(M\), estimates the total snow ablation between the first stake measurement (04/29) and the UAV survey (06/06).

<table>
<thead>
<tr>
<th>Dates</th>
<th>04/29-05/20</th>
<th>05/21-06/17</th>
<th>05/21-06/06</th>
</tr>
</thead>
<tbody>
<tr>
<td>Positive degree days (\circ\text{C})</td>
<td>159.798</td>
<td>148.234</td>
<td>95.204</td>
</tr>
<tr>
<td>Surface material</td>
<td>snow</td>
<td>snow</td>
<td>snow</td>
</tr>
<tr>
<td>In situ melt (m)</td>
<td></td>
<td>1.275</td>
<td></td>
</tr>
<tr>
<td>In situ melt (m w.e.)</td>
<td></td>
<td>0.638</td>
<td></td>
</tr>
<tr>
<td>DDF ((K_S; \text{m}/\circ\text{C}))</td>
<td></td>
<td></td>
<td>0.009</td>
</tr>
<tr>
<td>DDF ((K_S; \text{m w.e.}/\circ\text{C}))</td>
<td></td>
<td></td>
<td>0.004</td>
</tr>
<tr>
<td>Extrapolated melt ((M; \text{m}))</td>
<td>159.79 * 0.009 =</td>
<td>95.20 * 0.009 =</td>
<td></td>
</tr>
<tr>
<td></td>
<td>1.374</td>
<td>0.819</td>
<td></td>
</tr>
<tr>
<td>Extrapolated melt ((M; \text{m w.e.}))</td>
<td>0.687</td>
<td>0.409</td>
<td></td>
</tr>
</tbody>
</table>
Table 4. Positive degree-day model calculations used to estimate total ice ablation at stake 1 between 07/15/18 and 10/06/18. I used aerial imagery of snow coverage to estimate the date at which ice began to melt at stake 1 (07/15). The degree day factor for ice (KI) is calculated from ice melt measured between 07/15 and 09/17 and the number of positive degree days (PDDi) occurring at stake 1 during this time interval. The established KI (0.005 m/°C) is inserted into equation 2 to calculate the total ice melt for 07/15-10/06, given 876.214 positive degree days. This total, M, estimates the total ice ablation between the last stake measurement (09/17) and the final UAV survey (10/06).

<table>
<thead>
<tr>
<th>Dates</th>
<th>06/18-07/14</th>
<th>07/15-9/17</th>
<th>07/15-10/06</th>
</tr>
</thead>
<tbody>
<tr>
<td>Positive degree days</td>
<td>256.072</td>
<td>778.523</td>
<td>876.214</td>
</tr>
<tr>
<td>Surface material</td>
<td>snow</td>
<td>ice</td>
<td>ice</td>
</tr>
<tr>
<td>In situ melt (m)</td>
<td>3.725</td>
<td>3.931</td>
<td></td>
</tr>
<tr>
<td>In situ melt (m w.e.)</td>
<td>1.8625</td>
<td>3.5379</td>
<td></td>
</tr>
<tr>
<td>DDF (KS/I; m/°C)</td>
<td>0.015</td>
<td>3.931 / 778.523 = 0.005</td>
<td></td>
</tr>
<tr>
<td>DDF (KS/I; m w.e./°C)</td>
<td>0.007</td>
<td>3.931 / 778.423 = 0.004</td>
<td></td>
</tr>
<tr>
<td>Extrapolated melt (M; m)</td>
<td>(0.005 \times 876.214)</td>
<td></td>
<td>4.424</td>
</tr>
<tr>
<td>Extrapolated melt (M; m w.e.)</td>
<td>3.981</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure 1. (A) Inserting an ablation stake into the glacier after using the steam drill to drill a hole in the snow and ice, (B) Ablation stake 2, (C) Northwest cruisers snowmobile team transporting installation gear.
Appendix II

Stream Monitoring Efforts

I attempted to quantify stream discharge in two meltwater creeks below the Easton glacier’s terminus by installing automated stream gauges in each creek. Each site had a stilling well of PVC pipe attached to a large boulder. The pipe had a Solinst instream level logger connected to a data logger to record a continuous time series of stream stage. I used a March-McBirney Flowmate to measure river velocity and stream discharge at a range of flows at the same cross-section location during each field visit. By plotting the discharge measurements against stage height at each visit, I attempted to construct a rating curve to convert the stage data into a continuous record of discharge throughout the summer. After accounting for precipitation, atmospheric pressure changes, and groundwater contributions, streamflow changes should relate to glacier ablation. Unfortunately the meltwater creeks on the Easton glacier are dynamic and a high flow event carried one of the level loggers downstream in the middle of August. Additionally, substantial aggradation and erosion of the stream bed near the level loggers likely skewed the stage data. I was able to derive a meaningful rating curve at outlet creek 2 (Figure 4), but not in the main meltwater channel (outlet creek 1). I attempted to scale outlet creek 1 to outlet creek 2’s discharge, but I ultimately decided the data had too much uncertainty to draw meaningful conclusions.

Figure 2. Stage data (kPa) for meltwater creek 1 between July 23rd and August 30th, 2018. Stage readings were recorded every two minutes. A baralogger, located next to the stream, was used to correct for air pressure.
Figure 3. Stage data (kPa) for meltwater creek 2 between July 23rd and September 16th, 2018. Stage readings were recorded every two minutes. A baralogger, located next to the stream, was used to correct for the atmospheric pressure. There is a gap in data between August 21st and 29th, when the level logger dislodged and moved downstream.

Figure 4. Diurnal trends in stream temperature and hourly average stage at outlet creeks 1 and 2. The grey markers are temperature, the orange markers are hourly averaged stage at outlet creek 2, and the blue are hourly averaged stage at outlet creek 1.
Figure 5. Rating curve for outlet creek 2. Discharge measurements were taken on 08/04, 08/21, 08/29, 08/30, 09/12, and 09/16.

\[ y = 1.5358x^{1.5684} \]
\[ R^2 = 0.952 \]

Figure 6. Outlet creek 2.