Multi-scale investigation of snow accumulation on alpine glaciers

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Abstract

Glacier mass balance is strongly affected by snow accumulation. Processes such as orographic lifting, preferential deposition, and wind redistribution strongly affect the distribution of snow. To better understand the effects of these processes, statistical models have been developed to relate meteorological and topographic variables to snow accumulation. However, accurate empirical measurements of these variables — which are needed to inform models — are scarce, particularly in remote or difficult to access glaciers such as those within the St. Elias Mountains. The proposed study aims to use snow probing to measure accumulation, and various statistical techniques to examine accumulation variability at the point, hillslope, watershed, and regional scale on three glaciers in the Donjek Range, St. Elias Mountains. These measurements will be used to investigate measurement uncertainty and relevant length scales, the role of topography in determining snow distribution, winter balance measurement optimization, and the transferability of statistical relationships and regional differences in accumulation across a range. This study will provide valuable insight into the variability of snow on glaciers at various scales, and the processes and conditions that influence this variability.

Contents

1 Introduction

Snow accumulation, as the dominant input of mass to alpine glaciers, plays an important role in governing their mass balance and the hydrology of alpine catchments more broadly. This has implications not only for the availability of water for local ecological and human use (Barnett and others, 2005; O'Neel and others, 2014), but also for rates of global sea-level rise (Gardner and others, 2013). It is therefore necessary to understand the spatial distribution of snow on glaciers. However, achieving such an understanding is complicated by the fact that snow distribution in alpine regions is not uniform or static, but rather highly variable and influenced by diverse and dynamic processes operating on multiple spatial and temporal scales. Although previous research has attempted to account for these processes through the development of various techniques of measurement and modelling, little is known about how they operate in glacierized alpine environments. This severely limits possibilities of quantifying and predicting snow distribution on glaciers, particularly in remote locations where frequent empirical measurements are difficult (Nolan and others, 2015).

This proposal examines what is currently known and unknown about the topic of snow accumulation on glaciers and its spatial variability. The following section begins with an overview of accumulation variability within alpine regions in general, as there is a considerably greater breadth of studies devoted to snow in non-glacierized alpine basins. Section 3 reviews different ways of modelling the distribution of snow in alpine environments, while Section 4 describes methods for measuring accumulation and their relative merits and challenges. Section 5 compiles studies that specifically look at accumulation variability on glaciers and summarizes their key findings. Section 6 section focuses on the St. Elias Mountains, a large glacierized region where little is known about snow distribution and its effects on glacier mass balance. The final section described the proposed methodology for measurement of snow distribution in an area of the St. Elias Mountains and statistical techniques for the analysis of these observations.

2 Accumulation variability

The spatial distribution of snow accumulation can vary significantly. This is a result of interactions between spatially and temporally variable atmospheric conditions and heterogeneous topography (Deems and Painter, 2006; Liston and Elder, 2006). Understanding and predicting snow distribution therefore requires accounting for factors that include atmospheric circulation, precipitable water, air pressure, air temperature, wind speed and direction, elevation, slope exposure, presence of orographic barriers, surface slope and aspect, surface roughness, and relief (Schweizer and others, 2008b; McGrath and others, 2015).

2.1 Topographic scales

Snow accumulation is spatially variable on point scales (<5 m), hillslope scales (1–100 m), watershed scales (100–10,000 m) and regional scales (10–1000 km) (Clark and others, 2011). The features and conditions that lead to variability at these scales differ (see Table 1) and their relative importance depends on the topography and climate of the study area. Inclusion

Table 1: Relevant spatial scales for snow variability on glaciers. Information from Clark and others (2011).

Scale	**Length**	**Associated glacier feature**
Point	$<$5 m	Crevasses
Hillslope	1–100 m	Local surface topography (curvature, slope), avalanching
Watershed	100–10,000 m	Elevation, aspect
Regional	10–1000 km	Horizontal precipitation gradient across mountain range

of parameters that describe relevant processes at multiple scales has been shown to improve models that aim to explain measured snow distribution (Marchand and Killingtveit, 2005; Clark and others, 2011).

Point-scale variability is generally associated with surface roughness effects and the presence of small obstacles. These effects can be significant in vegetated landscapes or when the surface is very rough (e.g. boulder field) (López-Moreno and others, 2011). Many parts of a glacier though are characterized by a relatively smooth surface, with roughness lengths on the order of centimeters (Hock, 2005). In these areas, point-scale variability of snow depth is low. However, in heavily crevassed regions, point-scale variability can be large and thus exert a dominant control on snow distribution in the area (McGrath and others, 2015).

Hillslope-scale variability is caused by variations in the surface topography of the glacier. The curvature and slope of the surface as well as the presence of local ridges or depressions can affect where snow is located (Blöschl, 1999; Sold and others, 2013). Avalanching can also redistribute snow, especially on the margins of a glacier (Blöschl and others, 1991; Mott and others, 2008).

Watershed-scale variability results mainly from the effects of changing elevation and aspect on atmopsheric conditions (Clark and others, 2011). In particular, orographic lifting and shading can result in higher elevation and north-facing areas of the glacier having more snow than other areas (Mott and others, 2008; Sold and others, 2013). Gradients in temperature from elevation changes also affect the freezing level, which determines whether precipitation falls as snow or rain (Blöschl and others, 1991). For example, Machguth and others (2006) found a strong influence of elevation in determining accumulation on Findel Glacier in Switzerland.

Regional variability occurs when areas within a mountain range have differing amounts of snow. Often, this results from horizontal precipitation gradients and rain shadows forming on the lee side of topographic divides. Areas with large, steep mountains are especially affected by these processes.

Generally, spatial variability increases with spatial scale (Clark and others, 2011). Extent and spacing of measurements must therefore capturing variability both across the study area and at smaller scales. Clark and others (2011) note that studies of snow water equivalent (SWE) that have been conducted in alpine environments vary considerably in the extent and spacing of their measurements.

2.2 Snow drift and preferential deposition

Snow drift and preferential deposition are crucial factors that influence the distribution of snow (Lehning and others, 2008; Winstral and others, 2002; Clark and others, 2011). Sharp changes in topography cause convergent and divergent airflows close to the surface, leading to turbulence and vorticity. This terrain induced turbulence modifies mean wind (and snow particle) velocities, and can thus influence snow distribution via snow drift and/or preferential deposition (Mott and others, 2008; Lehning and others, 2008; Dadic and others, 2010).

Snow drift is the erosion and deposition of already deposited snow (Dadic and others, 2010). In general, erosion on the windward side is caused by increased wind speeds and deposition on the lee side of ridges is due to decreased wind speeds (Liston and Sturm, 1998; Mott and others, 2008; Dadic and others, 2010). Mott and others (2010) found that creep, which is the rolling of snow particles on the snow surface, and saltation, which is the bouncing and dislodging of snow particles, are primarily responsible for the formation of cornice-like features.

Preferential deposition is inhomogeneous precipitation in the absence of local erosion (Lehning and others, 2008). It is mainly governed by winds, where higher wind velocities and updrafts on the windward side of ridges cause reduced deposition while reduced wind velocities on the lee side enhance deposition. This process can occur at relatively low wind speeds because it does not require the lifting of already deposited snow — instead, it only needs to act with or against the falling snow (Mott and others, 2008; Dadic and others, 2010). For example, Mott and others (2011) found that the spatial structure of snow distribution in an alpine bowl was dominated by the preferential deposition of precipitation due to altered air flow fields.

Both processes described can occur at multiple spatial scales. Enhanced accumulation has been observed on the point-scale in small depressions and on lee sides of obstacles, on the hillslope scale on the lee side of ridges, and at the watershed scale on sheltered aspects (Harrison, 1986; Blöschl and Kirnbauer, 1992; Mott and others, 2008; Winstral and others, 2002; Clark and others, 2011).

3 Snow distribution models

The distribution and variability of a parameter, such as snow water equivalent (SWE), can be estimated using either dynamic or statistical models. These models help to determine relevant processes that affect the distribution of snow, generally by relating its distribution to meteorological and topographic descriptors or conditions. Inferences made from these models drive the direction of future studies and provide valuable insight into understanding why variability arises. Accounting for wind in snow distribution models is especially important because it plays a dominant role in spatial patterns of accumulation (Winstral and others, 2013).

3.1 Dynamic models

Deposition and redistribution of snow can be represented using physically based, spatially distributed models. The general aim of these models is to simulate surface processes and how they vary spatially and temporally (Mott and others, 2008). These models usually consider atmospheric conditions including freezing level, precipitation rates, relative humidity, and wind speed and direction, as well as processes such as orographic lifting, cloud formation, downslope evaporation, advection and fallout, snow metamorphism, and wind redistribution of snow (erosion, saltation) due to terrain induced turbulence (Smith and Barstad, 2004; Liston and Elder, 2006; Lehning and others, 2008; Mott and others, 2008). Modelling the dynamically induced flows of these components together describes the preferential deposition and redistribution of snow in alpine environments (Lehning and others, 2008; Mott and others, 2008; Dadic and others, 2010).

Many models have been developed to describe preferential deposition and redistribution. Early models were developed for flat or gently rolling terrain where boundary layer flow is better understood (Dadic and others, 2010). Boundary layer flow in steep terrain is generally non-linear though (Mott and others, 2008; Dadic and others, 2010), so a number of different approaches have been applied. For example, Dadic and others (2010) and (Lehning and others, 2008) solved non-hydrostatic, compressible Navier-Stokes equations in 3-D and aimed to conserve momentum, heat, mass, and states of water, to model wind flow velocities. Smith and Barstad (2004) employed Fourier transforms in a linear orographic model, which allowed for the more accurate representation of complex terrain.

Dynamic models are a valuable way to determine accumulation variability. Since they use physically consistent processes, they can be applied to any site and generate values for each grid cell. They can also be used in different climatic conditions, allowing for predictions of accumulation change (Clark and others, 2011). Furthermore, a historic data set of variables is not needed to generate a meaningful output. Another advantage of dynamic models is that they allow for high temporal resolution (e.g. Mott and others (2008) have a 1 hour time step), which allows for snowpack evolution to be examined.

Application of dynamic models is however operationally complex and computationally expensive, and also requires a diverse set of observations. Input parameters usually include temporally varying values for precipitation, wind speed and direction, air temperature, and relative humidity (Liston and Elder, 2006). Although these parameters can be obtained from meteorological stations, spatially distributed values — found using atmospheric models — are also required (Liston and Elder, 2006; Mott and others, 2008). For example, Dadic and others (2010) used meteorological data from three automatic weather stations located throughout the study basin, while in the study done by Mott and others (2008) monthly stake measurements were needed. Such a well-monitored basin can be difficult to achieve in remote or inaccessible areas. A number of other dynamic models, such as those described in Fowler and others (2007), use general circulation model (GCM) values to drive local circulation models, but it was shown that model output strongly depended on GCM boundary choice. Even with sufficient and appropriate input data, the models must assume a number of parameters and simplify parameter relationships (e.g. constant mean wind speed (Mott and others, 2008)) to characterize the atmosphere, which may not realistically describe air movement and stability. Additionally, the models do not account for all modes of snow transport. For example, snow deposition due to avalanching can be an important process of snow transport that is not captured in current models (Mott and others, 2008).

3.2 Statistical models

Statistical models of snow variability establish empirical relationships between snow distribution and external variables (Fowler and others, 2007). These models assume that local distribution is forced by external factors, such a meteorological conditions or topography.

3.2.1 Statistical downscaling

Statistical downscaling is the process of determining an empirical relationship between large-scale atmospheric conditions and regional climates (Fowler and others, 2007). In general, this relationship is expressed as a function F such that regional variables R are found by $R = F(X)$ where X encompasses large-scale climate variables (Fowler and others, 2007). These models are trained and validated using gridded reanalysis data from GCMs (X) and point observations (R). Performance of these models is measured using correlation coefficients, distance measures (e.g. root mean squared error), or explained variance (Fowler and others, 2007).

There are three main types of statistical models (Fowler and others, 2007). The first is a regression model, which directly quantifies a relationship between a local variable and a number of large-scale variables. Statistical methods such as multiple linear regressions (Hanssen-Bauer and Førland, 1998), principal component analysis (Kidson and Thompson, 1998), canonical-correlation analysis (Busuioc and others, 2001), neural networks (Zorita and Von Storch, 1999), and singular value decomposition (Widmann and others, 2003) can be employed. The second type of model is a weather typing scheme, which relates the occurrence of a particular 'weather class' to local variables. Weather classes can be found using empirical orthogonal functions or cluster analysis (Fowler and others, 2007). The third model uses weather generators that simulate local precipitation occurrences with a chosen distribution of precipitation amount (Fowler and others, 2007). The large-scale input variables that are usually chosen (e.g. sea-level pressure, geopotential heights) for statistical downscaling are representative of large-scale circulation. Increasingly, other variables such as humidity are being incorporated into analyses to account for mechanisms that rely on thermodynamics and vapour content (Fowler and others, 2007). When focusing on precipitation, integrated vapour transport (IVT) can be used as a proxy for precipitable water in the atmosphere and can be used to identify corridors of large water vapour transport that correlate with intense precipitation events (Neiman and others, 2008).

Statistical models are a simple way to examine variability. They are computationally efficient and comparatively easy to apply because they are based on standard and accepted statistical procedures (Fowler and others, 2007). Furthermore, generating values for specific point-scale variables does not require prior knowledge of all the processes that affect it. This allows for a function to be established for variables where all processes are currently not accounted for or where many processes are equally important.

Although the application of statistical downscaling is simple, it has a number of disadvantages. The method is difficult to apply in areas that have a small amount of observed historical data, as model performance is better when a long and reliable data set is used for calibration (Fowler and others, 2007). Statistical downscaling also assumes that the relationship between large-scale and local variables is stationary (Fowler and others, 2007). This means that use of the determined function is limited for projecting the variable of interest and further implies the need for long-term data observations. Furthermore, the empirical relationship assumes that there is no climate system feedback and the data generated through the empirical function are subject to the same biases as those of the original data set (Fowler and others, 2007). Wilby and Wigley (2000) also notes that the choice of large-scale variable domain (location and spatial extent) exerts a strong effect on the accuracy of the empirical function.

3.2.2 Terrain-based parametrization

Terrain-based parametrization is the linking of topographic indices determined from terrain modelling and observed conditions. To determine topographic indices, the terrain in the study area is divided into grid cells where terrain parameters (e.g. slope, curvature, aspect, "northness", wind exposure, topographic similarity) are calculated (Anderson and others, 2014; McGrath and others, 2015). The variable of interest is then measured in the study area and a relationship between grid-cell terrain parameters and observed data can be established (e.g. Blöschl and others, 1991; Liston and Sturm, 1998; Anderton and others, 2004; McGrath and others, 2015).

This method requires a good terrain model and a meaningful network of observed data. For example, Molotch and others (2005) found that that there were significant differences in modelled snow distribution when different terrain models were used. This is likely because the terrain-model grid size affects the value of the calculated terrain parameter for the cell. Terrain-based parameterization also needs sufficiently high resolution and spatial extent of the observed data. When measuring accumulation, the variability within the study area needs to be captured and all areas should be well represented.

Relating terrain model parameters with observed data is often accomplished with simple statistical methods. Multiple linear regressions (Marchand and Killingtveit, 2005; Sold and others, 2013; McGrath and others, 2015), mixed-effects multiple regression (Kasurak and others, 2011), parametric probability distributions (Clark and others, 2011), bivariate screening (Anderton and others, 2004), probability distribution functions (Kerr and others, 2013), and regression tree models (Elder and others, 1998; Winstral and others, 2002; Molotch and others, 2005; Revuelto and others, 2014; Wetlaufer and others, 2016) are among the more popular models. These models statistically relate snow distribution to terrain parameters with varying success. For example, the multiple linear regression model developed by Sold and others (2013) explained about 50% of the variance of snow depth while the model developed by Anderton and others (2004) explained 70–80% of the variance in snow water equivalent. A number of studies have found that elevation and wind redistribution parameters explain the majority of the variance in observed snow depth or snow water equivalence (e.g. Erickson and others, 2005; Trujillo and others, 2009; Schirmer and others, 2011; Grünewald and others, 2014; McGrath and others, 2015). Interaction parameters (e.g. $slope \times orientation$) have also been found to be significant predictors for precipitation distribution in alpine areas (Basist and others, 1994). Erxleben and others (2002) and Molotch and others (2005) note that many relationships between accumulation and controlling parameters are nonlinear so use of regression tree models yields better results. Combining statistical models has also been seen to improve model accuracy. Examples include combining linear regressions with generalized additive models (López-Moreno and Nogués-Bravo, 2006) as well as binary decision trees with kriging (Balk and Elder, 2000).

Relating topographic parameters and observed data is a simple approach to understanding processes that affect variability. Although no physically-based relations are employed, terrain parameters can act as proxies for processes that are known to occur (McGrath and others, 2015). In this way, dominant processes can be inferred through easy to find statistical relationships. This approach is especially powerful in areas such as alpine environments where topography strongly affects local conditions.

While terrain-based parametrization is easy to employ, its usefulness in understanding snow distribution can be limited. This method assumes that variability between cells is larger than within cell variability, which may not necessarily apply to all cells, especially in steep terrain or where grid size is comparatively large. Marchand and Killingtveit (2005) found that the standard deviation within the 30 m by 30 m grid cells used in the study was slightly larger than the between-grid variability. Additionally, Grünewald and others (2013) observed that local statistical models are able to perform well but that they cannot be transferred to different regions and that regional-scale models are not able to explain the majority of variance. The temporal transferability of terrain-based parameterization is also not reliable. Grünewald and others (2013) found that local models could be applied between years while Revuelto and others (2014) found that snow distribution variability could not be explained by their model in low snow years. Furthermore, the use of terrain parameters as proxies does not provide meaningful insight into relevant processes. This is important when attempting to predict distribution of variables in a different climate or location.

3.2.3 Variograms

Length scales are distances over which data is correlated and they describe the degree of spatial dependence of a variable, which is likely due to a controlling processes. It is important to identify scaling processes of snow properties so that more effective models that account for processes at many scales can be developed (Blöschl, 1999; Deems and others, 2006). For example, the correlation length of snow depth can give insight into the effects of underlying topography and wind effects. The length scale of snow depth changes with time and is dependent on the amount of snow, influence of wind, and occurrence of melt.

Geostatistical techniques and fractal analysis are the two main categories of tools that can be used to determine the length scale. Geostatitical techniques characterize a spatial pattern as composed of random deviations from a mean value and neglects any spatial structure that could exist in the random field (Deems and others, 2006). These tools are used for distinguishing between autocorrelated and uncorrelated scale regions and have been widely used for snow science applications (e.g Blöschl, 1999; Deems and Painter, 2006; Marshall and others, 2006). Fractal analysis allows for stochastic structure within the random component and is able to identify multiple scales of variability (Deems and others, 2006). Changes, or breaks, in variability scales are linked to important changes in processes that affect the observed spatial pattern. Fractal geometry (also known as scale invariance or self-similarity), has been observed in snow distribution patterns (Shook and Gray, 1996; Granger and others, 2002; Deems and others, 2006; Trujillo and others, 2007; Deems and others, 2008) and has been used to interpolate between snow depth observations (Shook and Gray, 1997).

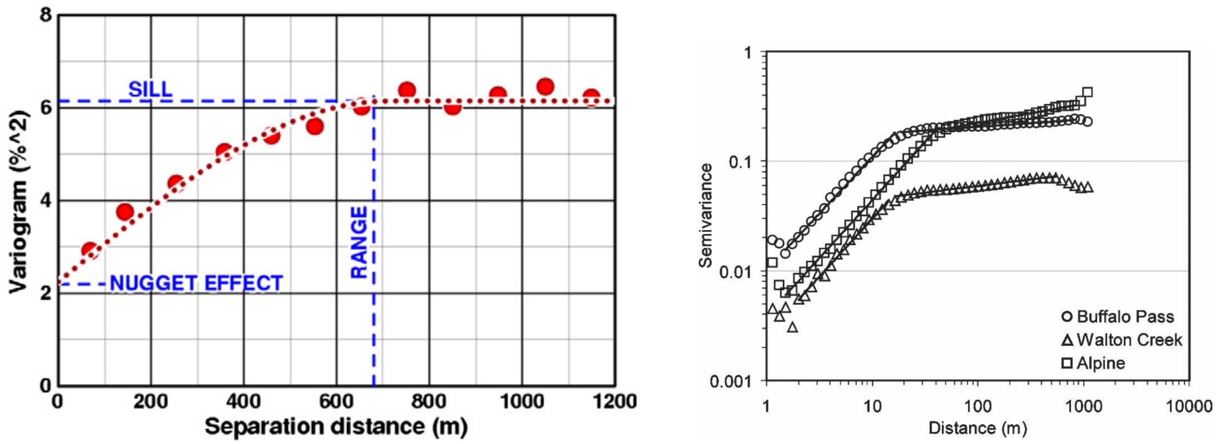
The most traditional geospatial technique is to construct a variogram (or semivariogram), which is a plot of the variance between points at two locations (Figure 1(a)). The goal of the variogram is to estimate the autocorrelation structure of the underlying stochastic process. The semivariance $\gamma\,(h)$ is calculated by

$$\gamma(h) = \frac{1}{2\,|N(h)|} \sum_{N(h)} \left(z_i - z_j\right)^2 , \tag{1}$$

where $N(h)$ is the set of all pairwise Euclidean distances $i - j = h$, $|N(h)|$ is the number of distinct pairs of $N(h)$, and z_i and z_j are the data vales at spatial locations i and j, respectively (Schirmer and Lehning, 2011). Variance will increase as the distance between points increases, indicating that the values at these points are increasingly less dependent on each other. The distance at which this value no longer increases is the correlation length (also called the range) and the variance at this point (random field) is referred to as the sill ($\lim_{h\to\infty} \gamma(h)$) (Srivastava, 2013). In practice, the correlation length exists only for stationary processes with a constant mean and is defined as the distance at which semivariance reaches 95% (Deems and others, 2006). Variability that was not captured in the measurement, due to insufficient sampling resolution and measurement error, is referred to at the nugget and is the y-intercept of a variogram. Data with a high degree of spatial structure will exhibit a steep slope for the sill is reached and data with low spatial structure would show a more gradual slope. Other analyses that present similar information to the semivariogram include autocorrelation and covariance.

Information from variograms is often used for interpolating point measurements to find spatial patterns. Reuter and others (2016) claim that geostatistical methods are preferred over simple regressions for interpolating data because they incorporate spatial autocorrelation directly from measurements. Kriging is a commonly used interpolation technique that assigns weights for observed values based on the data covariance. There are many types of kriging, including ordinary, universal, block, and external drift, which have different assumptions about the mean of the data and include different components of the data (e.g. residuals) (Webster and Oliver, 2001).

Fractal analysis attempts to identify scale-invariant spatial patterns, which means the observed variable has similar statistical properties at multiple scales. In this case, spatial pattern characteristics can be transferred from one scale to another using a scaling factor and this can also provide information about the scale, scope, and resolution of modelling and sampling efforts (Deems and others, 2006). The most common way of identifying scale-invariance and scale breaks is by analyzing the slope of a log-transformed variogram (see Figure1 (b)), where log-linear segments indicate self-similar, fractal distributions (Deems and others, 2006). For example, Schirmer and Lehning (2011) used fractal analysis to confirm the effect of dominant wind direction on snow distribution by examining differences in scale break between windward, lee, and cross-loaded slopes. Power spectra (log-log plots) can also be used to examine scale-invariant patterns with the wave number and spectral exponent representing the spatial scale and degree of variability, respectively (Trujillo and others, 2007, 2009).



(a) Variogram

(b) Log-log variogram

Figure 1: Example of a variogram (from Srivastava (2013)) and a log-log variogram with scale breaks (from Deems and others (2006)).

4 Measuring accumulation

Determining accumulation requires knowledge of snow density and depth. Measuring these parameters within a glacierized basin has many challenges. Basin location and topography affect accessibility, while the cost and time required to conduct measurements can be prohibitive. In most cases, the resolution of measurements over a large area is insufficient to approximate the true variability (Blöschl, 1999; Deems and others, 2006).

The chosen scientific question also guides choices in measurement tools and sampling design. Drivers of variability should be considered prior to sampling so that a sampling pattern is designed to capture this variability and to avoid sampling bias. When designing a snow survey, the support, spacing, and extent of the observations need to be defined (Blöschl and Sivapalan, 1995). The support refers to the area or footprint of each measurement (tool dependent), the spacing is the distance between measurements, and the extent is the region that is being sampled.

Snow density can be measured directly or with models of snow density change. To measure bulk density, a column of snow with a known volume is excavated (in a snow pit or with a firn corer) and weighed (Sold and others, 2013, 2014). Usually, a number of snow column densities are measured and the average density is taken as representative of glacier-wide density (e.g. Machguth and others, 2006; Grunewald and others, 2010; McGrath and others, 2015). This can result in error when calculating snow water equivalence (SWE) because density can vary spatially and temporally (due to total snow depth, elevation, solar radiation, and wind effects) in a way that is not captured by a limited number of snow density measurements (Grunewald and others, 2010; Wetlaufer and others, 2016). However, snow density has been seen to vary over greater spatial scales than snow depth so fewer density measurements are usually made (Elder and others, 1998; Clark and others, 2011).

Snow and firn density can also be calculated using models that account for relevant processes such as compaction from overlaying snow and refreezing of meltwater (Herron and Langway Jr, 1980; Sold and others, 2014). Densification processes are difficult to capture in models though, so they should be applied with caution (Mellor, 1974).

Three main methods are currently used to measure snow depth. Probing involves taking *in situ* point measurements of snow depth, GPR surveying involves using radar to detect the snow depth along continuous lines, and DEM subtraction involves taking the difference between the glacier surface at the end of the ablation and at the end of the accumulation seasons to find snow depth. Methods are selectively applied based on desired spatial resolution, cost effectiveness, and equipment availability. Each is prone to different sources of error and there is ongoing research to reconcile these approaches (Sold and others, 2014).

4.1 Snow probing

4.1.1 Measurement

The most direct way of measuring snow depth is by probing. To determine the snowpack thickness, the height of the snow above the end of the previous year's ablation surface is measured. Usually, a number of snow height measurements are obtained close to each other and the mean value is taken to be representative of that location. For example, Machguth and others (2006) took the mean of nine snow probe measurements within a 7 m radius as representative of a test site in the ablation zone.

Figure 2: Digging a snowpit in the accumulation area of Haig Glacier, Rocky Mountains

In the ablation zone, snow depth is easy to measure because the interface between the end of summer melt surface and the beginning of winter accumulation is well defined (i.e. glacier ice) (McGrath and others, 2015). In the accumulation zone however, the snow surface at the end of the melt season may not be easily distinguishable from the winter accumulation (Grunewald and others, 2010). It is common for the accumulation zone to have a heterogeneous surface at the end of the melt season — some areas do not experience any melt, while other areas experience some melt and the melt water percolates through the snow and firn. Melt water generated from warm weather or rain events, especially in the early and late parts of the accumulation season, can refreeze in the snowpack to form ice lenses (Sold and others, 2014). As a result, the interface can be difficult to observe and contain a combination of dense or compacted snow, ice lenses, and/or firn. Probing in the accumulation zone can therefore result in erroneous depth measurements — penetration into the dense snow or firn will make it seem like a deeper snowpack exists and probing to an ice lens within the snowpack can make it seem like shallower snow is present (Sold and others, 2013). Snow pits and firn cores are therefore used to examine snow and firn layers to determine where the current season's snow begins.

To determine the glacier-wide SWE, point snow depth measurements from probing need to be interpolated and extrapolated. This is often done using a statistical regression on parameters such as slope, aspect, curvature, and susceptibility to wind redistribution (e.g. Wheler and others, 2014; McGrath and others, 2015). A regression generates an equation that is site specific and is used to estimate SWE for each grid cell based on the values of its relevant parameters.

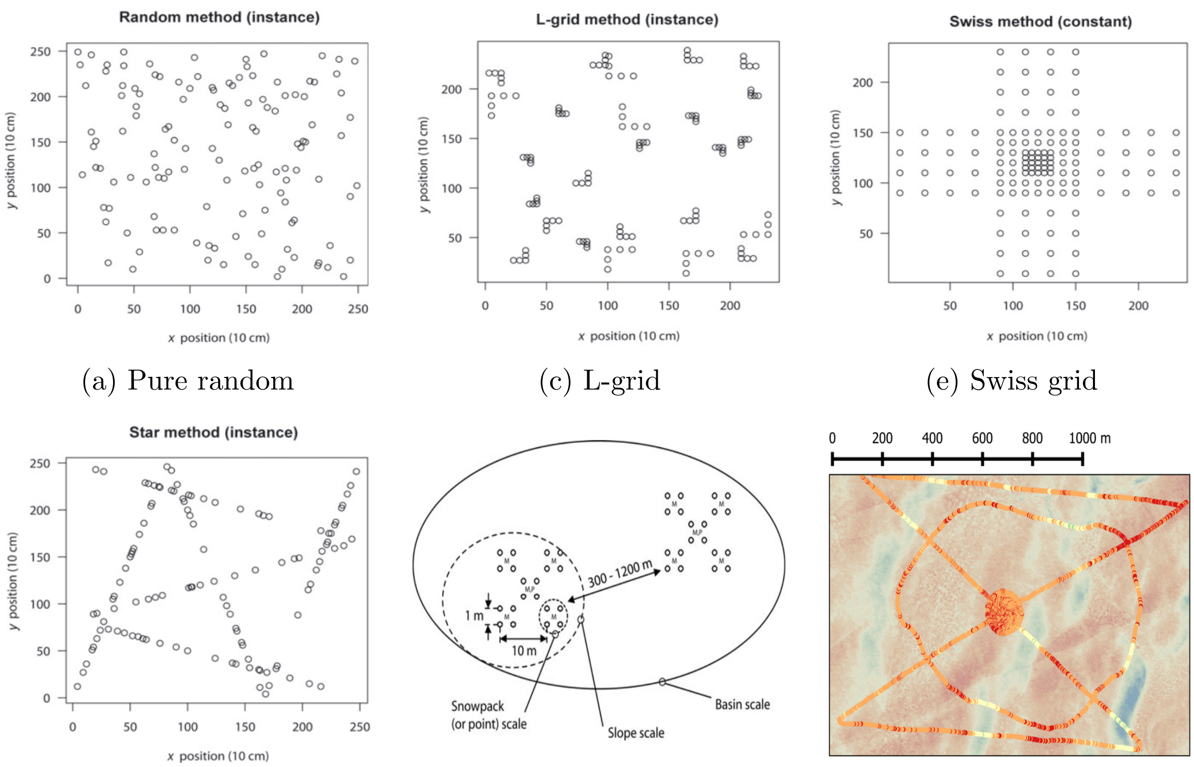
Snow probing is the simplest and oldest method used to determine accumulation. At the most basic level, it requires little more than a probe to determine depth, a way to determine location (such as a hand-held GPS), and a shovel to dig snow pits (see Figure 2). Furthermore, this method directly measures snow depth so no data processing or corrections are needed and depth uncertainty is simple to quantify (often multiple depth measurements are taken close together) (Sold and others, 2013).

There are however many drawbacks to this method. *In situ* probing and digging snow pits are incredibly time-consuming (Deems and Painter, 2006). This limits the number of measurements that can be made, which means that accumulation measurements are under-represented and spatial variability in accumulation is difficult to capture (Sold and others, 2014). Measurement is also limited to areas that are both accessible and safe for researchers. In complex terrain many areas cannot be surveyed, resulting in data gaps. Sold and others (2013) noted that this systematic bias can result in incorrect values of glacier-wide accumulation — particularly because inaccessible areas such as cliffs and ridges have relatively shallow accumulations (due to wind erosion), while heavily crevassed areas can accumulate deep snow packs.

4.1.2 Sampling Design

Optimal sampling schemes for snow probing are central for accurately estimating snow distribution and mass balance from *in situ* measurements. Measuring snow depth and travelling between measurement locations is both time consuming and can disturb the snow so care must be taken to choose a sampling scheme that avoids bias, allow for the greatest variability to be measured, and minimize distance travelled (Shea and Jamieson, 2010). A design that maximizes accuracy and minimizes effort is therefore desired (Elder and others, 1991) and both theoretical (Trujillo and Lehning, 2015) and applied (Kronholm and others, 2004; Shea and Jamieson, 2010) investigations of various sampling designs have been pursued. There are a number of different designs that have been employed to obtain point measurements, including pure random, linear random, nested, gridded random, and gridded.

A purely random distribution of points is favourable because it avoids all bias, has the best correlation with true distribution, and is likely to capture the most variability (Kronholm and Birkeland, 2007; Shea and Jamieson, 2010). Logistically though, it is difficult to successfully measure all points in the study area because some may be impossible to access and some may be disrupted during travel or measurement. This design often results in inefficient travelling routes, which decreases the number of possible point measurements. Elder and



(a) Pure random (c) L-grid (e) Swiss grid

(b) Star (d) Nested (f) Hourglass

Figure 3: Examples of snow sampling schemes. Figures (a), (b), (c), and (e) from Shea and Jamieson (2010). Figure (d) from Schweizer and others (2008a). Figure (f) from Parr, C., (2016 personal communication).

others (1991) show a simple basin-wide random sample is a less optimal sampling scheme than a stratified random sample that accounts for known variation. One instance of a purely random sampling scheme can be seen in Figure 3(a).

Linear-random sampling schemes impose a structure to traverse as much of the study area as possible but allow for a random distance between sampling points. An example of this scheme is the ‘star’, created by Shea and Jamieson (2010). A significant advantage of this scheme is that it was designed to minimize distance travelled while still measuring snow properties in various orientations and various distances apart, which reduces bias. However, since the observer travels in straight transects there is still a potential to miss smaller features or ones that parallel the transects (spatial autocorrelation). Shea and Jamieson (2010) used comparative Monte Carlo simulations to validate that the star scheme performs equivalently or better than other (more structured) sampling methods and that it converges to the true distribution as well as a purely random scheme. One instance of a linear-random sampling scheme can be seen in Figure 3(b). Linear-random sampling can also be done in an ‘hourglass’ shape with an inscribed circle (referred to as an hourglass sampling scheme). As seen in Figure 3(f), this pattern allows sampling in all directions and captures a wide range of snow depths from the underlying snow distribution (Parr, C., 2016 personal communication).

Gridded-random designs involve dividing the study area into equal sized areas and then sampling randomly within each area. The L-grid is an example of this scheme (Bellaire and Schweizer, 2008; Elder and others, 2009; Bellaire and Schweizer, 2011). In this scheme, the study area is divided into a grid and in each cell a random location is chosen as the start of the transect. The cardinal direction and measurement spacing of the transect are chosen randomly. Transects consist of five measurements, with three in the first direction selected and two perpendicular to this, forming an 'L' shape. This scheme has a small error bias (maintains randomness), while allowing for more efficient measurement (Shea and Jamieson, 2010). Compared to the star scheme, the L-grid does not have a consistent travel distance and involves constant reorientation and finding of transect start locations, which decreases its efficiency. One instance of a gridded-random sampling scheme can be seen in Figure 3(c).

Nested sampling is a scheme that maintains a certain sampling pattern and applies it at various length scales. For example, Schweizer and others (2008a) took four point measurement in a 1 m square and did this 10 m apart to form another square. The set of measurements was then repeated 300–1200 m away to capture basin scale values. Hierarchal trees that incorporate selected parameters are often used to determine nested sampling locations. Watson and others (2006) and Kasurak and others (2011) use hierarchal sampling to divide the study area into regions (often discontinuous) that are likely to have a similar snow distribution and low variance, which means they require lower density sampling. Nested sampling requires that the observer predetermine parameters that affect the spatial pattern of the variable. Often, remote sensing is used to obtain these parameters so the resolution of regions is limited to that of the remote sensing images (typically 30 m resolution). The variability that exists at scales less than the grid-size of the images is classified as being caused by 'random' effects, which are assumed to be unbiased and unpredictable (Watson and others, 2006). Nested sampling is well suited for regions with many complex and interacting parameters. For example, Watson and others (2006) used a hierarchal tree with time (traveling between locations), elevation, vegetation, and solar radiation at various length scales to create subgroup to sample. A nested sampling scheme can be seen in Figure 3(d).

Gridded sampling designs use regular measurement intervals in a grid pattern. Many variations of this scheme exist (Molotch and Bales, 2005; Kronholm and Birkeland, 2007; López-Moreno and others, 2011) with the most popular one being the Swiss cross (Kronholm and others, 2004). This nested arrangement allows for a larger area to be covered than a fully quadratic grid and measures at various spatial scales, leading to more reliable geospatial statistics. This method allows for easy measurement and reveals details at various scales. However, measurements are biased by regular spaced intervals and linear orientation, which could result in an under representation of the snow distribution further from the centre. A gridded sampling scheme can be seen in Figure 3(e).

4.2 GPR

Ground penetrating radar (GPR) can be used to find snow depth along continuous lines. This method is used to calculate the distance from the radar source to a boundary with a strong contrast in dielectric permittivity, which corresponds to a change in material properties (Sold and others, 2013). When the speed of the radar wave through the material is known, the travel time can be measured and from this the distance calculated. On a glacier, the radar wave is able to penetrate snow and ice at MHz frequencies and the strongest reflections arise when water is present (Sold and others, 2013). To measure snow depth, GPR units are mounted on aircrafts or snowmobiles that then travel over snow covered areas (see Figure 4) (Machguth and others, 2006; McGrath and others, 2015). The resulting processed radargram (e.g. Figure 5) gives a continuous snow depth profile. Processing of the radargrams involves using tracking algorithms that are able to trace continuous layers. Interpolation between transects is then done to find the glacier-wide accumulation. McGrath and others (2015) describe this process in five steps: (i) acquisition of GPR and probing data (ground truthing), (ii) calculation of snow density and radar velocity, (iii) calculation of snow thickness and resulting SWE, (iv) application of a correction to measured accumulation based on ground truth data, and (v) use of a multiple regression model to extrapolate SWE across the glacier. The extrapolation of SWE can also be done using an inverse approach with a coupled surface energy-balance snow model (van Pelt and others, 2014).

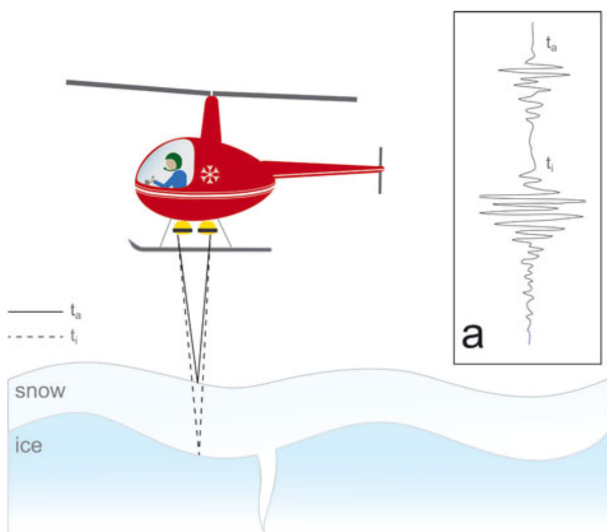


Figure 4: Schematic diagram of a helicopter-borne radar snow survey. The travel time for the signal to interact with the snow surface is shown as a solid line and the signal travel time of the interaction with the ice is shown as a dashed line. Together, these values can be used to determine snow depth. The inset (a) is an example waveform that would be recorded from these two events. Figure taken from Gusmeroli and others (2014).

GPR is an effective tool for measuring accumulation. It provides continuous snow depth transects, which means that spatial variability is well represented along the radar lines. Surveys also need be conducted only once to gain depth observations, which makes data collection fast and and reduces logistical efforts(Machguth and others, 2006). GPR snow depth estimates are not affected by glacier dynamics and the ability to fly over steep or inaccessible regions means that all areas of a glacier can be measured.

A large limitation of GPR is the difficultly of processing radargrams. Areas where the snow–ice boundary is not well defined (i.e. the accumulation area) lack clearly contrasted material properties, which can lead to misinterpretation of their internal layers (McGrath and others, 2015). In the ablation area, the presence of crevasses can also result in radargram misinterpretation (Machguth and others, 2006). Variation and uncertainty in radar wave speed due to differing snow density and liquid water content can also affect depth calculations (Sold and others, 2013). Often, wave speed is only measured in a few reference locations so changes in snow pack properties cannot be accounted for. Lastly, there is no universal procedure for processing GPR data. Selection of parameters and processing algorithms is dependent on available equipment, field conditions, survey design and intention (Sold and others, 2013), which hampers the reproducibility of surveys.

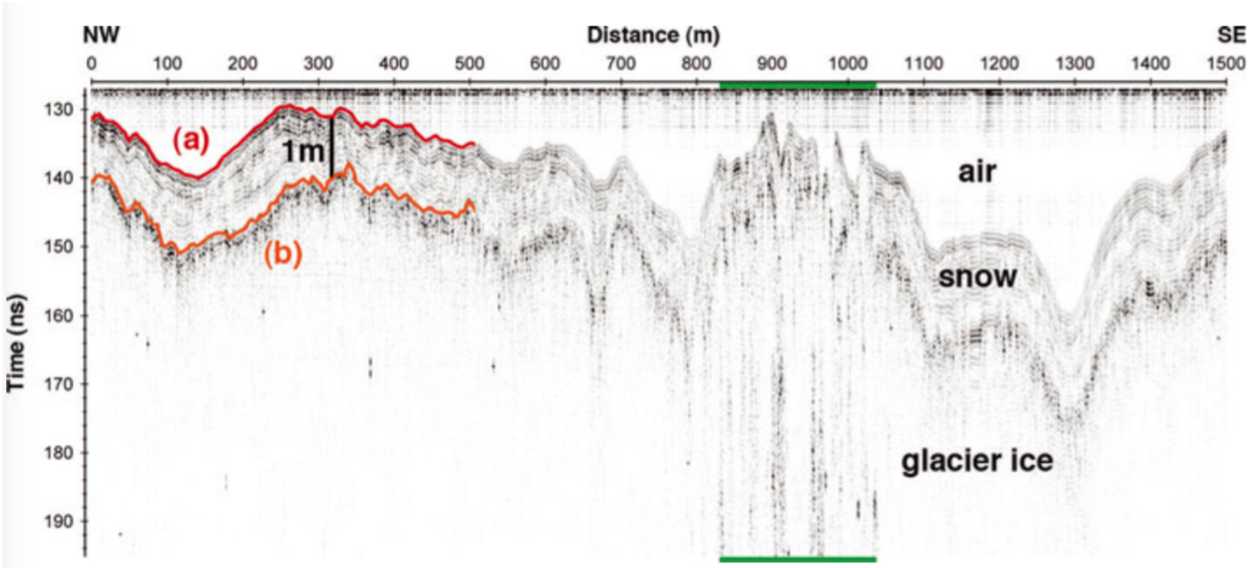


Figure 5: Radargram from the accumulation area of Findelengletscher, Valais, Switzerland. (a) The reflection at the air-snow interface. (b) The reflection at the snow-ice interface. Figure taken from Sold and others (2013).

4.3 DEM subtraction

Digital elevation model (DEM) subtraction involves taking two detailed surface topography scans — one at the end of the melt season and one at the end of the accumulation season — and subtracting them from each other to find the snowpack height. The largest advantage of this remote sensing method is that it provides a highly resolved spatial measurement of snow depth over an entire basin (Deems and Painter, 2006; Sold and others, 2013). Data collection is fast, although two surveys must be conducted. This technique is sensitive to other processes that change the glacier surface elevation, including vertical displacement due to ice flow (positive in the ablation area and negative in the accumulation area), firn compaction, and surface lowering due to melt after the acquisition of the end of melt season DEM (Sold and others, 2013). For example, (Sold and others, 2013) found that a first-order approach (where the observed elevation change was interpreted as snow accumulation) was inconsistent with snow depth probing — DEM subtraction showed decreasing accumulation with an increase in elevation. Corrections can

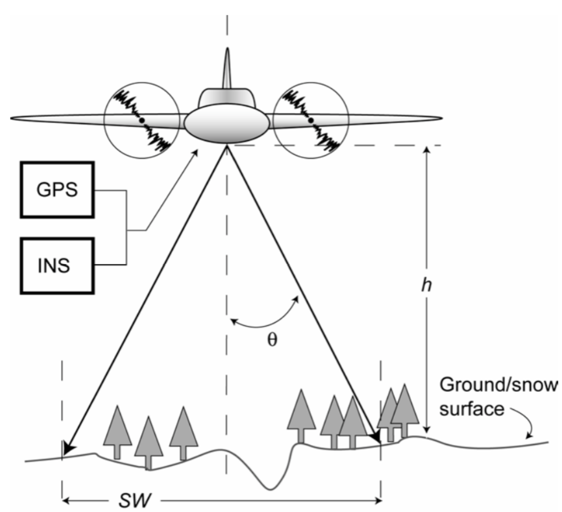


Figure 6: Schematic of airborne LiDAR system geometry. Scan angle (θ), height (h), and swath width (SW) are shown. Figure from Deems and Painter (2006).

be made to account for these discrepancies but they rely on *in situ* measurement of snow depth, knowledge of long-term mass balance, or information about the vertical displacement of ice from GPS towers (Sold and others, 2013).

Lidar and photogrammetry are the two main methods of producing DEMs. Lidar produces a surface elevation model by calculating the distance to a target (by measuring the time between an emitted and return laser signal) (Deems and Painter, 2006; Sold and others, 2013). Terrestrial lidar systems involve stationary units placed in vantage points from which they are able to scan the basin surface (Grunewald and others, 2010). Large basins require multiple overlapping scans to acquire a complete surface profile. Airborne lidar systems (see Figure 6) can also only scan a certain size footprint so the aircraft must fly over all parts of the basin to acquire a full surface profile. These systems also require an accurate global positioning system (GPS) — which is often corrected by referencing to a stationary GPS — to determine the 3D locations of the surface (Deems and Painter, 2006). Airborne systems are widely applied and favourable in large basins or ones where no vantage point exists or is inaccessible. However, these systems are expensive so terrestrial scanners, which are comparatively more cost effective, are becoming popular (Grunewald and others, 2010).

Complex topography and multiple laser reflections can cause problems when producing a DEM from lidar data (Deems and Painter, 2006). Significant vertical changes result in the spreading of the laser footprint and an incorrect interpretation of distance. (Deems and Painter, 2006) shows that an error of 50 cm can result from lidar scans of 45° terrain from 1000 m flight height. Careful planning of flight paths can reduce this error. Scattering of laser light and penetration into the snow pack can also introduce error into height calculations, although its magnitude is small (∼cm) (Deems and Painter, 2006). When subtracting the two measured DEMs, misclassification of corresponding point locations can occur, resulting in error in the final accumulation value (Deems and Painter, 2006).

Photogrammetry uses photographs to produce a series of DEMs that can be subtracted to find snow depth. Early attempts in the 1960s at applying this technique in snow covered areas suffered from poor vertical resolution due to overexposed photographs and the necessity of manual differencing (Nolan and others, 2015). Modern photography equipment, GPS, and software technology have allowed for an increase in accuracy and lowering of costs associated with photogrammetry (Nolan and others, 2015). Current photogrammetry software is able to determine a snow surface profile relative to stable, snow-free points within the mapped area (Farinotti and others, 2010). The photos collected for DEM creation can also be used to identify suspect changes in the snow pack (Nolan and others, 2015). Errors in photogrammetric measurements arise from sensor noise and poor lighting. Camera sensor noise is present in all digital photographs and its location changes from picture to picture. These erroneous pixels can be misinterpreted by the software as actual differences in height and thus lead to significant topographic noise, especially in steep mountainous terrain (Nolan and others, 2015). Additionally, having sufficient contrast in the photographed snow surface is critical for determining surface profile. Flat light conditions can reduce the resolution of the DEM or result in an absence of data in those parts of the photograph (Nolan and others, 2015). These effects can be avoided by waiting for better lighting.

4.4 Comparison of methods

The three methods of measuring accumulation differ in the extent of spatial information, collection techniques and costs, and processing needs. Spatial footprint is lowest for probing, which means that data must be interpolated. Although the actual measurement is simple and has relatively low uncertainty, the interpolation of points can lead to misrepresentation of spatial variability and significant errors that are not quantified. GPR provides continuous snow depth profiles, but interpolation is still needed between lines. Further, significant errors can arise from interpretation of layers in radargrams. DEM differencing has the advantage of allowing for the measurement of surface topography across the whole basin, but error can result when glacier dynamics affect surface height changes and in the conversion between snow height and to water equivalent.

Large differences in data collection time and cost also exist. Probing has low equipment costs but requires a large amount of human hours for ground-based measurements. GPR and DEM subtraction both require the use of aircrafts and expensive electronic equipment. However, these methods require lower logistical effort and data collection occurs quickly.

The three methods also have different data processing requirements. Simple statistical relations can be used for interpolating accumulation found by probing. However, GPR and DEM subtraction both require specialized software and knowledge of image processing methods, which increases the likelihood that misinterpretations of observations will occur. GPR has an advantage over DEM subtraction because it is not subject to elevation changes due to glacier dynamics and firn compaction. However, DEM subtraction has the advantage of more easily detecting the previous year's surface in the accumulation area and provides complete coverage of the study area (Sold and others, 2013).

In general, Machguth and others (2006) and Sold and others (2013) found good correlations between probing measurements, GPR, and DEM subtraction. However, (Sold and others, 2013) found that the methods did not always corroborate each other, particularly in crevassed areas and marginal regions. In crevassed areas, accumulation has large variation on small scales. The footprint of the GPR was usually too large to detect changes in snow depth and the movement of crevasses with time affected the lidar-derived snow depth. Marginal regions were misrepresented in the probing-derived profile because measurement was often not conducted in these areas (Section 4.1). This area also included the uppermost part of the glacier where wind erosion had a significant effect on accumulation.

Choice of measurement technique for a snow survey is therefore dependent on project specific needs. Resolution, cost, and equipment availability need to be considered when selecting the most appropriate method. To reduce errors and misrepresentation of measurements, multiple methods can also be applied (Machguth and others, 2006).

4.5 Temporally resolved methods

Temporally resolved methods measure accumulation continuously to provide a time series of snow accumulation. Usually, these methods involve relatively sparse point measurements so they do not represent spatial variability well. However, they are especially useful for identifying large snowfall events (rapid increases in accumulation) and wind erosion (gradual decreases in accumulation).

There are a number of methods of measuring SWE with time. Snow depth sensors, such as the SR50, measure the time between emission and return of an ultrasonic pulse (Ryan and others, 2008). As snow accumulates, the distance between the sensor and the snow surface decreases. SWE is then calculated using an assigned density. Snow pillows, which are large (3 m diameter) bladders filled with antifreeze solution, directly measure SWE (Archer and Stewart, 1995). As snow accumulates on the pillow, the weight of the snow forces an equivalent amount of the solution from the pillow to a standpipe. The height of the solution in the pipe is then recorded. Another method for measuring snow depth involves using multipath modulation of GPS signals (Larson and others, 2009; McCreight and others, 2014). Multipath modulation involves isolating GPS signals that are reflected from horizonal, planar reflectors, such as a snow surface. The distance between the geodetic GPS receivers and the reflection point will change during the accumulation season, thus recording changes in snow depth. This method allows for the measurement of average snow depth in a $\sim$1000 m^2 area around the antenna and an assigned density is then used to find SWE (McCreight and others, 2014).

5 Snow distribution on glaciers

While studies of snow distribution in alpine regions are plentiful (Clark and others, 2011, and sources within) there are comparatively few studies on the distribution of snow on glaciers. Although glaciers are often found in alpine environments, they present a different setting for accumulation. The freezing temperatures of glacier ice allow for snow to stick earlier than on the surrounding rocks, which can be above freezing especially in the early part of the accumulation season. Additionally, the surface of a glacier is often less steep than the surrounding peaks, which allows for snow to deposit more easily. The margin of the glacier can also accumulate snow from avalanches released from the surrounding peaks (Blöschl and others, 1991; Mott and others, 2008). Further, most glaciers do not support vegetation, which has significant effects on snow accumulation in many alpine basins (Pomeroy and others, 1999). Alford (1985) found that in open alpine areas with snow fields and small cirque glaciers there was a wide range of SWE over a relatively small range of elevation, while in the montane areas there was a strong relationship between elevation and SWE where SWE values increased rapidly with elevation. Since few studies have been done on this topic, it is difficult to say whether snow distribution on glaciers is fundamentally different than that of an alpine basin. This lack of snow variability quantification points to a significant gap in the literature.

Winther and others (1998) conducted one of the first accumulation variability studies on a glacier. A GPR system was used to measure snow depth along three large transects on Spitsbergen, Svalbard. It was found that the accumulation-elevation gradients varied considerably and that regional variability was large, with almost 50% more accumulation on the eastern coast and a minimum in accumulation in the inland locations. A number of subsequent accumulation studies in Svalbard have since been conducted. Pälli and others (2002) used GPR along longitudinal profiles of Nordenskjöldbreen and found 40-60% spatial variability over short distances. Grabiec and others (2011) compared snow distribution on four types of glaciers in Svalbard. It was observed that the land-terminating mountain glacier had a simple altitudinal gradient while the outlet glacier had a much weaker correlation and more wind-redistributed snow. It was thought that the orientation and shape of the glacier also had a significant impact on snow accumulation, with the glaciers that were oriented parallel to the dominant wind direction having stronger altitudinal gradients. Another glacier that was observed had no altitudinal gradient, so its distribution was determined by complex local conditions. The ice cap that Grabiec and others (2011) studied had all of these types of distributions in different areas.

Machguth and others (2006) conducted an airborne GRP survey of two adjacent glaciers in Switzerland. The lower part of the larger valley glacier showed a clear correlation between altitude and snow accumulation. The upper part of the glacier and the adjacent smaller glacier had no altitudinal trend and the fluctuations in depth were large. Additionally, the accumulation was 40% lower on the smaller glacier. The altitudinal trend is a well documented pattern and was thought to be a result of melt that occurred during warmer weather, which is more pronounced at lower elevations. Spatial variability of precipitation and redistribution of snow were believed to have resulted in the high spatial variability in higher parts of the study area. Since the majority of the precipitation events originated from one direction and the large glacier was on the lee side of a ridge, it experienced preferential deposition. Meanwhile, the smaller glacier was further along the storm track so it received less precipitation. Overall, Machguth and others (2006) showed that snow distribution on glaciers is not simply a function of altitude, which corroborated research done in other alpine catchments.

The most recent and comprehensive study of snow distribution on glaciers was done by McGrath and others (2015). This study focused on seven Alaskan glaciers of various sizes, orientations, and distances from the Pacific Ocean. McGrath and others (2015) found that SWE was highly variable (40% differences) on hillslope scales and especially large in the ablation area (which has a rough surface due to the presence of crevasses). The dominant control on SWE distribution was altitude, but multiple terrain parameters were needed to capture most of the variance — after elevation, wind exposure explained the most variance.

The study done by Walmsley (2015) contains the longest record of spatial distribution of snow accumulation. Walmsley (2015) analyzed 48 and 44 year records of two Norwegian glaciers for inter-annual stability in distribution patterns. It was found that snow accumulation is spatially heterogeneous yet it exhibits robust time stability in distributions. Reliability maps were then used to reduce the sampling scheme to one index site as well as a transect with 50 m elevation intervals for each glacier. Although winter balance reconstructions produced values within 0.15 m water equivalent, it was determined that a centreline transect underestimated winter balance. Transverse transects were therefore recommend as an addition to the sampling scheme to improve reliability. Additionally, several strongly irregular snow spatial distribution years were identified, which were inconsistent with the overall reduced sampling schemes.

The majority of studies that have examined snow distribution on glaciers have been done with either airborne or ground-based radar (e.g. Winther and others, 1998; Machguth and others, 2006; Grabiec and others, 2011; van Pelt and others, 2014; McGrath and others, 2015). In general, the radargrams provided valuable information but ground truthing by probing was always conducted. Gusmeroli and others (2014) also did a small GPR survey on an Alaskan glacier and found that GPR reflections were difficult to identify in areas of the glacier that had high debris content on the surface or in the upper part of the accumulation area. Sold and others (2013) did an extensive study that compared snow distribution values obtained by using probing, GPR, and DEM subtraction with lidar. All three methods showed an overall altitudinal trend but with significant small-scale variability (for a comparison of the three methods and their relative benefits, see Section 4.4). van Pelt and others (2014) used GPR and a coupled surface energy balance-snow model to examine accumulation variability. It was found that the terrain parameters such as slope and curvature resulted in preferential deposition. Additionally, van Pelt and others (2014) calculated that small-scale variability of snow accumulation had a negligible effect on the mean net mass balance in the accumulation zone and a negative impact of -0.09 m w.e. a^{-1} in the ablation area.

Dadic and others (2010) is the only study thus far that has examined snow distribution on glaciers using a dynamic model. This study specifically looked at the effect of wind on snow accumulation, and found that glacierized areas with the largest accumulation also experienced the lowest horizontal wind speeds and increasing downward wind velocity. Preferential deposition was highest (positive or negative) in troughs located close to steep slopes, where updrafts and down drafts led to decreased and enhanced deposition, respectively. In general, the wind speed was controlled by small-scale topography and had a significant impact on accumulation.

Fractal analysis has only been conducted in one alpine location by Arnold and Rees (2003). This study focused on small-scale (mm to 100 m) spatial variability and found that snow depth, surface albedo, and surface roughness were all self-similar over the range investigated. In particular, snow depth had a longer correlation length during winter than summer but in both cases, a constant variance was observed after approximately 50 m. Arnold and Rees (2003) suggest that future studies should measure snow depth along transects at least 100 m long with an intensive spacing of 1–2 m to identify the range at which variance becomes constant and that these transects should be completed every 1–2 km to determine whether this range differs across the glacier. McGrath and others (2015) plotted mean SWE difference with distance and found that four of the study glaciers exhibited a rapid increase in variability over the first $\sim$150 m and a slow increase in variability beyond but the three other study glaciers exhibited a gradual increase in variability over the entire range. Although this was not a detailed investigation of observed length scales, it points to potentially heterogenous nature of snow distribution length scales on glacier and the need for their increased measurement.

Although there are still few studies of snow distribution on glaciers, the work described above provides a good starting point for such investigations. Comparisons of variability between neighbouring glaciers and within a basin are both important areas of study.

6 Glaciers in the St. Elias Mountains

Snow data are generally sparse in mountain regions, especially those that are isolated from humans (Marcus and Ragle, 1970). The St. Elias Mountains (Figure 7) are one such area. These mountains contain the largest non-polar ice field and the longest valley glaciers outside of Greenland and Antarctica (Marcus and Ragle, 1970; Danby and others, 2003). Steep climatic gradients across the mountains create sharp changes in glacier cover and mass

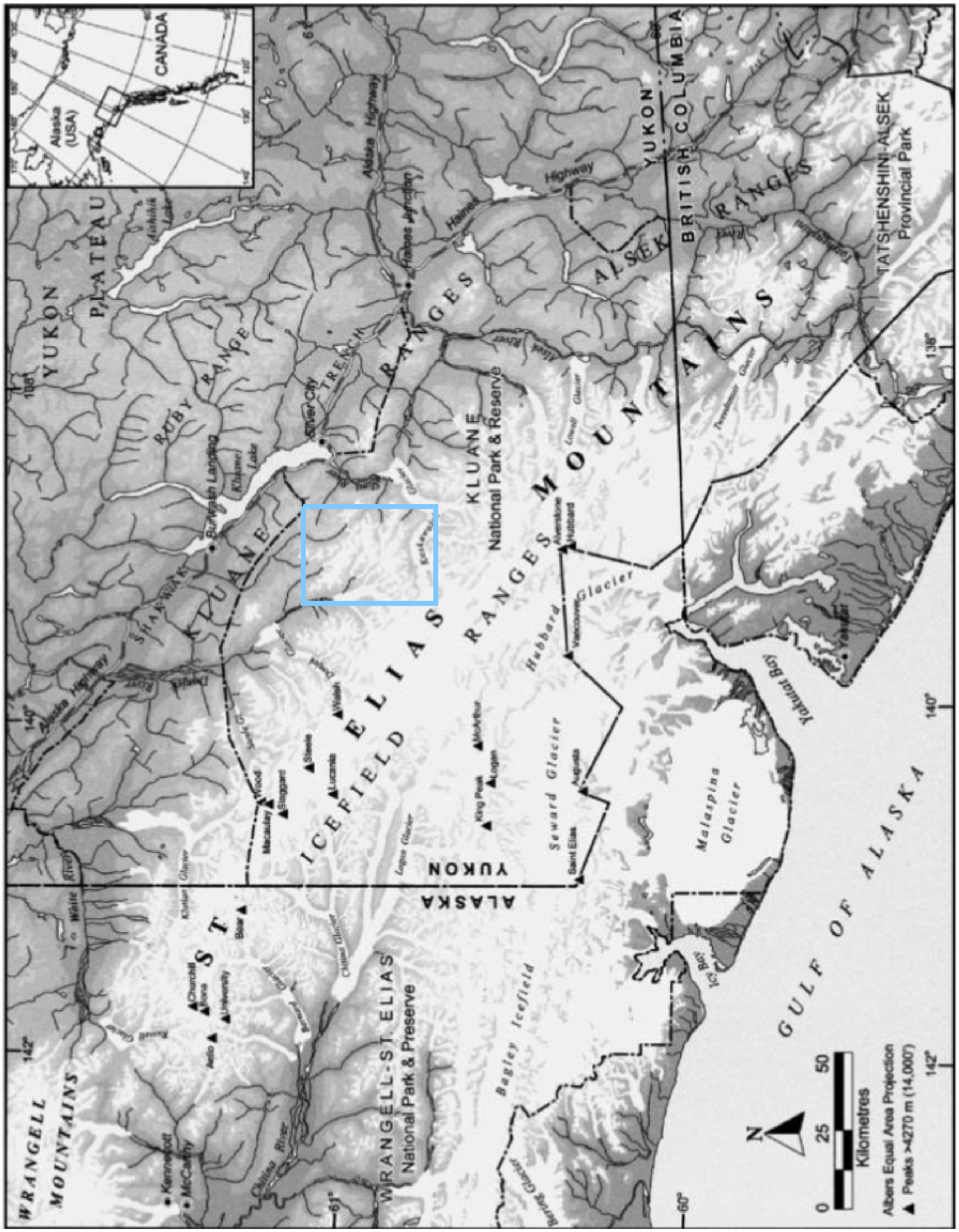


Figure 7: Map of the St. Elias Mountains and surrounding area. Figure taken from Danby and others (2003).

balance (Clarke and Holdsworth, 2002). This region currently has the most negative mass budget and is the largest contributor to sea-level rise in the world (Kaser and others, 2006; Gardner and others, 2013). Understanding how local glacier mass balance is affected by distribution of snow is therefore critical for accurate predictions of glacier response to a warming climate.

Research on snow distribution and glacier mass balance in the St. Elias is limited. The first significant investigations took place under Project Snow Cornice (Wood, 1948). Researchers looked at snow accumulation and ice formation as well as ice-mass thermal regime, density, and depth. Studies were conducted primarily on large glaciers such as the Kaskawalsh and Seward, and thus provided insights into large-scale accumulation patterns. This initiative was then followed by the Icefield Ranges Research Project (IRRP), which was established in 1961 (Danby and others, 2003). A number of subsequent long-term studies have been established in the St. Elias since IRRP (e.g. Clarke and others, 1984; Paoli and Flowers, 2009). Wheler and others (2014) determined the end-of-winter accumulation for the mass balance of a small alpine glacier in the Donjek Range. The study measured snow depth at a number of fixed stakes and used a multiple linear regression model — that accounted for slope, curvature, and elevation — to extrapolate these points and estimate basin-wide SWE. Arendt and others (2008) also briefly studied the mass balance of a number of glaciers in the St. Elias.

Two ice cores have been retrieved from the St. Elias Mountains. The first one was taken from the summit of Mt. Logan (5340 m) in 1980 and was 103 m long. The accumulation history in this core has been used to study the local (Holdsworth and others, 1991) and regional (Moore and others, 2002) climate history. A second core, called the Eclipse core, was taken from a site 45 km northeast of Mt. Logan, 2 km lower in altitude, with an accumulation almost five times as large (Wake and others, 2002). This core is 160 m and has also been used for studying local and regional climate history (Wake and others, 2002).

The weather in the St. Elias varies considerably over the range. The west side of the mountains is characterized by a cool, Marine West Coast climate due to the influence of the Pacific ocean, while the eastern side (just 250 from the ocean) is considered subarctic (Marcus and Ragle, 1970). Taylor-Barge (1969) studied the relationship between synoptic weather conditions and basin weather conditions across the St. Elias. It was observed that the mountains are oriented perpendicular to frequent and intense storms that originate in the ocean, which resultes in considerable interaction between weather and topography. When weather moves from the Gulf of Alaska, it is orographically lifted, which creates significant precipitation. If the front is perpendicular to the mountains, it can be deflected north or south, depending on the upper atmospheric flow. Fronts that are more or less aligned parallel to the mountains or very strong perpendicular fronts travel without deflection. The fronts can also stall on the west side of the mountains. Eventually, the fronts spill over the mountain divide (located to the west of the Kaskawalsh Glacier) and descend along the eastern side, which often results in decreased precipitation. This rain shadow is likely the major cause of the significant difference in accumulation and equilibrium line altitude (ELA) between the two sides of the mountains — the marine side has an ELA of $\sim$1100 m and the continental side has an ELA of $\sim$2100 m, while at the same elevation there is three times more accumulation on the marine side (Marcus and Ragle, 1970).

Although the characterization of synoptic conditions by Taylor-Barge (1969) is useful, it was conducted during the summer when weather conditions are considerably different than during winter. Taylor-Barge (1969) does note though that the weather patterns observed would likely be strengthened during winter because many of the spatial gradients are enhanced. The synoptic air masses present during the winter produce a strong temperature and moisture gradient, with warm, moisture-laden air coming from the Pacific Ocean and cold, dry air coming from the Mackenzie basin. These gradients would likely result in even more precipitation and stronger winds. The presence of a high pressure Arctic system could decrease the ability of low-pressure systems to pass over the divide, leading to a further enhancement of precipitation on the western side of the mountains.

The study done by Taylor-Barge (1969) also found weather effects on multiple scales. Synoptic conditions, including front movement, affected regional scale differences in weather and precipitation patterns. Watershed scale topography was responsible for differences in weather for nearby basins and affected wind speed and direction most significantly, while point scale topography had strong effects on snow accumulation. Orographic effects were found to be significant on all scales.

A study done by Pomeroy and others (1999) looked at snow mass balance in a non-glacierized alpine basin within the St. Elias. It was found that wind had a significant impact on the distribution of snow — up to 79% of the snow was redistributed from alpine areas to (primarily) hillsides, where accumulation was tripled. In the study basin, measured accumulation ranged from 54% to 419% of the actual snowfall. However, in a subsequent study year, which had two large wet snow events, the redistribution of snow was minimal and accumulation variability was much lower. The type of snow and how susceptible it is to wind effects therefore also plays a critical role in distribution. Additionally, areas within the basin can have different accumulation patterns throughout the winter. One area within the basin studied by Pomeroy and others (1999) had almost no redistribution (despite heavy winds) from the beginning of winter through to March. After this, all of the snow was lost even though additional accumulation events occurred in the basin. This could indicate a dependence of redistribution on weather conditions such as temperature, or that a critical depth was reached that allowed for redistribution to occur. Sublimation was also observed in the basin, but the amount of snow lost through this process could not be determined. Yet given that sublimation occurs several orders of magnitude faster when blowing snow is present and approximately 20% of winter days were observed to have blowing snow, it could have a significant impact.

There is clearly a strong need for a more comprehensive understanding of snow accumulation in the St. Elias Mountains. Although a few studies have examined accumulation, no studies have examined the distribution of snow and how it varies spatially. This is especially true of small alpine glaciers in the St. Elias Mountains, since most of the accumulation differences have been observed on large glaciers. It is likely that orographic lifting as well as wind redistribution and preferential deposition play major roles in determining accumulation on small alpine glaciers, so future studies should focus on the impact of these factors.

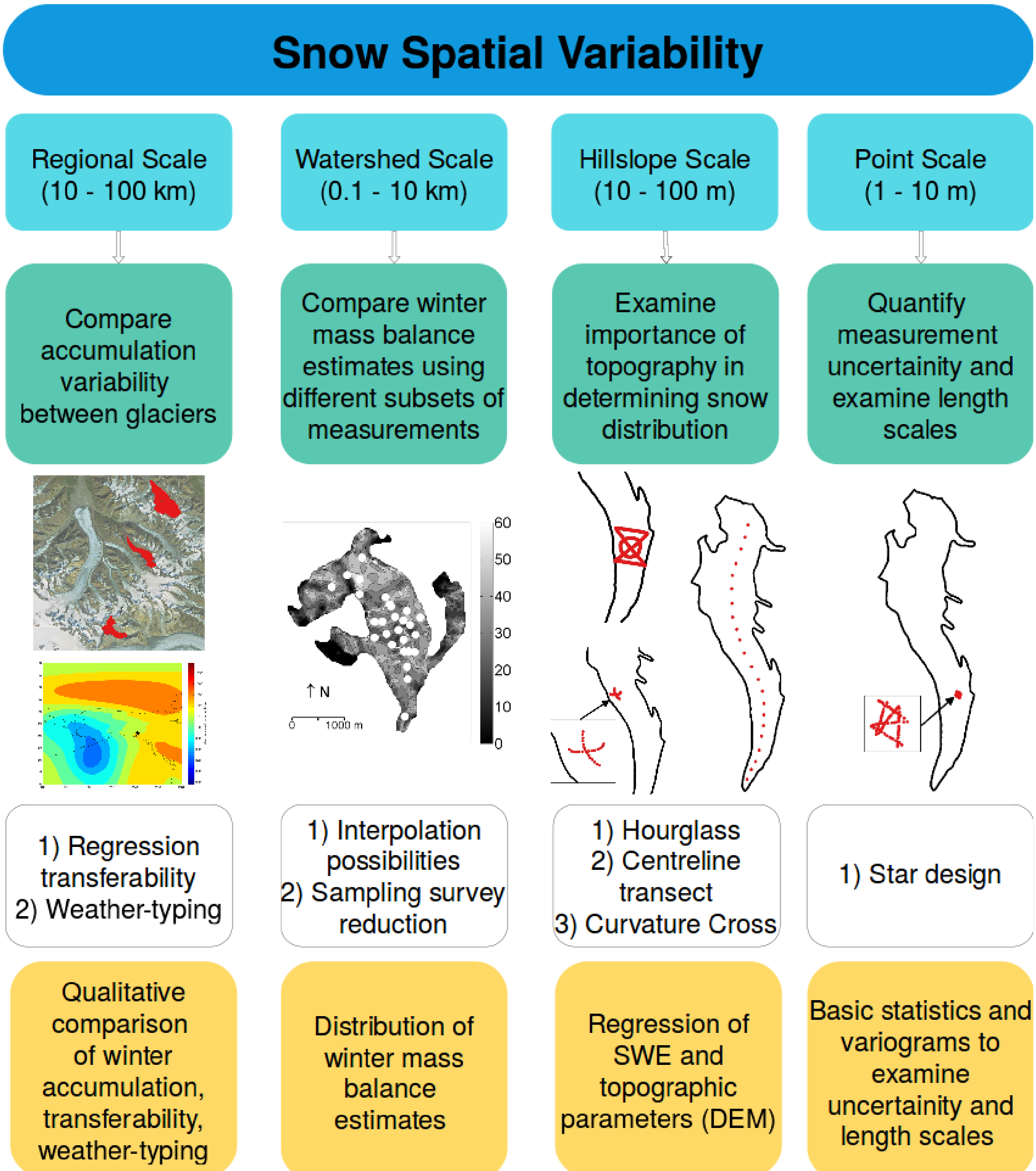


Figure 8: Visual representation of proposed research

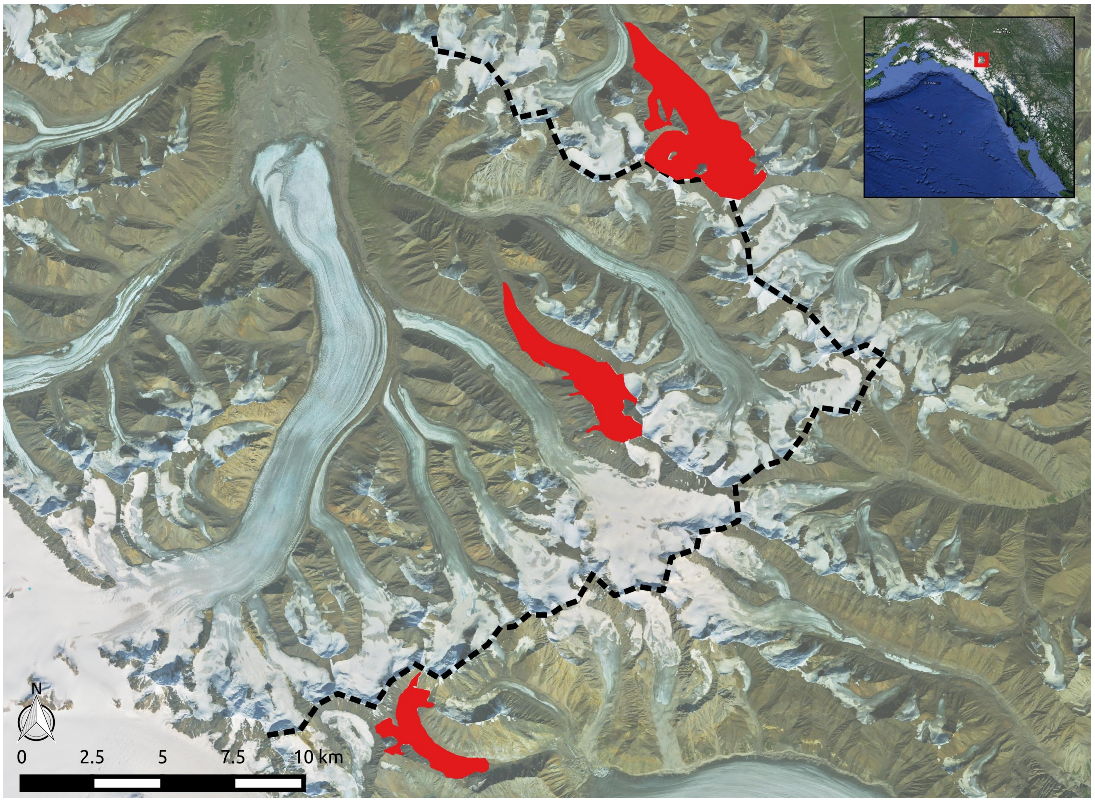
7 Proposed Research

The proposed research aims to examine the spatial and temporal variability of snow distribution in the St. Elias Mountains using direct measurements and statistical models (Figure 8). The spatial variability will be measured by conducting an extensive snow survey in May 2016. Accumulation will be measured using a combination of probing, firn coring, and snow pits on three alpine glaciers in the Donjek Range, located in the eastern part of the St. Elias Mountains. A combination of statistical techniques, including basic statistics, variograms, and regressions, will then be used to investigate spatial variability at multiple scales. Weather-typing will also be used to examine meso-scale weather conditions that affect precipitation distribution.

The three glaciers chosen for this study can be seen in Figure 9. Glaciers in the Donjek Range are unnamed but working names have been employed by Crompton and Flowers (In Press) and are adopted for this work. Glacier 4, Glacier 2, and Glacier 12 were selected. These glaciers were chosen because they are safe to travel on (with most of the glacier accessible on skis), have SPOT5 DEM coverage (highest resolution), and are a small enough (see Table 2) to allow for reasonable coverage using point measurement can be obtained. The selected glaciers also have similar orientations and one central glacier-filled valley (similar shape). Additionally, these glaciers are spread throughout the Donjek Range and are located increasingly further from the large-scale topographic divide (located at the head of the Kaskawalsh Glacier (Taylor-Barge, 1969)) allowing for the potential investigation of distance from mountain divide as an effect on regional snow distribution. The three glaciers are also located on different sides of the range-scale topographic divides, which run roughly from west to east in the southern area and from south to north in the eastern area and form an 'L' shape. Glacier 4 is located on the southern side of the first arm, Glacier 2 is located on the northern side of the first arm and the western side of the second arm, and Glacier 12 is located on the eastern side of the second arm. From anecdotal observations, these different configurations likely affect the winter balance of glaciers in the Donjek Range with glaciers on the southern side of the first arm receiving more snow.

7.1 Snow Measurement

Each glacier was divided into seven regions to determine sampling schemes and their locations. This allowed for snow depth and density be taken throughout a large portion of the glacier area. The glaciers were divided into three attitudinally distributed areas, with the upper area encompassing the accumulation zone, the middle area encompassing the upper half of the ablation zone, and the lower area encompassing the lower half of the ablation zone. During the snow survey, approximately one day will be spent taking measurements in each area. With these divisions, the ablation area will have a higher density of snow probe, firn core, and snow pit measurements than the accumulation area. Since measurements are more difficult and time consuming to make in the accumulation area, measurement efforts will be focused on the ablation area, where the snow pack is well defined. Higher density sampling in the ablation zone may also be warranted because McGrath and others (2015) found that the highest variability in SWE was in the ablation zone, especially in sections that had rough and crevassed ice.



(a) Donjek Range with study glaciers. From left to right: Glacier 4, Glacier 2, Glacier 12. Dashed line indicates local topographic divide, which forms an 'L' shape.

(b) Study glaciers with 50 m contour lines. From left to right: Glacier 4, Glacier 2, Glacier 12.

Figure 9: Study glaciers in the Donjek Range.

Table 2: Area, length, and elevation descriptors of three chosen glacier.

	Area (km^2)	**Length (km)**	**Elevation (m)**		
			Minimum	**Maximum**	**Mean**
Glacier 4	5.26	6.2	1573	2854	2321
Glacier 2	6.91	7.4	1906	3098	2472
Glacier 12	25.59	9.5	1775	3037	2434

The middle and lower areas were further divided into left, central, and right areas. The left and right areas encompass the margin of the glacier and were defined as being within 300 m of the ice edge (as mapped by the Randolph Glacier Inventory (RGI 5.0) (Pfeffer and others, 2014)), while the central area encompasses the central part of the glacier between the margins. This division was chosen because the snow distribution is likely to be affected by different processes at the margins compared to the centre. The margins are closer to steep rock slopes, which can affect the wind patterns and radiation at the snow surface. Avalanching can also affect snow distribution along the margins (Blöschl and others, 1991).

Probing will be done at point, hillslope, and watershed scales on the three glaciers (regional scale) in a number of patterns and orientations in an attempt to estimate snow distribution at multiple scales (Figure 10). For the point scale, the star design will be used to measure snow depth within 40x40 m grids (SPOT5 DEM resolution). For the hillslope scale, the hourglass with inscribed circle (referred to just as hourglass) will be used to measure snow depth within an entire elevation band of the glacier, with the upper and lower transects also serving as transverse transects. Cross transects will be completed opportunistically when areas of high curvature are present. A centreline profile and additional transverse transects will also be probed, which can be combined with hourglass measurements to estimate watershed-scale mass balance. A comparison of these measurements between the three study glaciers will allow for investigation of regional scale variability. Sampling schemes will be placed to ensure that SWE is measured in locations that represent a full range of topographic parameters found on each glacier.

For each glacier, the following series of snow depth measurements have been planned:

- Star design in a random location in thirteen regions of the ablation zone, with seven in the lower ablation area and six in the upper ablation zone (125 points each, random distance between 0.1 m and 1.5 m)
- Hourglass in two regions of the glacier ablation zone and span the entire width (random distance between 5 and 20 m). The upper and lower portions will thus also serve as cross-glacier transects.
- Transverse transects in three locations: one close to the terminus and two between the hourglass schemes (random distance between 5 and 20 m).
- A centre-line transect with a low resolution ($\sim$100m).

Snow density will be measured in various areas of each glacier. A firn corer will be used to determine bulk density of the snow column by extracting snow cores (known volume)

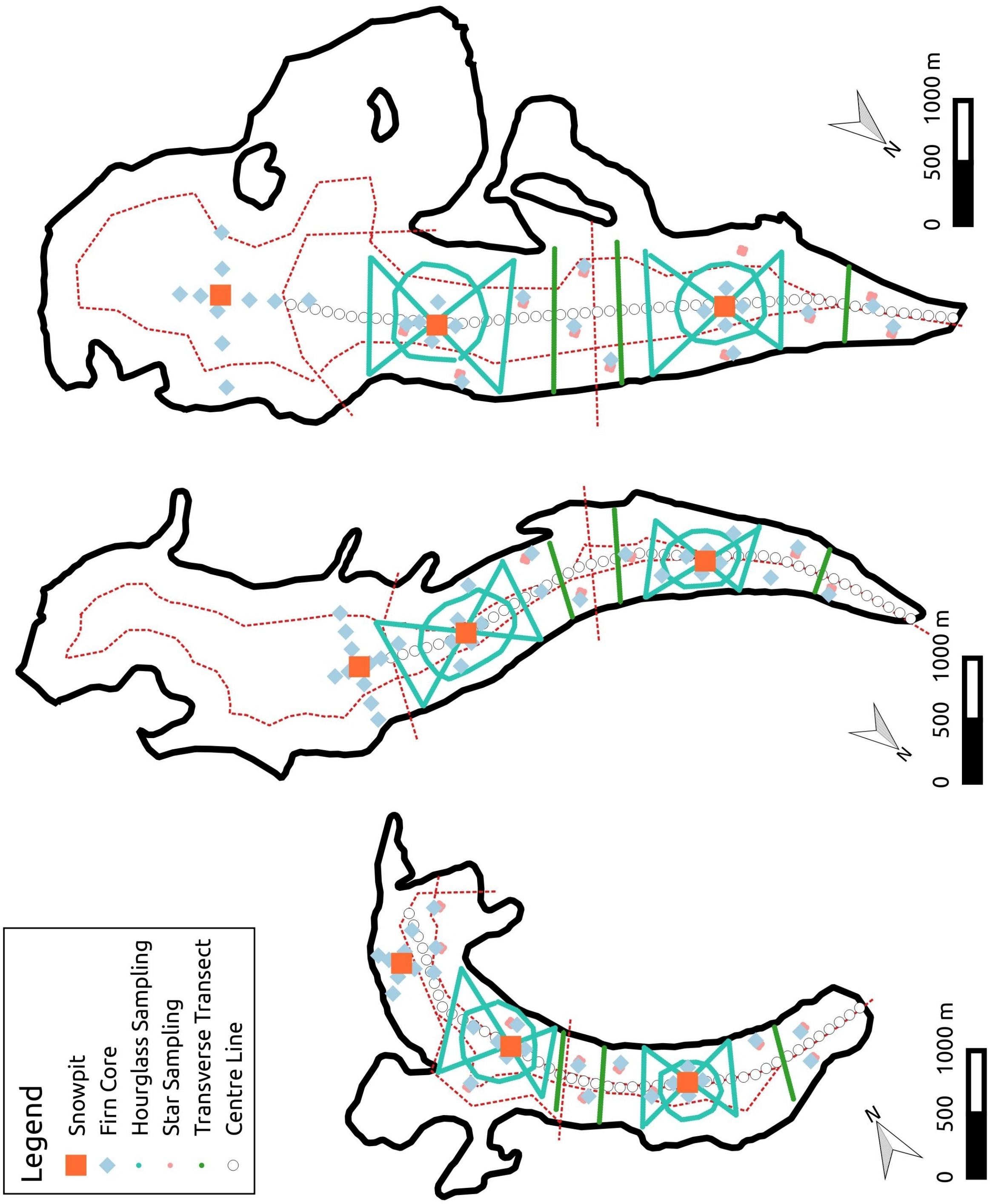


Figure 10: Planned sampling design on Glaciers 4, 2 and 12.

and weighting them. This instrument will be used because it allows for faster estimation of snow column density than snow-pit derived density. The measurements obtained with the firn corer will be compared to density measured in a snow pit using a wedge cutter, which is considered the most accurate way to measure density but is more time consuming (Østrem and Brugman, 1991). In areas with shallow snow packs, a Federal sampler (i.e. SWE tube) will be used for snow density measurements. Other snow properties observed in the snow pit, including stratigraphy and snow pack temperature will also be recorded.

For each glacier, the following series of snow density measurements have been planned (note that each measurement also serves as a depth measurement):

- Firn core at the edge of each star design located throughout the glacier.
- Four firn cores at the centre of each hourglass scheme.
- Approximately six firn cores in the transverse direction and four in the along-flow direction of the accumulation area.
- Snow pit measurements at the centre of each hourglass schemes and one in the accumulation aree (allows for comparison of firn core and snow-pit density)

7.2 Statistical Analysis

Spatial variability of snow accumulation at multiple scales will be investigated using a number of statistical techniques. Variograms will be constructed from all snow-depth measurements within a basin. This will allow for an investigation of relevant length scales and scale breaks that may address the autocorrelation of snow depth on glaciers and identify how much variance can be observed using our methods (variogram nugget). Identification of scale breaks could also help determine processes that drive snow distribution patterns at various scales.

At the point scale, the goal is to determine measurement error. To accomplish this, variability of snow depth within a grid cell will be quantified using basic statistical descriptors (e.g. standard deviation) of measurements from the star sampling schemes. It is valuable to estimate the uncertainty of a snow depth measurement at the resolution of the DEM used.

Regressions of SWE and DEM-derived terrain parameters will be done to investigate variability at the hillslope scale. Terrain parameters, including elevation, slope, curvature, "northness", aspect, and wind exposure/shelter, will be calculated from SPOT5 DEMs (40 x 40 m resolution). Both linear statistical methods (e.g. multiple linear regression) and non-linear methods (e.g. regression tree models) will be applied to best try to capture the relationships between geography and accumulation. The regression method that is able to capture the most variance will be used to interpolate between accumulation measurements to obtain a spatial distribution of SWE.

Variability at the watershed scale will be investigated by comparing winter balance estimates using different combinations of observed snow distribution and terrain parameters used for the regression. Different subsets of measured SWE values will be used to determine regressions and estimate winter balance, which will provide a range of possible winter-balance values. Comparing interpolated values of SWE from various regressions with measured values will examine the error when interpolating between point observations. This analysis is likely to provide insight into optimizing future glacier snow surveys.

At the regional scale, various analyses will be conducted. First, the transferability of regressions will be investigated by applying the regressions from one glacier to the other two and evaluating their performance.

Second, an investigation into the mesoscale weather conditions in the Pacific northwest during the 2015–2016 winter season will be done in an attempt to explain regional scale differences in accumulation. Weather conditions, including geopotential height, integrated vapour transport (IVT), and mean sea-level pressure, in the North Pacific will be obtained from ERA-Interim reanalysis data and statistical methods, including PCA and neural networks, will then be applied to find weather patterns. The frequency and types of patterns observed at the mesoscale, such as atmospheric rivers of IVT (Neiman and others, 2008; Roberge and others, 2009), could provide insight into moisture sources for the Donjek Range and how they affect glaciers at different distances from the large-scale topographic divide as well as various orientations to range-scale topographic divides.

Links with other studies that examined relationships between local precipitation and mesoscale and synoptic weather conditions will aid in understanding processes that affect regional snow distribution differences and in formulating hypotheses for future research. For example, Serreze and others (1995) and Lackmann and others (1998) found that vapour fluxes originating in the Pacific Ocean tend to peak in the lower troposphere (800 hPa) so the presence of a high mountain range is likely to block these sources. Additionally, Neiman and others (2008) observed that on the west coast, atmospheric rivers are responsible for twice as much precipitation as all storms and that atmospheric rivers have a tendency to increase SWE in the autumn and winter and decrease SWE in the spring. Furthermore, work done by Roberge and others (2009) found that atmospheric rivers, which result in intense cold-season precipitation events in the Yukon, can be clustered by the general area of moisture origins and that they are associated with certain synoptic conditions in the North Pacific. A study done by Matthews (2013) found that including weather type found from reanalysis data improved the simulation of daily ablation by up to 14% compared to a temperature-index model. Although this study did not examine precipitation, it is possible that investigation of weather conditions could also improve precipitation modelling through statistical down-scaling and account for inter-annual variability in precipitation. Preliminary work to correlate synoptic patterns to glacier-wide accumulation reveals significant correlations between patterns in the 750 hPa geopotential, found using a self organizing map (SOM), and winter balance on one glacier in the Donjek Range between 2007 and 2011, which is consistent with observations made by Taylor-Barge (1969) in the St. Elias Mountains.

7.3 Expected Outcomes

The goal of this project is to improve understanding of processes and parameters that affect snow accumulation on glaciers. Currently, there is little work that addresses length scales of SWE on glaciers. A large portion of the proposed work will therefore focus on identifying length scales within glacierized basins. This will be done by constructing and analysing variograms using data from star sampling schemes as well as hourglass, transverse, and centreline

Table 3: Research time line

Year	Date	Task
2016	May	Field work
	Late May, Early June	Organize and plot data, field report, preliminary correlation and geospatial statistics
	Late June	Glaciology Summer School
	July–August	Point scale analysis (variogram, autocorrelation)
	September – October	Hillslope scale analysis (regressions with topographic parameters from DEMs)
	October	Northwest Glaciologists' Meeting
	November – December	Watershed scale analysis (range of winter mass balance estimates)
2017	January – February	Weather pattern analysis (Integrated Vapour Transport)
	March – April	Regional scale analysis (compare computed relationships between glaciers)
	May	Write and submit paper for review
	June–July	Thesis writing and submission
	August	Defence and graduation

sampling schemes. From this, lag distances between 0.1 m and ~5 km can be plotted, which allows for identification of multiple relevant length scales. Fractal analysis could also be attempted, which may aid in the identification of scale breaks. Determining length scales is likely to provide insight into conditions and processes that affect snow distribution, including underlying topography such as crevasses as well as wind redistribution.

Mass balance models often estimate winter balance by using interpolation methods based on only a small number of observations. It is likely that these accumulation estimates are poor representations of mass balance input. The proposed work will attempt to improve how these estimates are made by determining uncertainty in SWE measurement and interpolation, identifying topographic parameters that affect snow distribution, and examining the transferability of statistical models within a mountain range. While these aspects have been investigated in mountain environments, there are few studies that look at winter balance estimates on glaciers. By examining various statistical techniques used when calculating regressions between observed accumulation and topographic parameters (e.g. linear regression versus binary tree) as well as different methods for interpolating between measurements (e.g. kriging), a range of winter mass balance estimates can be complied. Various subsets of measured SWE can also be used in these regressions and interpolations to examine the effect of sample size and measurement location on accumulation estimates. Valuable insights into optimizing snow survey design are likely to be gained from this process. Transferability of regressions between nearby basins will also be addressed, which will aid in determining winter mass balance across mountain ranges. Examining potential impacts of mesoscale weather systems (weather typing) on mountain range snow accumulation will also contribute to improving current understanding of conditions that may affect observed snow distribution.

There is a need for a multi-scale investigation of snow accumulation on glaciers. The proposed work aims to characterise the snow distribution by identifying relevant length scales and by investigating the uncertainty, techniques, and controlling factors associated with calculating spatial patterns of SWE. A paper that summarizes these methods and results will be submitted by the end of the project. A projected time line for field work and data analysis is presented in Table 3.

8 Summary

Snow accumulation plays a central role in alpine hydrology and has a prominent impact on glacier mass balance. In mountainous regions, accumulation is highly variable on point, hillslope, watershed, and regional scales. The contribution of accumulation to glacier mass balance is controlled mainly by the distribution of snow. Processes such as orographic lifting, preferential deposition, and wind redistribution, all arising from the interaction of atmospheric conditions and topography, strongly affect snow distribution. Statistical models have been used to relate meteorological and topographic variables to snow accumulation in order to better understand the effects of these processes. These models rely on accurate measurement of snow distribution, which can be achieved by determining SWE from snow density and depth. Results from previous studies of accumulation on glaciers have shown large spatial variability at many scales and a dependence on multiple processes that affect snow distribution.

Accumulation in the St. Elias Mountains is poorly understood, largely because the glaciers are remote. There is a need to quantify snow accumulation in this region and how it varies both between glaciers and within glacierized basins. The proposed study would be the first within the St. Elias Mountains to examine accumulation variability at the point, hillslope, watershed, and regional scale. Well-established methods will be applied to measure accumulation variability. These measurements will be used to investigate measurement uncertainty and relevant length scales, the role of topography in determining snow distribution, optimizing winter balance measurement, and the transferability of statistical relationship as well as regional differences in accumulation across a range. This comprehensive approach to examining spatial patterns in snow distribution on glaciers will contribute to the current understanding of processes and parameters that affect winter balance variability on glaciers.

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