Snow distribution on glaciers

Alexandra Pulwicki
Earth Science SFU

December 21, 2015

Abstract

Alpine glacier mass balance input is controlled mainly by the spatial distribution of snow. Processes such as orographic lifting, preferential deposition, and wind redistribution strongly affect snow distribution in alpine regions, and arise from interactions between atmospheric conditions and steep topographic gradients. Dynamic and statistical models have been used to relate meteorological and topographic variables to snow accumulation so that the effects of these processes can be better understood. These models rely on accurate measurements of snow distribution. Snow probing, ground penetrating radar, and digital elevation model subtraction are all methods of obtaining such measurements that vary considerably in their spatial resolution, cost, and ease of use. Although snow distribution on glaciers has not been studied extensively, results from previous studies have shown there to be large spatial variabilities at multiple scales due to a number of processes. Accumulation distribution in the St. Elias Mountains is particularly poorly understood, largely because the glaciers of this region are remote and difficult to access. There is therefore a need to understand snow accumulation in this region and how it varies, both between glaciers and within glacierized basins.

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 anthropogenic uses (O'Neel and others, 2014; Barnett and others, 2005), but also for rates of global sea-level rise (Gardner and others, 2013). It is therefore critical to thoroughly understand the spatial distribution of snow on glaciers. Achieving such an understanding however 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 glacialized 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 literature review 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. The final 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.

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 (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 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;

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

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). 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 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 amount 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) notes 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 modifies mean wind (and snow particle) velocities, and thus can 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).

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).

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, 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.

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) has 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 global 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 conditions. 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). This method assumes 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).

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 achieve this, 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), 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) 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 while the model developed by Anderton and others (2004) explained 70–80% of the variance. 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.

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 variability can be limited. This method assumes that variability between cells is larger than within cell variability. This 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, the use of terrain parameters as proxies does not provide meaningful insight into relevant processes. This is important when attempting to predict distribution or variability of variables in a different climate or location.

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.

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 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). 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 interpolation of *in situ* point measurements, GPR survey 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 and knowledge. 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

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 years 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 probes within a 7 m radius as representative of a test site in the ablation zone.

In the ablation zone snow depth is easy to measure because the snow surface at the end of the melt season is well defined (McGrath and others, 2015). In the accumulation zone however, this same snow surface is determined by a combination of snow transformations — including melt, settling, and redistribution by wind — that are influenced by meteorological and topographical factors (Grunewald and others, 2010). This results in a heterogeneous snow surface that can have dense or compacted snow, ice lenses, or exposed firn. Furthermore, altering of the snow pack during the accumulation season (especially in the early and late parts) from warm weather or rain events can result in water refreezing in the snowpack (Sold and others, 2014). 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). As a result, snow pits and firn cores are often 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 extrapolated. This is often done using a statistical regression that accounts for grid cell values of slope, aspect, curvature, and susceptibility to wind redistribution (e.g. Wheler

and others, 2014; McGrath and others, 2015). This generates a linear equation that is site specific and is able to predict 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 1). 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 taken, 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. This means that 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.

Figure 1: Digging a snowpit in the accumulation area of Haig Glacier, Rocky Mountains

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

2)(Machguth and others, 2006; McGrath and others, 2015). The resulting radargram (e.g. Figure 3) 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) describes this process in five steps: (i) acquisition of GPR and probing data (ground trotting), (ii) calculation of snow density and radar velocity, (iii) calculation of snow thickness and resulting SWE, (iv) applying a correction to measured accumulation based on ground truth data, and (v) using 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).
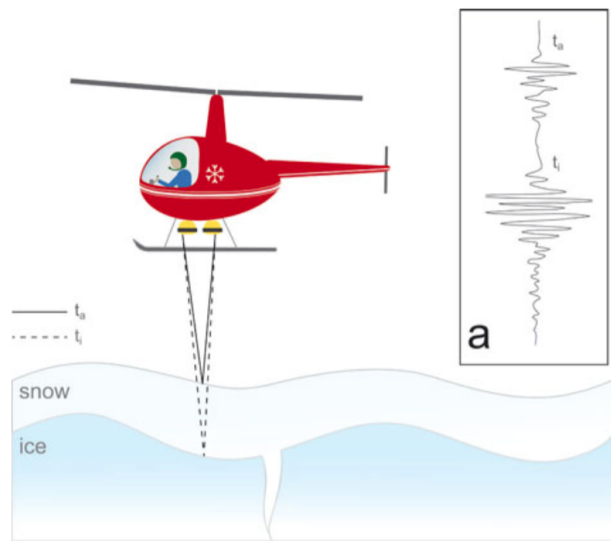

Figure 2: Schematic of a helicopter-bore radar snow survey. The travel time for the signal to interact with the snow surface is snow as a solid line and the signal travel time of the interaction with the ice is shown as a dashed line. Together, these values can be used to determine snow depth. The inset (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. GPR also has the major advantage that processes that affect snow height, including melt, settling, and glacier dynamics, do not affect the measurement accuracy (Machguth and others, 2006). Furthermore, the ability to fly over steep or inaccessible regions means that all areas of a glacier can be measured. GPR surveys also need be conducted only once to gain depth observations, which makes data collection fast and able to cover large areas (Machguth and others, 2006).

A large limitation of GPR is the difficulty 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 mislabeling (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 field conditions, survey design and intention (Sold and others, 2013), which hampers the reproducibility of surveys.
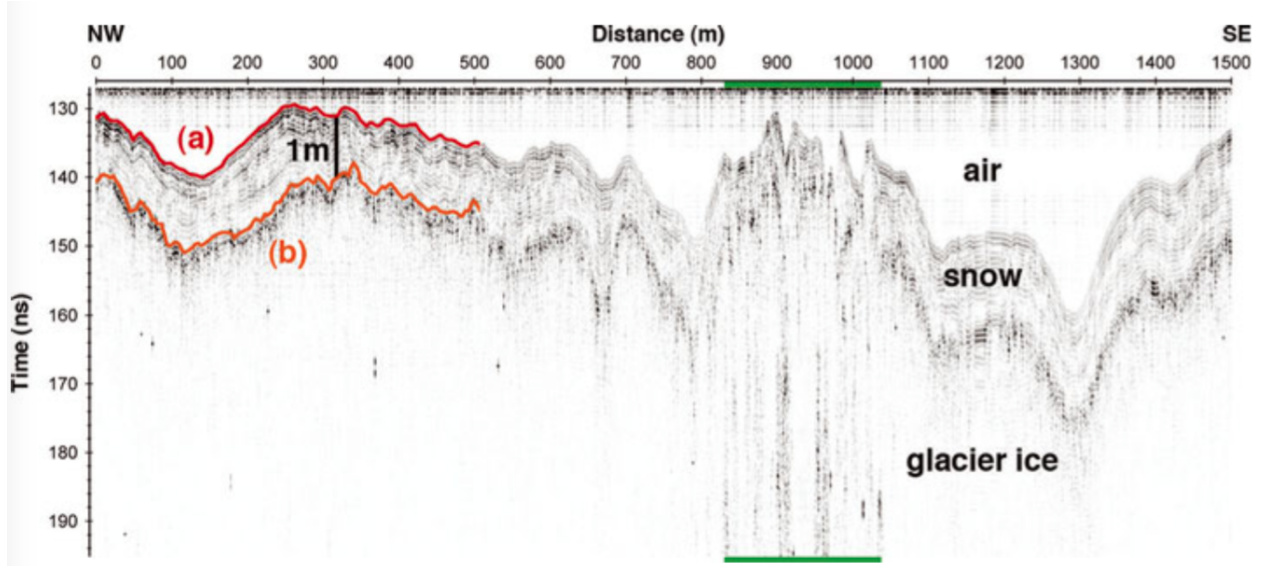
Figure 3: Radargram from the accumulation area of Findelengletscher, Valais, Switzerland. (a) The reflection at the air-snow interface. (b) The reflection at the snow-ice interface. Figure taken from Sold and others (2013).

4.3 DEM subtraction

Digital elevation model (DEM) subtraction involves taking two detailed surface profile 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

Figure 4: Schematic of airborne LiDAR system geometry. Scan angle (θ), height (h), and swath width (SW) are shown. Figure from Deems and Painter (2006).

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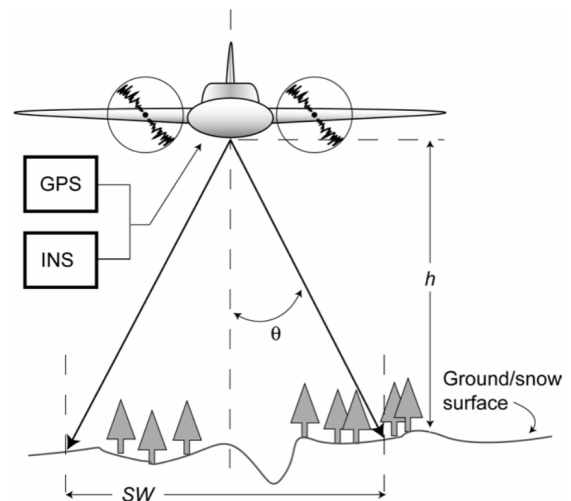
be made to account for these discrepancies but they rely on *in situ* measurement of snow depth, knowledge of long-term mass balance, or information about the vertical displacement of ice from GPS towers (Sold and others, 2013).

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 4) 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 resolution is lowest for probing, which means that data must be interpolated. Although the actual measurement is simple and has 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 profile across the whole basin, but error can result when glacier dynamics affect surface height changes.

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 have low human hour requirements 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 (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 ~ 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, glaciers do not support vegetation, which has significant effects on snow accumulation in many alpine basins (Pomeroy and others, 1999). 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 majority of studies that have examined snow distribution on glacier have been done with either airborne or ground-based GPR (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 subtracting 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 slowest 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,
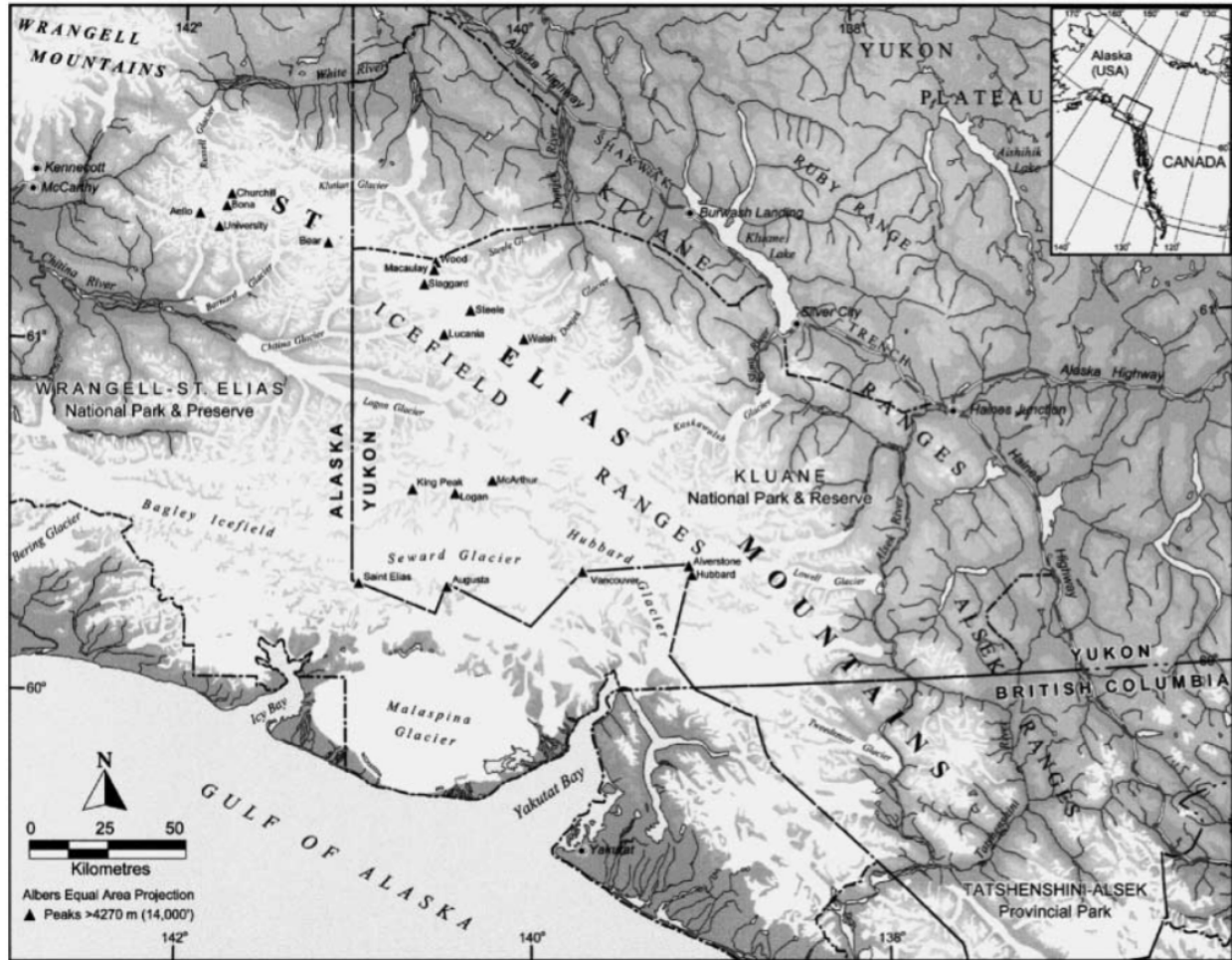
Figure 5: Map of the St. Elias Mountains and surrounding area. Figure taken from Danby and others (2003).

the wind speed was controlled by small-scale topography and had a significant impact on accumulation.

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 5) 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 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 Kaskalwalsh 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. The mechanisms associated with surge type glaciers were studied by Garry Clarke and his colleagues on Trapridge Glacier Clarke and others (1984). More recently, Gwenn Flowers began studying smaller alpine glaciers in the eastern part of the St. Elias Mountains (e.g. Paoli and Flowers, 2009). Wheler and others (2014) determined the end-of-winter accumulation for the mass balance of these glaciers by recording snow depth at a number of fixed stakes and using a multiple linear regression model — that accounted for slope, curvature, and elevation — to extrapolate these points and find 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 ∼1100 m and the continental side has an ELA of ∼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; watershed scale topography was responsible for differences in weather for nearby basins and affected wind speed and direction most significantly; and 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 the majority of 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.

7 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. This is due to complex interactions between dynamic meteorology and dramatic topography that result in steep gradients in temperature and wind. Processes such as orographic lifting, preferential deposition, and wind redistribution are a result of these interactions and they affect where snow will accumulate.

The distribution of snow can be described using dynamic and statistical models. Dynamic models employ equations that describe physical processes to determine local atmospheric conditions and from that, calculate where snow is likely to accumulate. These models are spatially and temporally transferable and provide insight into relevant processes. However, they are computationally expensive and require a large array of input data. Statistical models, on the other hand, are computationally simple and usually require only basic data. However, their use is restricted to areas with long historical records, and their transferability limited by the fact that the parameters they employ are proxies for processes.

Measuring snow accumulation can be done a number of different ways. Snow probing is the simplest method and requires little equipment and knowledge. However, the measurements acquired are usually quite sparse, require a long time to complete, and cannot be done in areas that are difficult to access. GPR is another common way of measuring snow depth and has been widely applied in snow accumulation studies. This technique provides continuous lines of snow depth data and is not affected by surface changes due to glacier dynamics. Processing and interpreting data can be difficult though, especially in the accumulation area. DEM subtraction allows for a fully distributed measurement of accumulation but complex topography can create significant errors in the interpretation of distance. Factors such as cost, processing needs, and extent of spatial information should all be considered when determining which method to apply.

Snow accumulation on glaciers is a relatively new area of research. The few studies that have examined this topic have found high variability in accumulation both within a glacierized basin and between nearby glaciers. Perhaps the most challenging aspect of studying accumulation on glaciers is acquiring data with sufficient spatial extent and resolution to come to meaningful conclusions. Determining processes relevant to snow distribution also presents considerable difficultly since dynamic models are sparse and statistical models must employ proxy parameters. In the St. Elias Mountains, little is known about the variability of snow distribution on alpine glaciers. No studies have been done to examine differences in accumulation between glaciers within the same range and accumulation distribution within those glaciers. Given the importance of snow for local water availability and glacier mass balance, there is clearly a need to better understand its distribution within the St. Elias Mountains.

References

Anderson BT, McNamara JP, Marshall HP and Flores AN (2014) Insights into the physical processes controlling correlations between snow distribution and terrain properties. *Water Resources Research*, **50**(6), 4545–4563 (doi: 10.1002/2013WR013714)

Anderton S, White S and Alvera B (2004) Evaluation of spatial variability in snow water equivalent for a high mountain catchment. *Hydrological Processes*, **18**(3), 435–453 (doi: 10.1002/hyp.1319)

Archer D and Stewart D (1995) The installation and use of a snow pillow to monitor snow water equivalent. *Water and Environment Journal*, **9**(3), 221–230 (doi: 10.1111/j.1747-6593.1995.tb00934.x)

Arendt AA, Luthcke SB, Larsen CF, Abdalati W, Krabill WB and Beedle MJ (2008) Validation of high-resolution GRACE mascon estimates of glacier mass changes in the St Elias Mountains, Alaska, USA, using aircraft laser altimetry. *Journal of Glaciology*, **54**(188), 778–787 (doi: 10.3189/002214308787780067)

Barnett TP, Adam JC and Lettenmaier DP (2005) Potential impacts of a warming climate on water availability in snow-dominated regions. *Nature*, **438**(7066), 303–309 (doi: 10.1038/nature04141)

Blöschl G (1999) Scaling issues in snow hydrology. *Hydrological Processes*, **13**(1415), 2149–2175

Blöschl G and Kirnbauer R (1992) An analysis of snow cover patterns in a small alpine catchment. *Hydrological Processes*, **6**(1), 99–109 (doi: 10.1002/hyp.3360060109)

Blöschl G, Kirnbauer R and Gutknecht D (1991) Distributed snow melt simulations in an alpine catchment. *Water Resources Research*, **27**(12), 3171–3179

Busuioc A, Chen D and Hellström C (2001) Performance of statistical downscaling models in GCM validation and regional climate change estimates: application for Swedish precipitation. *International Journal of Climatology*, **21**(5), 557–578

Clark MP, Hendrikx J, Slater AG, Kavetski D, Anderson B, Cullen NJ, Kerr T, Örn Hreinsson E and Woods RA (2011) Representing spatial variability of snow water equivalent in hydrologic and land-surface models: A review. *Water Resources Research*, **47**(7) (doi: 10.1029/2011WR010745)

Clarke GK, Collins SG and Thompson DE (1984) Flow, thermal structure, and subglacial conditions of a surge-type glacier. *Canadian Journal of Earth Sciences*, **21**(2), 232–240 (doi: 10.1139/e84-024)

Clarke GK G K and Holdsworth G (2002) Glaciers of the St. Elias mountains. *US Geological Survey professional paper*, (1386J)

Dadic R, Mott R, Lehning M and Burlando P (2010) Wind influence on snow depth distribution and accumulation over glaciers. *Journal of Geophysical Research: Earth Surface (2003–2012)*, **115** (doi: 10.1029/2009JF001261)

Danby RK, Hik DS, Slocombe DS and Williams A (2003) Science and the St. Elias: an evolving framework for sustainability in North America's highest mountains. *The Geographical Journal*, **169**(3), 191–204 (doi: 10.1111/1475-4959.00084)

Deems JS and Painter TH (2006) Lidar measurement of snow depth: accuracy and error sources. In *Proceedings of the International Snow Science Workshop*, 1–6

Elder K, Rosenthal W and Davis RE (1998) Estimating the spatial distribution of snow water equivalence in a montane watershed. *Hydrological Processes*, **12**(1011), 1793–1808 (doi: 10.1002/(SICI)1099-1085(199808/09)12:10/11¡1793::AID-HYP695¿3.0.CO;2-)

Erxleben J, Elder K and Davis R (2002) Comparison of spatial interpolation methods for estimating snow distribution in the Colorado Rocky Mountains. *Hydrological Processes*, **16**(18), 3627–3649 (doi: 10.1002/hyp.1239)

Farinotti D, Magnusson J, Huss M and Bauder A (2010) Snow accumulation distribution inferred from time-lapse photography and simple modelling. *Hydrological Processes*, **24**(15), 2087–2097 (doi: 10.1002/hyp.7629)

Fowler H, Blenkinsop S and Tebaldi C (2007) Linking climate change modelling to impacts studies: recent advances in downscaling techniques for hydrological modelling. *International Journal of Climatology*, **27**(12), 1547–1578 (doi: 10.1002/joc.1556)

Gardner AS, Moholdt G, Cogley JG, Wouters B, Arendt AA, Wahr J, Berthier E, Hock R, Pfeffer WT, Kaser G and others (2013) A reconciled estimate of glacier contributions to sea level rise: 2003 to 2009. *Science*, **340**(6134), 852–857 (doi: 10.1126/science.1234532)

Grabiec M, Puczko D, Budzik T and Gajek G (2011) Snow distribution patterns on Svalbard glaciers derived from radio-echo soundings. *Polish Polar Research*, **32**(4), 393–421

Grunewald T, Schirmer M, Mott R and Lehning M (2010) Spatial and temporal variability of snow depth and ablation rates in a small mountain catchment. *Cryosphere*, **4**(2), 215–225 (doi: 10.5194/tc-4-215-2010)

Gusmeroli A, Wolken GJ and Arendt AA (2014) Helicopter-borne radar imaging of snow cover on and around glaciers in Alaska. *Annals of Glaciology*, **55**(67), 78–88 (doi: 10.3189/2014AoG67A029)

Hanssen-Bauer I and Førland E (1998) Long-term trends in precipitation and temperature in the Norwegian Arctic: can they be explained by changes in atmospheric circulation patterns? *Climate Research*, **10**(2), 143–153

Harrison W (1986) Seasonal accumulation and loss of snow from a block mountain catchment in central Otago. *Journal of Hydrology (NZ)*, **25**(1), 1–17

Herron MM and Langway Jr CC (1980) Firn densification: an empirical model. *Journal of Glaciology*, **25**, 373–385

Hock R (2005) Glacier melt: a review of processes and their modelling. *Progress in Physical Geography*, **29**(3), 362–391 (doi: 10.1191/0309133305pp453ra)

Holdsworth G, Fogarasi S and Krouse H (1991) Variation of the stable isotopes of water with altitude in the Saint Elias Mountains of Canada. *Journal of Geophysical Research: Atmospheres (1984–2012)*, **96**(D4), 7483–7494 (doi: 10.1029/91JD00048)

Kaser G, Cogley JG, Dyurgerov MB, Meier MF and Ohmura A (2006) Mass balance of glaciers and ice caps: consensus estimates for 1961–2004. *Geophysical Research Letters*, **33**(19) (doi: 10.1029/2006GL027511)

Kerr T, Clark M, Hendrikx J and Anderson B (2013) Snow distribution in a steep mid-latitude alpine catchment. *Advances in Water Resources*, **55**, 17–24 (doi: 10.1016/j.advwatres.2012.12.010)

Kidson JW and Thompson CS (1998) A comparison of statistical and model-based downscaling techniques for estimating local climate variations. *Journal of Climate*, **11**(4), 735–753 (doi: 10.1175/1520-0442(1998)011¡0735:ACOSAM¿2.0.CO;2)

Larson KM, Gutmann ED, Zavorotny VU, Braun JJ, Williams MW and Nievinski FG (2009) Can we measure snow depth with GPS receivers? *Geophysical Research Letters*, **36**(17) (doi: 10.1175/2007JTECHA947.1)

Lehning M, Löwe H, Ryser M and Raderschall N (2008) Inhomogeneous precipitation distribution and snow transport in steep terrain. *Water Resources Research*, **44**(7) (doi: 10.1029/2007WR006545)

Liston GE and Elder K (2006) A distributed snow-evolution modeling system (SnowModel). *Journal of Hydrometeorology*, **7**(6), 1259–1276 (doi: 10.1175/JHM548.1)

Liston GE and Sturm M (1998) A snow-transport model for complex terrain. *Journal of Glaciology*, **44**(148), 498–516

Machguth H, Eisen O, Paul F and Hoelzle M (2006) Strong spatial variability of snow accumulation observed with helicopter-borne GPR on two adjacent alpine glaciers. *Geophysical Research Letters*, **33**(13) (doi: 10.1029/2006GL026576)

Marchand WD and Killingtveit Å (2005) Statistical probability distribution of snow depth at the model sub-grid cell spatial scale. *Hydrological Processes*, **19**(2), 355–369 (doi: 10.1002/hyp.5543)

Marcus MG and Ragle RH (1970) Snow accumulation in the Icefield Ranges, St. Elias Mountains, Yukon. *Arctic and Alpine Research*, 277–292

McCreight JL, Small EE and Larson KM (2014) Snow depth, density, and SWE estimates derived from GPS reflection data: Validation in the western US. *Water Resources Research*, **50**(8), 6892–6909 (doi: 10.1002/2014WR015561)

McGrath D, Sass L, O'Neel S, Arendt A, Wolken G, Gusmeroli A, Kienholz C and McNeil C (2015) End-of-winter snow depth variability on glaciers in Alaska. *Journal of Geophysical Research: Earth Surface*, **120**(8), 1530–1550 (doi: 10.1002/2015JF003539)

Mellor M (1974) *A review of basic snow mechanics.* US Army Cold Regions Research and Engineering Laboratory

Molotch N, Colee M, Bales R and Dozier J (2005) Estimating the spatial distribution of snow water equivalent in an alpine basin using binary regression tree models: the impact of digital elevation data and independent variable selection. *Hydrological Processes*, **19**(7), 1459–1479 (doi: 10.1002/hyp.5586)

Moore G, Holdsworth G and Alverson K (2002) Climate change in the North Pacific region over the past three centuries. *Nature*, **420**(6914), 401–403 (doi: 10.1038/nature01229)

Mott R, Faure F, Lehning M, Löwe H, Hynek B, Michlmayer G, Prokop A and Schöner W (2008) Simulation of seasonal snow-cover distribution for glacierized sites on Sonnblick, Austria, with the Alpine3D model. *Annals of Glaciology*, **49**(1), 155–160 (doi: 10.3189/172756408787814924)

Nolan M, Larsen C and Sturm M (2015) Mapping snow-depth from manned-aircraft on landscape scales at centimeter resolution using Structure-from-Motion photogrammetry. *The Cryosphere Discussions*, **9**, 333–381 (doi: 10.5194/tcd-9-333-2015)

O'Neel S, Hood E, Arendt A and Sass L (2014) Assessing streamflow sensitivity to variations in glacier mass balance. *Climatic change*, **123**(2), 329–341 (doi: 10.1007/s10584-013-1042-7)

Pälli A, Kohler JC, Isaksson E, Moore JC, Pinglot JF, Pohjola VA and Samuelsson H (2002) Spatial and temporal variability of snow accumulation using ground-penetrating radar and ice cores on a svalbard glacier. *Journal of Glaciology*, **48**(162), 417–424 (doi: 10.3189/172756502781831205)

Paoli LD and Flowers GE (2009) Dynamics of a small surge-type glacier using one-dimensional geophysical inversion. *Journal of Glaciology*, **55**(194), 1101–1112 (doi: 10.3189/002214309790794850)

Pomeroy J, Hedstrom N and Parviainen J (1999) The snow mass balance of Wolf Creek, Yukon: effects of snow sublimation and redistribution. *Wolf Creek, Research Basin: Hydrology, Ecology, Environment, edited by: Pomeroy, JW and Granger RJ*, 15–30

Ryan WA, Doesken NJ and Fassnacht SR (2008) Evaluation of ultrasonic snow depth sensors for US snow measurements. *Journal of Atmospheric and Oceanic Technology*, **25**(5), 667–684 (doi: 10.1175/2007JTECHA947.1)

Smith RB and Barstad I (2004) A linear theory of orographic precipitation. *Journal of the Atmospheric Sciences*, **61**(12), 1377–1391 (doi: 10.1175/1520-0469(2004)061¡1377:ALTOOP¿2.0.CO;2)

Sold L, Huss M, Hoelzle M, Andereggen H, Joerg PC and Zemp M (2013) Methodological approaches to infer end-of-winter snow distribution on alpine glaciers. *Journal of Glaciology*, **59**(218), 1047–1059 (doi: 10.3189/2013JoG13J015)

Sold L, Huss M, Eichler A, Schwikowski M and Hoelzle M (2014) Recent accumulation rates of an alpine glacier derived from firn cores and repeated helicopter-borne GPR. *The Cryosphere Discussions*, **8**(4), 4431–4462 (doi: 10.5194/tcd-8-4431-201)

Taylor-Barge B (1969) The summer climate of the St. Elias Mountain region. Technical report, DTIC Document

van Pelt J, Ward J, Pettersson R, Pohjola VA, Marchenko S, Claremar B and Oerlemans J (2014) Inverse estimation of snow accumulation along a radar transect on nordenskiöldbreen, svalbard. *Journal of Geophysical Research: Earth Surface*, **119**(4), 816–835 (doi: 10.1002/2013JF003040)

Wake CP, Yalcin K and Gundestrup NS (2002) The climate signal recorded in the oxygen-isotope, accumulation and major-ion time series from the Eclipse ice core, Yukon Territory, Canada. *Annals of Glaciology*, **35**(1), 416–422 (doi: 10.3189/172756402781817266)

Wheler BA, MacDougall AH, Flowers GE, Petersen EI, Whitfield PH and Kohfeld KE (2014) Effects of temperature forcing provenance and extrapolation on the performance of an empirical glacier-melt model. *Arctic, Antarctic, and Alpine Research*, **46**(2), 379–393 (doi: 10.1657/1938-4246-46.2.379)

Widmann M, Bretherton CS and Salathé Jr EP (2003) Statistical precipitation downscaling over the Northwestern United States using numerically simulated precipitation as a predictor. *Journal of Climate*, **16**(5), 799–816 (doi: 10.1175/1520-0442(2003)016¡0799:SPDOTN¿2.0.CO;2)

Wilby RL and Wigley T (2000) Precipitation predictors for downscaling: observed and general circulation model relationships. *International Journal of Climatology*, **20**(6), 641–661 (doi: 10.1002/(SICI)1097-0088(200005)20:6¡641::AID-JOC501¿3.0.CO;2-1)

Winstral A, Elder K and Davis RE (2002) Spatial snow modeling of wind-redistributed snow using terrain-based parameters. *Journal of Hydrometeorology*, **3**(5), 524–538 (doi: 10.1175/1525-7541(2002)003¡0524:SSMOWR¿2.0.CO;2)

Winther J, Bruland O, Sand K, Killingtveit A and Marechal D (1998) Snow accumulation distribution on spitsbergen, svalbard, in 1997. *Polar Research*, **17**, 155–164

Wood WA (1948) Project "Snow Cornice": the establishment of the Seward Glacial research station. *Arctic*, 107–112

Zorita E and Von Storch H (1999) The analog method as a simple statistical downscaling technique: comparison with more complicated methods. *Journal of Climate*, **12**(8), 2474–2489 (doi: 10.1175/1520-0442(1999)012¡2474:TAMAAS¿2.0.CO;2)