Spatial and temporal variability of the snow environment in the Western Canadian Arctic

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Spatial and temporal variability of the snow environment in
the Western Canadian Arctic

By

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THESIS

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Abstract

Snow cover in the Western Canadian Arctic is a significant input to the hydrological mass balance, it produces shelter and habitat for animals and humans, and supports interactions with vegetation and climate. The Arctic-tundra snow cover is greatly impacted by wind erosion, redistribution and deposition of snow during high wind events over the winter months. As a result, the end of winter snow cover is characterised by significant small-scale (on the order of a few meters) spatial variations in snow cover depth, density, and thus snow water equivalent (SWE), and runoff. Future climate related changes to snow cover depth and density will have significant consequences to the hydrology, ecology and climatology of the Arctic. This thesis reviews a multi-year record of snow studies in Siksik Creek, a sub-catchment of Trail Valley Creek (TVC) located in the western Canadian Arctic. TVC is located in the taiga-tundra transition zone, dominated by tundra, but with shrub and forest patches. Wind speed, snow depth, temperature and snowfall were measured over the full annual cycle, while end of winter snow accumulation was measured through ground based snow surveys and aerial imagery from an unmanned aerial system (UAS). The snow cover of TVC is highly influenced by its vegetation, topography and climate. Therefore, as the climate and vegetation continues to change in the coming decades, it is expected that there will be great changes in snow cover and, consequently, impacts on water resources, animal habitats and vegetation. The results from this thesis will provide information on improved methods to measure the snow environment, and the data sets needed to test snow models required for understanding future changes in snow.
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Chapter 1: Introduction

1.1 Research Background

Over the past few decades, Arctic amplification has resulted in warming across the entire Arctic at a rate almost twice as fast as the global average (Screen and Simmonds, 2010), causing changes to all aspects of the cryosphere, including sea, lake and river ice, snow cover and permafrost (Graversen et al., 2008). Such rapid changes are also occurring across Canada’s three northern territories (Yukon, Northwest Territories, and Nunavut), with strong increases in temperature (Stewart et al., 1998; Déry, 2001; Woo et al., 2008; DeBeer et al., 2016), significant, but spatially variable, increases and decreases in precipitation, decreasing floating ice, and melting permafrost. All Arctic ecosystems are of course dependent on the cryosphere, and many of the inhabitants of the Canadian Arctic also rely on the cryosphere for hunting and fishing, recreation, employment and traditional ways of life (Vaux et al., 2013). It is well known that these changes to the Arctic climate and cryosphere will continue for centuries or even millennia (Williams et al., 2007; Collins et al., 2013), but the details of the past and future changes to the cryosphere at the local scale of importance to the northern residents is poorly known.

Snow is a fundamental component of the Arctic and sub-Arctic cryosphere (Derksen et al., 2009), with significant implications to, and interactions between, vegetation, permafrost, hydrology, ecology and society (Quinton and Marsh, 1998; Pomeroy et al., 1999; Ménard et al., 2014b), as well as feedbacks to the regional and global climate system. The primary snow properties of interest to hydrologists, ecologists, climatologists and remote sensors are snow depth, density, and snow water equivalent (SWE), and the spatial distribution of these across the Arctic landscape. Unfortunately,
due to the small number of weather stations across the Arctic as compared to more temperate regions and difficulties in measuring snowfall and sampling the Arctic snow cover, our knowledge of past changes in snowfall and snow cover is very limited (Fleming et al., 2000). Airborne and satellite remote sensing, as well as modelling studies, have reduced some of this uncertainty at regional to pan-Arctic scales and have clearly demonstrated considerable decreases in snow cover duration (Brown et al., 2010; Derksen and Brown, 2012; Liston et al., 2002). However, to the author’s knowledge, there are few studies of multi-decadal changes in the Arctic snow environment at the local scale, that include changes in snowfall and spatially distributed snow depth, density and SWE, and no such studies in the Canadian Arctic. More detailed description of advances and limitations in observations and modelling of the Arctic snow cover follow.

A significant deficiency in our understanding of Arctic snow cover is due to the large errors in measuring snowfall in the high wind speed environment of the Arctic. For example, there are over 50 different types of precipitation gauges in use around the world (Sevruk and Klemm, 1989), each with its own systematic error of precipitation measurement (Goodison et al., 1998). In attempts to better understand this issue, the World Meteorological Organization (WMO) has produced multiple intercomparison reports on solid precipitation measurements, comparing various precipitation gauges and wind shields to obtain catch ratios dependent on wind and/or temperature for each gauge (Goodison et al., 1998; Sevruk et al., 2009). Goodison et al. (1998) provided details on these, and more recently, there has been multiple solid precipitation correction studies on weighing precipitation gauges at automated sites (Smith, 2007; Rasmussen et al., 2012; Pan et al., 2016). A key result from these studies is what WMO considers the DFIR
(double fenced intercomparison reference) gauge to be the reference standard, while all others suffer from undercatch during high winds and there is still no new technology that solves this issue.

The impact of blowing snow is of special importance in the Arctic where tundra dominates (Pomeroy et al., 1997; Déry and Yau, 2002), with blowing snow entraining large amounts of snow into the lower atmosphere during periods of high wind. Across the Arctic, the number and magnitude of high winds, and therefore blowing snow events, varies greatly from year to year, with some years with frequent and extreme wind storms, and other years with only a few such events. The three key effects of blowing snow are to redistribute snow across the landscape, to greatly increase the rates of sublimation over the winter and to create wind slab and depth hoar layers within the snowpack.

Sublimation can be sufficiently large, significantly reducing the average watershed SWE (Li and Pomeroy, 1997; Liston et al., 2016; Nolan et al., 2015). Pomeroy and Male (1992) and Pomeroy and Essery (1999) were among the first to measure sublimation using eddy covariance methods over short periods of time during blowing snow events, and clearly showed that sublimation during blowing snow can be extremely large. However, technical limitations have limited the ability to run eddy covariance over entire winter periods in the Arctic as needed to determine total winter sublimation. Instead, some investigators (Benson, 1982; Liston and Sturm 1998; Pomeroy and Essery, 1999; Liston and Sturm, 2004) have estimated total winter snow sublimation from blowing snow models to suggest that sublimation can remove 15-50% of total precipitation. As a result, blowing snow across open tundra areas with extensively flat terrain, will typically have an average SWE that is considerably less than winter precipitation.
The second implication of extensive blowing snow in the Arctic, is that the spatial variability in SWE is extremely large over very small horizontal scales. Snow depth variability varies from snow bedforms (Filhol and Sturm, 2015) which are typically from a few centimeters to a few meters across, with a coefficient of variation varying from 0.16 to 0.34 (Pomeroy et al., 1998; Pohl and Marsh, 2006), to terrain controlled snow drifts that form on lee slopes, in stream channels and along lake edges. Drifts on slopes are controlled by slope aspect, angle, and extent, while those along stream channels and lake edges are controlled by the channel width and slope angle for example. In hilly tundra terrain, the hilltops may have snow only a few centimeters in depth, while lee slopes may accumulate snow many meters in depth – these areas of deeper than average snow are typically called snow drifts. Liston et al. (1995) and Sturm et al. (2001b) documented the size and location of major slope controlled drifts up to 6 m in depth in Arctic Alaska and showed that they typically occur in the same locations from year to year. With drifts coming in all shapes and sizes, Rees et al. (2014) related snow properties to the drift properties. They analyzed the slope and aspect of the drift and related that to observed snow depths, densities and SWE. Their results confirmed that slopes have a great effect on SWE and snow depths, with steeper slopes capturing approximately 3 to 4 times greater snow depth and SWE than flat land. Woo and Sauriol (1980) described drifts in stream channels up to 6 m in depth. Aspect plays a role in the year to year variation of snow depths and SWE on terrain controlled drifts, depending on the varying dominant winter wind direction. Although such snow drifts have been described extensively in the literature, surprisingly they are poorly defined quantitatively. For example, consider a profile of snow accumulation along a transect that crosses a hilltop, a steep lee slope, and
a flat area at the base of the slope. Typically, the snow depth will be low across the hilltop, will gradually increase in depth as you progress down the hill, reaching a maximum at some point, and then decreasing in depth across the flat area at the base of the slope. How deep does the snow need to be on the slope to be classified as a drift? Which area of the slope would be classified as a drift? How would you map snow drifts across a watershed? Does the drift area evolve over the winter or should it be mapped as an area of some basic properties such as slope angle? Few studies have specifically addressed these definitions, and as a result, comparing information on snow drifts from various sources is difficult. This uncertainty also challenges estimating basin SWE from landscape based snow surveys as it is extremely difficult to properly define drift areas and to properly sample the very large number of drifts across the landscape.

In addition to terrain controlled drifts, drifts also form in patches of tall vegetation (shrub and tree) (Pohl and Marsh, 2006; Rees et al., 2014). Drifts that form in shrub patches on flat terrain, are dependent primarily on patch size and extent, as well as stem diameter, shrub height and the elastic modulus of the stem (Ménard et al., 2014a) which controls shrub bending (Marsh et al., 2010) and possible burial by the snowpack. In this case, saltation and sublimation are able to resume, limiting the further growth of the snowpack in shrub patches (Ménard et al., 2014a). Ménard et al. (2014a) hypothesized that bending of the shrubs occurs at near 0°C when snowfall is wet so that it can attach itself to the branches and weigh them down. They also suggested that bending is possible in isolated shrub patches during high wind events. Variations in shrub bending by shrub species, size or age are not well known (Ménard et al., 2014a). Drifts located in tree patches on flat terrain are impacted in similar ways as to those in shrub patches, but they
are not susceptible to bending and burial as are some shrubs. Snow up to 2 m in thickness has been reported in shrub patches that are not bent by the snow, but the maximum snow depth in patches and the controlling factors have not been described in the literature. For regions with large numbers of shrub patches, large drifts form on a small area of the windward edge of the patch, but there are of course interactions between slope, stream, lake and vegetation patch drifts. For example, in the low Arctic, slope drifts are often underlain by large patches of shrubs and trees, enhancing drift formation. In contrast, shrub patches that occur up-wind of slopes may limit the availability of snow for redistribution, thus limiting slope drift formation. Such interactions between shrub/tree patches and terrain controlled drifts have not been well studied, but are required to understand the impact of future changes in climate on snow cover.

Terrain and vegetation drifts with deep snow depths, and high snow density, typically hold a very high percentage of total watershed SWE. For example, Pomeroy et al. (1997) estimated that for a tundra watershed in the western Canadian Arctic, 27% of the total basin SWE was contained in drifts (defined as slopes greater than or equal to 9º) covering only 8% of the watershed, while only 43% of the total snow was contained on the windswept tundra areas that covered 70% of the basin area. Sturm et al. (2001b) suggested similar amounts of total watershed snow in drifts for Toolik River and Meade River sites in Alaska. This large volume of snow in drifts has a significant impact on snowmelt runoff, as the shallow snow covering a large percentage of the basin melts over the first few days of thaw and often will not contribute to runoff as the meltwater infiltrates and freezes within the frozen active layer. While large drifts, covering only a small portion of the watershed, will have a delayed contribution to runoff, but will
contribute to runoff for weeks or even months (Marsh and Woo, 1981) resulting in a streamflow regime that is similar to a glacierized watershed. Although described in the literature, few if any models have been shown to simulate these effects.

Snow drifts have important ecological effects by providing important protective habitat during the long cold winters for many plants and animals. This includes 400 species of vascular plants and lichens and mosses (Wohl, 2015), polar bears (Liston et al., 2016), lemmings and siksiks (Winner, 2003; Wohl, 2015), and impact foraging for all other Arctic animals, including caribou and muskox for example (Larter and Nagy, 2001). Drifts that are sufficiently deep to exist into the summer, provide habitat for animals such as caribou during the summer as the coolness limits insect harassment (Downes et al., 1986; Anderson and Nilssen, 1998; Hagemoen and Reimers, 2002).

Sampling this highly variable Arctic snow cover is extremely challenging. To date, most studies rely on extensive sampling using simple depth probes and snow cores, with sampling in one of two ways. First, snow probing is carried out along short transects within typical landcover types including windblown tundra, slope drifts (often defined by a slope angle of greater than or equal to 9° (Tabler, 1975; Woo et al., 1983; Pomeroy et al., 1997)), shrub patches, and channel and lake edges. The average SWE within each landcover type is then weighted by the area of each landcover type (Rees et al., 2014) to determine the average watershed SWE. Although studies have suggested that this method provides reasonable estimates of basin average SWE (Pomeroy et al., 1998; Bowling et al., 2003; Pohl and Marsh, 2006; Marsh et al., 2008), it is in fact very difficult to determine the accuracy of this method due, primarily, to the difficulty in sampling snow located in drifts across a watershed. Sampling extensive uplands with relatively uniform
snow cover likely has very low errors, while estimating the snow contained in numerous, but small in area, drifts is extremely challenging, and likely prone to very large, but poorly known, errors. This method doesn’t consider edge effect, that is the effect that discontinuities around the borders of landcover types have on snow distribution (i.e. edges of shrub or tree patches). Therefore, this method only provides average snow data for each terrain type, not fully spatially variable data at sufficiently high resolution required to: i) falsify spatially distributed hydrological models (including blowing snow and melt components), ii) provide distributed SWE as input to fully distributed hydrological models, or iii) provide data required to better understand the links between snow drifts, vegetation, and permafrost, and iv) between snow and animals.

A second sampling approach is to measure depth and density along long transects (Derksen et al., 2009; Derksen et al., 2014), and assume that these transects cover the full range of snow depths and are representative of the terrain. There are advantages of this approach for remote sensing applications (Derksen et al., 2014; Leinss et al., 2014), but for hydrological needs, it is unlikely that this approach quantifies average watershed SWE with the needed accuracy and does not provide appropriate maps of SWE required for model validation or testing. Although both methods likely do a reasonable job of sampling SWE within terrain types where the SWE does not vary greatly over small distances (for example extensive upland tundra), it seems very likely that they do a very poor job of sampling areas where the SWE is both very high, and varies greatly over small areas (i.e. drifts).

The deficiencies in snow survey methods noted above, have encouraged the development of numerous remote sensing techniques to better map snow cover depth and
SWE at local to regional scales. These include the following. Airborne and satellite based Synthetic Aperture Radar (SAR) (Leinss et al., 2014) has been applied to measure SWE at a resolution of 91 by 91 m. This method has significant promise, but is currently limited by the influence of snow structure on the signal. Satellite based microwave sensors (Dietz et al., 2012) have been used for many years to provide maps of SWE over grids of approximately 25 km. However, this method is limited by the insensitivity to deep snow in drifts that cover small areas. Recently, terrestrial and airborne LiDAR (Light detecting and ranging, Hopkinson et al., 2004; Deems et al., 2006; Grünewald et al., 2010; Hopkinson et al., 2012) has been applied to map snow depth across local to regional areas. Another recent advance is to combine aerial photogrammetry from both manned aircraft (Nolan et al., 2015; Bühler et al., 2015) and Unmanned Aerial Systems (UAS) (Vander Jagt et al., 2015; De Michele et al., 2016; Harder et al., 2016) with Structure-from-Motion (SfM; Westoby et al., 2012). This method has shown significant promise. Although the use of manned aircraft currently has advantages, the use of multi-rotor and fixed-wing UASs is increasing due to their semi-automation, repeatability, cost and improving accuracy (Colomina and Molina, 2014; De Michele et al., 2016). Both have their advantages and disadvantages, multi-rotor UASs possess greater stability in higher winds, however lack battery duration which minimizes aerial extent, whereas fixed-wing UASs are less stable in higher winds but can cover larger areas, and therefore are very useful on the spatial and temporal variability aspect. Currently all types of UAS available to researchers have a short range and typically can only cover up to a few km² within a few hours of flying, due to lack of battery power. It is expected that this range will increase dramatically in the coming years.
In addition to field observations and remote sensing, snow modelling has provided important insight into Arctic snow accumulation. Liston et al. (1993; 1995; 1998; 2002; 2006; 2016) have carried out various modelling studies using the SnowModel. These studies have generally focused on modelling drifts (Liston et al., 2016) and accumulation on landscapes with numerous shrub patches (Liston et al., 2002). In contrast, Pomeroy et al. (1988; 1993; 1995; 1997; 2002; 2006; 2008) have focused on the hydrological implications of blowing snow events. Primarily modelling snow erosion and accumulation in Hydrological Response Units (HRUs) needed for larger scale hydrological modelling using such models as the Prairie Blowing Snow Model (PBSM; Pomeroy, 1988; Pomeroy et al., 1993) and the Distributed Blowing Snow Model (DBSM; Pomeroy et al., 1997; Essery et al., 1999; Essery and Pomeroy, 2004). In general, these studies have shown considerable success in estimating snow cover in HRUs and in slope drifts, but little ability to simulate snow cover at high resolution across the Arctic landscape. Liston et al. (2016) has demonstrated the ability to predict the size of snow drifts of known location using a fine horizontal resolution of 2.5 m by 2.5 m, however have not demonstrated this ability at a watershed scale.

As suggested above, there remain many gaps in our understanding of the Arctic snow cover. Measuring key aspects of the snow environment (snowfall, snow on the ground, sublimation, and distribution of snow in drifts) using ground observations or remote sensing is very limited. In addition, there are significant uncertainties in the ability of snow models, including limitations in modelling wind flow around slopes and vegetation patches, and properties of the snow surface which effect the entrainment of snow into the atmosphere when wind speeds increase. As a result of these significant
uncertainties, to the author’s knowledge no studies have yet been able to complete a rigorous snow mass balance for an Arctic watershed. The following chapter outlines some of the challenges in improving our understanding of Arctic snow, and the research that will be carried out in this study to improve our understanding of some aspects of the Arctic snow environment.

1.2 Research Objectives

This research will focus on improving two key aspects of our understanding of the Arctic snow environment. First how has snowfall and snow on the ground changed over the last 60 years, and second have the snow survey approaches used over this period of record provided accurate estimates of snow accumulation across a tundra watershed (located 50 km north of the Inuvik weather station) where large numbers of deep snow drifts occur. Such an analysis is only possible due to a unique snow data set that includes 60 years of snowfall from the Inuvik Airport weather station (Inuvik-A), 26 years of spatially distributed snow surveys, and recently studies considering new observation methods, including UAS for mapping snow depth.
Chapter 2: Snow Environment of the Western Canadian Arctic

Abstract

This paper considers changes in snowfall and snow on the ground over a period of 60 years and uses recent advances in snow accumulation observations to better characterize the variability of end-of-winter snow depth, density and snow water equivalent (SWE) in the western Canadian Arctic. A combination of traditional snow surveys and novel photogrammetric techniques with an unmanned aerial system (UAS) were conducted to compare high resolution data sets for peak snow accumulation over complex shrub tundra landscapes. These data sets were applied to quantify the spatial distribution of SWE at high resolution across the study basin with the help of three spatial snow densification methods. Results show a decreasing trend in snowfall and maximum snow depth over the 60-year period as well as a decreasing trend in SWE collected from snow surveys for 1991-2017. End of winter snow depth root mean square errors (RMSE) of 49.0 and 27.6 cm were obtained between the manual and UAS snow depths for the two study years of a 1 km² watershed. Spatial variability in snow depth, density and SWE were then analyzed at a fine spatial scale (1 m x 1 m) in order to quantify the percentage of SWE captured in drift formations. This information is required for both improved runoff modelling, and to consider past and future changes to tundra landscapes.

Keywords: snow, SWE, drift, UAS, spatial variability
2.1 Introduction

Snow cover is a keystone component of the Arctic environment, with significant implications to, and interactions between vegetation, permafrost, hydrology and ecology (Quinton and Marsh, 1998; Pomeroy et al., 1999; McFadden et al., 2001; Essery and Pomeroy, 2004; Woo and Marsh, 2005; Ménard et al., 2014b), as well as feedbacks to the climate (Sturm et al., 2001a). Recent studies have suggested significant changes in the snow environment across the Arctic with increasing/decreasing snowfall and decreases in the spring snow cover area (Foster et al., 2008; Brown et al., 2010) and snow water equivalent (SWE) (Liston and Hiemstra, 2011) over the last few decades. Such changes have also been observed in the western Canadian Arctic (Marsh et al., 2002; Lantz and Kokelj, 2008; Marsh et al., 2010), along with later onset of fall snowfall and earlier onset of snowmelt. These changes in snow cover are of course primarily due to the combined effects of changes in air temperature and in precipitation. However, changes in snow cover are also impacted by changes in wind and blowing snow, as well as the effect of increasing shrubs across the tundra landscape (Sturm et al., 2001a; Lantz et al., 2010; Loranty and Goetz, 2012). Shrub patches can trap blowing snow, and as a result store a greater quantity of SWE compared to nearby open tundra. Trapping of snow in shrub patches reduces the fetch available for blowing snow, therefore reducing both sublimation during blowing snow (Pomeroy and Gray, 1995; Liston et al., 2002; Marsh et al., 2010), and the size of drifts formed on slopes for example.

Our understanding of changes in Arctic snow is limited by two significant issues. First, unlike other climatic variables such as temperature with over a hundred years of record and measurements from many locations, observations of changes in snow cover
have only been monitored relatively recently in the Arctic (Woo and Marsh, 1978), and as a result, the length of record is very limited in most locations and only in very few locations have measurements been consistent over a few decades. Second, there are significant errors in our ability to measure all major components of the Arctic snow environment, including: snowfall (Woo et al., 1983; Goodison et al., 1998; Smith, 2007; Pan et al., 2016), sublimation during blowing snow events (Déry et al., 1998; Xiao et al., 2000; Liston et al., 2002; Déry and Yau, 2002; Yang et al., 2010), and water vapour flux from frozen soil to snowpack (Santeford, 1978; Woo, 1982). In addition, there are large errors in our ability to accurately map SWE at the watershed scale (Woo and Marsh, 1978; Goodison and Walker, 1993; Koenig and Forster, 2004; Pulliainen, 2006; Takala et al., 2011). This limitation is due to the combination of a highly spatially heterogeneous snow cover and using simple approaches to measuring snow on the ground and its distribution across the landscape. For example, typical methods for measuring snow on the ground still follow the basic procedures of Church (1933) who pioneered the use of snow surveys using a simple rod to measure snow depth and a core tube to estimate snow density. These depth and density observations were, and still are, carried out along snow courses selected to represent average snow conditions. The well known limitations to this approach, suggest that there are large errors in past records of the snow environment across the Arctic, and that this limits our ability to develop and test appropriate models of snow cover at the scales needed to better understand many aspects of snow cover and to model snowmelt runoff for example. The result is that the typically short periods of record, and the larger errors in measuring all aspects of Arctic snow, combine to make it
likely that we have insufficient knowledge of the Arctic snow cover as needed to meet societal needs. This thesis will address aspects of this deficiency.

However, significant advances have been made in recent years to improve the measurement of one of the deficiencies noted above – namely the measurement of snow depth across watersheds in order to better quantify the spatial variability in snow depth. These advances have focussed on mapping of the snow surface elevation using a number of ground based and airborne methods. These advances include: using airborne Light detection and ranging (LiDAR) (Kirchner et al., 2014), ground-based LiDAR (Filhol and Sturm, 2015) and photogrammetry using manned aircraft and Structure-from-Motion (SfM; Westoby et al., 2012) imaging analysis techniques (Nolan et al., 2015). Recently, Unmanned Aerial Systems (UAS) have been tested as a new source of imagery for using SfM to map snow depth at high resolution and frequent intervals, and at a relatively low cost (Vander Jagt et al., 2015; Bühler et al., 2015; Nolan et al., 2015; Harder et al., 2016). However, few if any studies have tested and used these methods in the unique environment of the Arctic, or have used them to consider errors in the past records of snow on the ground.

Given these significant deficiencies, the overall objectives of this chapter is first to document changes in snow cover at a tundra site in the Canadian Arctic and second to build on recent advances in mapping snow depth to demonstrate a method to improve our ability to map snow cover, to consider errors in the past record of snow on the ground, and to make recommendations for future snow observations. Specifically, we will:

*Objective 1*: Characterize changes in snowfall and snow on the ground over the 60-year period of record at Inuvik Airport (Inuvik-A).
Objective 2: Combine traditional snow surveys, UAS derived maps of snow depth, and various methods to map snow density, in order to better understand the magnitude of snow drifts and to document errors in quantifying end of winter SWE at the Trail Valley Creek research watershed located 50 km north of Inuvik-A.

Objective 3: Make recommendations concerning future snow survey methods for use in the Arctic.

2.2 Study Sites

Field observations were carried out in the Trail Valley Creek (TVC) watershed at 68.75°N, 133.50°W located approximately 50 km north of Inuvik (Mike Zubko) Airport in the Northwest Territories (NT). TVC has been the focus of continuous streamflow observations by the Water Survey of Canada (WSC) since 1977 and snow hydrology research since 1991 (Marsh et al., 2008). The gauged portion of the watershed is approximately 57 km² in area (Marsh et al., 2008), and is characterized by rolling hills with some deeply carved river valleys. Elevations range from approximately 45 to 190 meters above sea level (Marsh et al., 2010). TVC is at the northern edge of the tundra-taiga interface (Callaghan et al., 2002; Bliss and Matveyeva, 1992) and is underlain by continuous permafrost between 150 and 350 meters thick (Heginbottom and Radburn, 1992), with an active layer ranging from 0.3 to 1.0 m (Marsh et al., 1995; Endrizzi et al., 2011).

Snow cover depth, density, and SWE at TVC, as across most of the Arctic, varies greatly over small spatial scales due to extensive blowing snow, transporting snow from
source areas (hilltops and lakes) to drift areas (lee slopes, shrub patches, leading edges of forest patches, stream channels and lake edges). Following Tabler (1975), Marsh and Pomeroy (1996) mapped slope drifts as those that form on slopes greater than 9 degrees, as well as those on lake edges and stream valleys. For TVC, these slope drifts were estimated to cover 8% of the watershed, similar to many other lowland, Arctic watersheds (Winstral and Marks, 2014; Rees et al., 2014). Marsh and Pomeroy (1996) classified the remainder of TVC by vegetation cover with tundra (69.8%), shrubs (21.5%), and forest patches (0.5%). The tundra is comprised of grasses, lichens (Lecidea) and mosses (Sphagnum). While shrub patches consist of 0.5 to 3 m high alder (Alnus viridis), willow (Salix glauca) and birch (Betula glandulosa). These shrub patches primarily occupy stream edges, lake edges, hill slopes and some upland area. Forest is comprised of white spruce (Picea glauca) up to 10 m in height, and occur in a range of environments from valley bottoms to upland plateaus in the lower reaches of TVC.

The climate of the region is marked by short summers and long cold winters, with about 8 to 9 months (September/October to May/June) of snow on the ground. The mean annual air temperature, rainfall and snowfall for Inuvik are -8.2°C, 114 mm and 159 cm (snowfall is uncorrected for wind undercatch) respectively, for a total of 241 mm of precipitation (Canadian Climate Normals 1981-2010, 2016). End of winter spatial variability in snow depth, density and SWE at TVC is very high, with typical upland snow cover averaging 40 cm in depth, while drifts are often up to 300 cm in depth (Marsh and Pomeroy, 1996), and in extreme cases up to 600 cm.

Field work for this study occurred during April and May of 2015 and 2016, and focused on Siksik Creek (Figure 1), a representative sub-basin of TVC that is
approximately 1 km² and located in the lower section of TVC. Although, this sub-
catchment has lower relief than TVC, with elevations ranging from approximately 45 to
100 m above sea level, the general characteristics of Siksi are similar to TVC with
rolling hills, short, steep slopes and shrub patches (but no forest patches) that are prone to
the development of snow drifts. The large number of end of winter snow drifts that are
typical of this region are illustrated by aerial photos (Figure 2) taken after the start of melt
when only drifts remain. Terrain dominated drifts in Siksi Creek were classified using
the same methods as for TVC (Marsh and Pomeroy, 1996) and only cover 5% of the
catchment. Following Marsh and Pomeroy (1996), the remainder of Siksi Creek was
classified by vegetation cover with tundra (79%), shrubs (vegetation dominated drifts;
16%), and forest patches (0%), similar to that of TVC. Snow measurements carried out in
Siksi will be complemented by snow depth and density data from key sites across TVC.
Figure 1. Land classification of Siksik Creek watershed, showing slopes with angles > 9° (black polygons), where terrain dominated drifts typically form, and shrub patches (green) defined as areas with vegetation heights > 0.5 m, and tundra (white) with vegetation heights ≤ 0.5 m. Slope and vegetation heights were derived from airborne Lidar data collected August 2008 (Hopkinson et al., 2009). Trail Valley and Siksik Creek (blue lines) flow to the east and southeast respectively. Inset map shows TVC located to the east of the Mackenzie Delta, south of the Beaufort Sea and north of Inuvik.
Figure 2. Aerial image of Siksik Creek, May 13th 2016, 18 days after the start of snowmelt. Remaining late lying snow patches (shown in white) illustrate the locations of end of winter snow drifts which had much deeper than average snow depth. Areas with slopes greater than 9 degrees, as in Figure 1, are outlined in black, while linear blue lines are stream channels. Snow free, wetted surfaces are darker in colour and dry surfaces in lighter colours. Trail Valley Creek flows to the east and Siksik Creek to the southeast.
2.3 Methodology

2.3.1 Snow observations

Although climate models suggest that precipitation will increase in most environments, various studies have clearly shown that changes in snowfall across the Arctic have been extremely variable. Understanding past changes in the snow environment in the Arctic is greatly complicated by typically short periods of weather station records, large errors in measurement of snowfall and snow on the ground, and few long-term records of natural snow cover across watersheds. The following section will outline a long term data set from the western Canadian Arctic that will allow for an improved understanding of the past changes in snow in this region of the Arctic.

Precipitation and related meteorological data are available from Inuvik Airport (Inuvik-A; YEV, 2202570) (Figure 1) for the period 1958 to present (climate.weather.gc.ca). This is one of the longest, continuous, period of precipitation and snow on the ground data in the western Canadian Arctic, and provides a unique opportunity to consider changes in winter snowfall in this region of the Arctic. There are various snow data sets from Inuvik-A that are available for analysis, each with various advantages and disadvantages. These include: snowfall estimated from ruler measurements of snow on the ground over a representative area (Metcalfe et al., 1994), while more recently snow on the ground has been estimated from SR50 ultrasonic ranging sensors, snowfall has been collected using a variety of manual or weighing precipitation gauges, and snow on the ground, under a natural open forest site, has been estimated from 20 point snow surveys. There are limitations to each of these data sets, including the following. Wang et al. (2017) used the ruler measurements of snowfall at Canadian
weather stations, but had to estimate snow density to estimate snowfall as a depth of water and collection of these data ended in 2003 for Inuvik-A. The accuracy and consistency of snowfall from precipitation gauges also suffers from a number of issues, including changes in gauge type over the period of record. From manual gauges with Nipher shields in the early 1960s, to automatic Geonor weighing gauge with Alter shields beginning in the early 2000s. In addition, snowfall is typically under caught by all gauges due to wind effects around the orifice of the gauge. Finally, the 20 point snow surveys produced excellent monthly estimates of snow on the ground below a forest canopy, but do not provide snowfall information. In addition, these observations were discontinued a number of years ago. After careful consideration of each of these data sets, it was decided to use the total precipitation data from Inuvik-A during the period of October to May as this record was the most complete, compared to the aforementioned snowfall data set that is missing recent data (post 2003). It should be noted here that there is no consistent data set of wind undercatch corrected winter snowfall for Canada. Wang et al. (2017) provide an adjusted data set of snowfall, for trace values, using ruler measurements of snow on the ground and estimates of snow density. However, this adjusted data set for Inuvik snowfall is limited to the period of 1958-2003. In contrast, the climate.weather.gc.ca data set has a relatively full period of record for total precipitation. In the Arctic, when all winter precipitation falls as snow, it is felt that it is most consistent to use the measured precipitation for Inuvik-A and separate snowfall based on air temperature. Therefore, annual snowfall at Inuvik-A was estimated by totalizing daily precipitation for days with mean air temperatures below 0ºC from October 1st to May 30th. Separating rain from snow when air temperatures are near 0ºC is difficult due to many complicating factors.
(Marks et al., 2013), one of which is large time steps of observational data (daily). Marks et al., 2013 showed that methods, to separate rain and snow, that use either a constant isotherm of 0°C for air temperature, dew-point temperature or wet-bulb temperature are reasonably effective when applied at daily time steps. In this study, we will use 0°C daily mean air temperatures to separate rain from snow for the period of interest. In addition to snowfall, we will use ruler and SR50 measurements of snow on the ground from Inuvik-A to compare to the snowfall for trends. However, we will not attempt to estimate snow density in order to estimate SWE.

In addition to the Inuvik-A snowfall measurements, snowfall and related meteorological observations from the TVC Main “Meteorological” station (TMM) (Figure 1), and TVC Forest Site (TFS) in the TVC catchment will be used. TMM has two weather stations, the Wilfrid Laurier weather station (TMM-W) that was installed in 1991 and a Meteorological Service of Canada (MSC) weather station (TMM-M; 220N005) installed in 1998 during the Mackenzie GEWEX Study (MAGS) (Stewart et al., 1998). TFS was installed in 2009. TMM-W and TMM-M are located approximately 20 meters apart in the central portion of the Siksik Creek watershed (Figure 1) and TFS is located in a sparse forest patch approximately 2 km south of the TMM stations (in the TVC watershed, but outside of Siksik). TMM-M, TMM-W and TFS currently measure precipitation (single alter shielded T-200B Geonor weighing gauge; Figure 4), air temperature and relative humidity (Vaisala model hygrothermometer), wind speed and direction (RM Young anemometer) and snow depth (SR50 ultrasonic ranging sensor). Snowfall has been measured at TMM-W since 1991, but with varying instruments. From 1991 to 1998 a Nipher-shielded MSC manual gauge was used, following that a Nipher...
shielded Fisher-Porter weighing gauge was in operation from 1998 to 2005. This gauge was destroyed by a bear in the summer of 2005 and an Alter shielded Geonor was installed in September 2006. However, winter winds speeds at this site are missing from 2006 until 2008 due to rime-ice covering the anemometer and in September 2008 a RM Young anemometer was installed at a higher height on the tower. Therefore, corrected Geonor weighing precipitation was only calculated from 2008 to 2016.

Duplication of the Geonor weighing gauges at TMM-W and TMM-M within a small area (~20 meters) provides both protection against gauge malfunctions and allows consideration of the range of errors between the same model precipitation gauges under different wind conditions of different wind undercatch for example. The TFS Geonor provides the opportunity for snowfall observations from a lower wind speed environment as it is shielded by the open forest (Figure 4), and potentially will provide a higher quality data set. Unlike the other two stations, TFS is also equipped with a 7.5 m$^2$ snow scale (CRREL/NRCS SWE snow scale sensor 002.1; Figure 3) mounted on 4 load cells to measure SWE accumulation over an area of 0.84 m$^2$. Pomeroy et al. (1997) suggested that with little blowing snow in the forest, and minimal canopy interception due to the widely separated trees, snow on the ground at this site should approximate total winter snowfall. If correct, snowfall, snow scale, and snow survey at this site should be very similar.
Figure 3. Snow scale located at the forest site equipped with 4 load cells under the centre panel. Each of the 9 panels are 0.91 by 0.91 m.

Figure 4. Single Alter shielded Geonor weighing gauges at a) TMM-W on the right and TMM-M on the left and b) TFS
2.3.2 Snowfall correction

During international studies of snow precipitation undercatch errors, the double fenced intercomparison reference (DFIR) gauge is considered to measure actual snowfall (Goodison et al., 1998). All other models of snow precipitation gauges (irrespective of gauge design, manual or automatic recording, or wind shield design) are known to significantly underestimate precipitation (Goodison et al., 1998; Smith, 2007) due to systematic error related to wind flow over and around the gauge’s orifice. To correct undercatch errors for single alter shielded Geonors, solid precipitation is adjusted for wind speed using the following standard approach (Smith, 2007; Pan et al., 2016):

\[ P_{\text{corr}} = \frac{P_{\text{obs}}}{CE} \]

(1)

where \( P_{\text{corr}} \) (mm) is the solid precipitation corrected for wind, \( P_{\text{obs}} \) (mm) is the observed solid precipitation and \( CE \) is the catch efficiency, with \( CE \) calculated as:

\[ CE = 1.18 \cdot \exp(-0.18 \cdot W_s) \]

(2)

Where \( W_s \) (m/s) is the wind speed at Geonor gauge height (approximately 2 m). Due to stronger wind speeds in the Arctic and because of a lack of tall vegetation at most sites (TMM for example), we used the following suggestions of Pan et al. (2016): a minimum threshold of 1.2 m/s and a maximum threshold of 9 m/s for the wind speed, resulting in catch efficiencies of 1 and 0.23 for winds below 1.2 m/s and above 9 m/s respectively.

Daily images, from a time lapse camera installed at TMM, were used to determine the onset of snow accumulation for winters of 2014/15 and 2015/16.

2.3.3 Spatially distributed watershed snow cover

The weather station snowfall and snow on the ground data outlined above are extremely useful in understanding long term trends in snowfall. However, actual snow on
the ground is also significantly impacted by other processes, including, blowing snow redistribution and sublimation, vapour transfer from the soil to the snowpack, and forest and shrub canopy interception and sublimation. To consider these effects, we will use spatially distributed snow survey data from Trail Valley Creek watershed.

2.3.3.1 Trail Valley Creek

From 1991 to 2017, end of winter snow surveys were carried out in TVC catchment at the same locations reported by Marsh and Pomeroy (1996). Surveys were carried out in terrain types that were representative of the TVC watershed, including: tundra (low vegetation less than 0.5 m; 69.8%), shrub (tall shrub vegetation greater than 0.5 m, but less than 3 m; 21.5%), forest (tree vegetation greater than 3 m; 0.5%) and drift (slopes greater than 9°; 8.2%). A total of 6808 snow depths and 854 snow density measurements were obtained over this period throughout 3 tundra sites, 2 shrub sites, 1 forest site and 3 drift sites. These data will be used to consider changes in snow on the ground over this period of record. However, it should be noted that caution must be used with these data due to changes in observers and observation methods. For example, 28 snow surveyors have measured end of winter snow accumulation since 1991, likely resulting in some discrepancies in techniques and sampling method between surveyors (Berezovskaya and Kane, 2007), and resulting errors. In addition, five types of snow core tubes for measuring snow density were used over this period. Large diameter tubes for shallow snow consisted of the Eastern Snow Conference model 30 (ESC-30; 30 cm² cross-sectional area), Meteorological Survey of Canada Type 1 (MSC; 39 cm² cross-sectional area) and SnowHydro (30 cm² cross-sectional area). Farnes et al. (1982)
describes the differences between the ESC-30 and MSC tubes, with MSC snow tube having a 7% average error and the ESC-30 an average error of 0.3%. In addition, small diameter tubes used for deeper snow consisted of the Mount Rose (11 cm² cross-sectional area) and Standard Federal (11 cm² cross-sectional area). Again, these are described by Farnes et al. (1982), with the Standard Federal snow tube having a 10% average error and the Mount Rose an average error of 4.1%. The various differences in tube design (metal with spaces, or acrylic for seeing the core) and differences in cutter teeth design (shape and number of teeth) between these tubes result in well known variances in density estimates (Farnes et al., 1982; Goodison et al., 1987; Dixon and Boon, 2011). For snow depth measurements, two types of probes were used, avalanche snow depth probes (2 m and 5 m maximum depth) and more recently the MagnaProbes (1.2 m maximum depth; Sturm and Holmgren, 2002; Marshall et al., 2006). The 1991-2017 snow survey period will be used to analyze for long term trends in snow depth, density and SWE, and to put the 2014/15 and 2015/16 winters into this long-term context.

Following Steppuhn (1976), the following equation is used to calculate SWE for each snow survey:

$$\overline{SWE} = \frac{\bar{\rho}_s}{\rho_w} \cdot \overline{d_s} \cdot 10$$  (3)

Where $\overline{SWE}$ (mm) is the mean SWE for a transect, $\bar{\rho}_s$ (g/cm³) is the mean density along that transect, $\rho_w$ is the density of water and is equal to 1 g/cm³, $\overline{d_s}$ (cm) is the mean snow depth of that same transect and the constant 10 is the unit conversion from cm to mm.
2.3.3.2 Siksik Creek

In addition to the snow surveys noted above and carried out across TVC, additional snow surveys were conducted within Siksik during April of 2015 and 2016. The Siksik snow surveys were conducted along transects selected to be representative of the catchment’s land classification (vegetation and topography) and well distributed throughout the catchment (Figure 5) in order to document the spatial variability in snow depth, density and SWE. Snow depths were sampled at every 2 meters along 15 transects (April 2015) and 24 transects (April 2016), of 50 to 180 meters in length, which covered tundra, shrubs and drifts land cover types. Where snow depths were greater than 1.2 m, total depth was estimated from an avalanche probe that can measure depths up to 5 meters. In total, 516 and 832 snow depth measurements were obtained in 2015 and 2016 respectively using a combination of MagnaProbe and avalanche probe measurements. More snow depths were measured in 2016 than 2015 due to the need for a well distributed data set.

Snow densities during the two study winters were obtained using a combination of ESC-30, SnowHydro and Standard Federal snow core tubes depending on the year and snow depth. Density measurements were typically made every 10 snow depths. A total of 46 and 84 end of winter snow density measurements were obtained for April 2015 and 2016 respectively along transects noted above. A mean SWE (mm) was calculated for each transect by using equation 3. Total basin SWE for Siksik Creek were calculated using weighted averages of SWE for each landcover type similarly done for the past snow surveys at TVC.
Figure 5. DEM of Siksik Creek catchment. Transects where manual snow depth measurements were obtained are shown in pink (2015) and in light blue (2016). GPS ground control points are shown in red (2015) and blue (2016). The catchment is shown as the red polygon and the creeks as black lines, with TVC flowing towards the east.
2.3.4 High resolution mapping of end of winter snow depth

We obtained high resolution images of the Siksik domain using a light-weight, fixed-wing eBee (SenseFly®) UAS equipped with a Cannon PowerShot ELPH 110 HS RGB camera with 16.1 megapixels. All images were obtained near the end of winter and immediately before the melt period. The UAS was flown on days with low winds (less than 5 m/s) and in a north-south direction at a height of 186 m above sea level, and with an image overlap of 60 to 80%. Prior to the flight of the UAS, 12 ground control points (GCPs; Figure 5) were distributed throughout the catchment in both study winters. Some GCPs were placed on the snow surface and others at the base of snow pits and directly on the snow free tundra. These GCPs were surveyed using Real Time Kinematic (RTK) GPS (Leica SR530 dual frequency) with accuracies of under 0.05 m horizontally and vertically. UAS imagery was collected on April 28th 2015 and April 23rd 2016, with 911 and 656 images respectively. Postflight Terra 3D, was utilized to post-process and stitch together each yearly set of images by matching 2D keypoints found on multiple images. The post-processing software then used the Structure-from-Motion (SfM) imaging technique to estimate the 3D structure from the assembled 2D imagery (Westoby et al., 2012). Postflight Terra 3D then analyzed the images and generated a digital surface model (DSM), for each date, at a horizontal and vertical resolution of 0.03 m covering an area of approximately 2 km² that covered all of Siksik Creek watershed. These DSMs were aggregated to a 1 m horizontal resolution and then subtracted from a snow-free LiDAR digital elevation model (DEM) (Hopkinson et al., 2009) with a horizontal and vertical resolution of 1 and 0.05 m in order to map snow depth across the entire Siksik
Creek watershed. The end result is snow depth maps with a horizontal resolution of 1 m and a vertical resolution of 0.05 m.

2.3.5 Mapping end of winter snow density

In order to use the maps of snow depth from the UAS to map SWE at high resolution, there is also a need for high resolution maps of snow density. Few, if any, studies have carried this out, and none to our knowledge in Arctic. The Airborne Snow Observatory (ASO) operationally maps snow depth using airborne LiDAR in the Sierra Nevada mountains (Kirchner et al., 2014), and since snow density in this environment is dominated by melt freeze and compaction processes, the ASO estimates snow density from a simple snow density model (Painter et al., 2014). For the Arctic, temporal and spatial variability in snow density are significantly more complicated. One process that is very important in the Arctic is the development of depth hoar of very low density. Snow density models have done a poor job at predicting the development of depth hoar within the snowpack because they have difficulties in reproducing the variability of the depth hoar properties within the same snowpack (Barman and Jain, 2016). Therefore, process based modelling of density is currently not an option. Instead we will estimate snow density across the watershed by applying simple relationships using end of winter snow depth and density data (1390 measurements) from the 1991-2016. Specifically, we will use: (1) mean density for the 1991 to 2016 period for each grid point across the entire Siksik basin (Rees et al., 2014), (2) an inverse distance weighting (IDW) interpolation of the observed 2015 and 2016 densities and (3) a depth/density relationship (Sturm et al., 2010) determined from the 1991 to 2016 density data. These methods were chosen based
on their different level of simplicity or ease of estimating density for the catchment, method 1 being the simplest to method 3 being the most complex. They were used to estimate spatial variation in density, and to allow mapping of SWE across the Siksik watershed.

2.3.6 Mapping end of winter SWE

SWE across the Siksik watershed for 2015 and 2016 was mapped using UAS estimated depths and the three density estimation methods noted above. Spatial distribution in end of winter SWE was obtained via maps created, multiple distributed mean SWE UAS estimates for the catchment were calculated for each density method listed above and compared to their respective manual snow survey distributed mean SWE. Finally, estimates of the total water available for melt from Siksik Creek are obtained.

2.4 Results

2.4.1 Past changes in the snow environment

2.4.1.1 Inuvik winter snowfall: 1958 to 2017

The snow record at Inuvik-A for 1958-2017 (Figure 6) shows a general decline in all measurements of snowfall and snow on the ground. For example, during the first 30 years (1958-1987), mean, maximum, minimum and standard deviation of snowfall was 131, 192, 80 and 32 mm respectively, while during the last 30 years (1988-2017), the values are 111, 179, 63 and 31 mm respectively. Figure 6 shows that in Inuvik, snowfall during the 2015 and 2016 study winters were extremely low, with the 2010 and 2014 winters holding the lowest snowfalls on record. Furthermore, Table 1 illustrates the
difference in mean snowfall, snow depth, snowfall as a percentage of yearly precipitation and air temperature between two 30-year periods at Inuvik-A. As shown in Table 1, there is a decrease in the four parameters except for air temperature between the two 30 year periods. With the lack of snow density data for the period of record, we can’t directly compare changes in snowfall to changes in snow depth. Using Mann-Kendall trend test, we see significant decline in snowfall (p = 0.0002) and maximum snow on the ground (p = 0.003) at Inuvik-A. These trends are subject to changes in instrumentation and measurement methods, for example, the station transitioned from manual measurements to automated measurements in 1995 and therefore we see a shift in the snowfall data.

\[\text{Figure 6. Inuvik yearly snowfall (grey columns), yearly maximum snow depth (solid black line) and percent of yearly precipitation as snow (solid red line) from 1958 to 2017. The dashed grey line and dashed horizontal black line are the trendlines for snowfall and maximum snow depth respectively. Snowfall and maximum snow depth data were acquired from the unadjusted daily data set at http://climate.weather.gc.ca.}\]

\[\begin{array}{|c|c|c|c|}
\hline
\text{Parameter} & \text{1958-1987} & \text{1988-2017} & \text{Difference} \\
\hline
\text{Mean snowfall (mm of SWE)} & 131 & 111 & -20 \\
\text{Mean snow depth (cm)} & 71 & 61 & -10 \\
\text{Mean snow percentage (\%)} & 52 & 45 & -7 \\
\text{Mean air temperature (°C)} & -9.6 & -7.2 & 2.4 \\
\hline
\end{array}\]

\[\text{Table 1. Inuvik-A 30-year means.}\]
2.4.1.2 Trail Valley snowfall: 2008 to 2016

As shown in Figure 7 for the period of snowfall observations at TMM, there are variations from year to year, with a maximum of 176 mm in 2011/12, and a minimum of 88 mm in 2015/16. Figure 7 also shows that the Inuvik-A uncorrected snowfall is typically larger than the TMM-W uncorrected snowfall but considerably smaller than the corrected TMM-W snowfall. The mean winter uncorrected snowfall at TVC was 71 mm and the mean corrected snowfall was 137 mm, lower than the uncorrected snowfall at Inuvik-A for this period of 94 mm. Although Inuvik-A wind speed is less than at TMM-W, it is unclear if the difference in snowfall is due to using uncorrected snowfall at Inuvik, or due to an actual difference in snowfall between the two stations. It should be noted that the long term MSC snowfall records show that Tuktoyaktuk, 80 km north of TVC on the Beaufort Sea coast has average winter snowfall of 103 mm compared to 159 mm at Inuvik (Canadian Climate Normals 1981-2010, 2016).
Pan et al. (2016) ran bias corrections for yearly precipitation (including rainfall, mixed precipitation and snowfall) at TVC for the years 2008 to 2014. They found that the annual mean corrected precipitation at TMM was 251 mm, while we attained a mean corrected snowfall of 137 mm of SWE at TMM for the years 2008 to 2016. This suggests that snowfall accounts for 55% of the yearly precipitation at TVC, slightly higher than the average of 49% for the Inuvik region.

Figure 8 shows a selection of the metrics collected in the two study winters (2014/15 and 2015/16) at TMM and TFS. Winter of 2015/16 saw colder temperatures, lighter winds and less uncorrected and corrected precipitation than the previous winter. Furthermore, the winds at TFS are lighter than those at TMM-W, suggesting that the gauge undercatch may be lower, that the forest cover hinders blowing snow events within
the forest patch and hence the forest snow cover may be closer in magnitude to the winter snowfall. Differences between snowfall measurements at TMM-W and TMM-M are 25 mm for 2014/15 and 22 mm for 2015/16, these differences can be explained by the placement of the gauges. TMM-M Geonor gauge is placed closer (within 2 m) to tall, wind stream obstructing, instruments whereas the TMM-W gauge is in an open area (10 m away from any obstruction) (Figure 4a). However, previous to the winter of 2015/16, the TMM-W gauge became tilted (approximately 10°) which increased the amplitude of the signal noise in the raw data, making it much harder to pick out snowfall events to wind pressure error. Therefore, we see less precipitation in 2015/16 at TMM-W than TMM-M. Much larger differences are present between the TFS gauge and the other two gauges due to lower wind speeds in the forest cover resulting in a lower wind-induced undercatch correction.
Lesack et al. (2014, Figure 4) showed that maximum annual snow on the ground measured by a daily average of multiple point measurements at Inuvik for the 56-year period of record (1957-2012) varied considerably. They indicated that snow depths at Inuvik based on the end of April have declined from an average of 59 to 25 cm between the periods of 1957-1985 and 1986-2012. Figure 6 shows another version of these data.

2.4.1.3 Changes in end of winter snow cover

Figure 8. Daily time series of meteorological data for the winters of 2014/2015 (left; 1-Oct-2014 to 6-May-2015) and 2015/2016 (right; 25-Sept-2015 to 25-April-2015). 2-meter air temperature at TMM-W (a, b), 10-meter wind speed at TMM-W and TFS (c, d) and total snowfall at TMM-W, TMM-M and TFS (e, f, g, h) in mm of SWE. For total snowfall, a comparison of the three Geonor gauges is shown for uncorrected snowfall (e, f) and for corrected snowfall for wind-induced undercatch (g, h). The red dashed line in g and h is the snow scale located at TFS, meters away from the Geonor gauge.
again illustrating that maximum snow depth has been slowly declining over the period of record. Note that these are snow depths, not SWE, and can not be directly compared to snowfall (Figure 6), but the trend is consistent between the two data sets. The average peak snow depth for the entire period of record is 66 cm, with a mean, maximum, minimum and standard deviation of 71, 99, 43 and 14 cm respectively during the first 30 years and 61, 116, 40 and 16 cm respectively for the last 30 years. A decrease in the 30 year mean is observed in the mean peak and minimum peak, however an increase in the 30 year mean is observed in the maximum peak and standard deviation peak between the two 30-year periods. This decrease in mean peak snow depth, and increase in variability is consistent with predictions from climate models. Changes in snow depth is extremely important as depth plays a significant role in ecology and controlling the thermal insulation provided by the snow to soil (Barrere et al., 2017), and hence influences soil temperature and permafrost.

Figure 9a, b and c illustrate that over the past 26 years (1991 to 2017), TVC end of winter average snow depths, density and SWE for each of the TVC landcover types have shown slight changes but also large variability. Snow depth, density and SWE have mean coefficients of variation (CV) of 0.30, 0.25 and 0.23 respectively. Average snow depth for the terrain types (tundra, shrub, forest and drift) was 37, 69, 73 and 185 cm respectively over the entire period of record. For 1991 to 2004, depths averaged 43, 80, 77 and 197 cm respectively, and for 2005 to 2017 depths averaged 39, 63, 69 and 183 cm. Little change in the drift class over the period of record is likely related to the fact that drift depth is likely more related to winter wind blowing snow events, and it is unknown whether there has been a change in the magnitude and frequency of such events.
End of winter SWE for tundra, shrubs, forest and drifts respectively ranged from: 60 to 170 mm, 80 to 340 mm, 80 to 290 mm and 33 to 1270 mm, with basin average end of winter SWE ranging from 108 to 253 mm. There is greater inter-annual snow depth and SWE variability in the drift landcover. Figure 9d also illustrates that over the past 26 years (1991 to 2017), the TVC end of winter snow cover has averaged 167 mm of SWE, with a maximum of 253 mm in 1991/92, a minimum of 108 mm in 2006/07 and a standard deviation of 38 mm. However, there has been no significant trend in end of winter basin SWE.

Decreasing snowfall and end of winter SWE is not only sensitive to decreasing precipitation, but also to the length of the snowfall period. As an example, snow survey dates at TVC have been occurring earlier due to the increasing uncertainty when melt will begin in the spring. Hence, the surveys could be missing snowfall that occurs later. A more rigorous analysis uses the Inuvik-A data over the full 60-year period of record. Inuvik winters are indicated by the period between the first and last three consecutive daily mean air temperatures below the 0°C isotherm. Changes in the end of winter using this definition is similar to that reported by Shi et al. (2015) who showed that spring melt is occurring approximately 9 days earlier using a slightly different definition. The changes in fall and spring temperature results in a reduction of total winter length by approximately 14 days over the 60-year period seen in the trendline in Figure 10. With the use of the Mann-Kendall trend test the total winter length (Figure 10) shows a significant (p = 0.04) trend.
Figure 9. End of winter snow surveys results at TVC from 1991 to 2017. Snow depth, density and SWE for tundra (blue), shrub (red), forest (green) and drift (gray) are shown in a, b and c respectively. Basin mean SWE (black) and aerial weighted SWE for each landcover type shown in d. The basin mean SWE was calculated for winters when certain landcover types were not sampled by using the period of record mean SWE for that landcover type.
2.4.1.4 Snowfall compared to end of winter SWE

SWE from snow surveys and from snowfall are compared in Figure 7 for the period 2008-2016 at TVC. Over this period, end of winter basin average SWE has been, on average, 24% greater than corrected snowfall and only four out of eight of the winters snowfall has been within 1 standard deviation of the snow surveys (Figure 7). This difference is not due to the effect of blowing snow sublimation, as sublimation would reduce basin SWE compared to snowfall. Note that Pomeroy et al. (1997) have estimated winter sublimation at TVC to be 20% of total snowfall. Larger SWE than corrected snowfall could be the result of two possibilities: i) the TVC basin may be a sink for blowing snow or ii) that true snowfall is actually larger than our estimate from snow gauges at TVC. Figure 7 shows that there doesn’t seem to be a year to year relationship between end of winter SWE on the ground and corrected snowfall. Nevertheless, in this 8-year period there is a slight decrease in both snowfall and end of winter SWE.

Figure 10. Length of winters in Inuvik, NWT, defined simply as the period between the first and last three consecutive daily mean air temperatures below the 0°C isotherm. The dashed line shows the 60-year trend.
2.4.2 Errors involved in estimating watershed snow cover SWE

As noted earlier, there are considerable concerns over the errors associated with measuring snowfall and snow on the ground in the Arctic. Errors in measuring distributed SWE is primarily due to the concerns of measuring SWE in snow drifts. For example, Pomeroy et al. (1997) suggested that drifts contain a similar quantity of SWE as that located on tundra, with drifts covering only 8% of the landscape and tundra 70%. While, for the 1991-2017 period, snow surveys at TVC suggest a similar distribution with the percent of total basin SWE contained in each terrain type (tundra, shrub, forest and drift) as 39.6 %, 20.6 %, 0.5 % and 39.2 % respectively. Here we see that drift and tundra landcover types are similar in terms of the percentage of SWE stored in the basin, in spite of the large differences in area that they represent in the basin. However, as there are likely considerable uncertainties in measuring both drift area and SWE in drifts, there are large errors in documenting the average snow contained in drifts and the area that they occupy. To consider this issue, new methods using a UAS were implemented to map snow depth, snow density and SWE over the Siksik watershed, and then to compare these to traditional snow surveys.
2.4.2.1 Snow depths

The snow depth maps in Figure 11 display high spatial variability in the snow depth across the Siksik catchment, with snow depths varying from 50 cm to above 200 cm over distances as short as 10 m. Deeper snowpacks are seen on the western side of the catchment with greatest snow depths located in terrain and vegetation dominated drifts. In contrast, shallower snowpacks are observed on plateaus where wind erosion is strongest throughout the winter season. In 2015, the in-situ GPS measured observations had a mean

![Figure 11. End of winter (April 28th 2015 (left) and April 23rd 2016 (right)) snow depth as measured by UAS for Siksik catchment at a 1 m resolution. “No Data” represents areas where Postflight Terra 3D was unable to find sufficient 2D keypoint matches. Manual observations (brown lines) and ground control points (GCP; red points) are illustrated. UAS estimated snow depth profiles (gray lines on maps) illustrated below maps.](image-url)
snow depth of 67.2 cm and a standard deviation of 38.0 cm, while the UAS derived mean snow depth was 79.5 cm and a standard deviation of 47.4 cm. As for 2016, the in situ mean snow depth was 60.5 cm and a standard deviation of 24.4 cm, while the UAS derived mean snow depth was 57.6 cm and a standard deviation of 26.5 cm.

The relationship between the snow depths as estimated from the UAS and probing for the two study winters are in Figure 12 and in Table 2. The 2015 flight compared 516 points (15 transects) with an absolute mean bias of 35.6 cm, a root mean square error (RMSE) of 49.0 cm and a $R^2$ value of 0.67, while the 2016 flight only compared 832 points (24 transects) with an absolute mean bias of 19.2 cm, a RMSE of 27.6 cm and a $R^2$ value of 0.40. The spread of snow depths along the 1:1 line, in Figure 12, is much greater in 2015, resulting in a $R^2$ value much larger than that of 2016. The estimated values have also been compared to the observed values in terms of their probability distribution functions (PDFs) of snow depth in Figure 13 and their statistics are given in Table 2. A two sample Kolmogorov-Smirnov test was computed for both study winters PDFs, resulting in an absolute maximum difference of 0.207 and critical difference of 63.6 for 2015 as well as 0.097 and 49.8 respectively for 2016. Both absolute maximum differences

![Figure 12](image-url)  
*Figure 12. Scatter plots of observed snow depth vs. UAS estimated snow depth, scatter plot of April 2015 (left) with a RMSE of 49.0 cm and a $R^2$ value of 0.67 and scatter plot of April 2016 (right) with a RMSE of 27.6 cm and a $R^2$ value of 0.40.*
values are smaller than their respective critical difference and therefore the UAS and observed snow depth frequencies are distributed similarly. This confirms that both samples are from the same distributed population. Furthermore, Figure 13c and f show that the observed snow depths have larger relative frequency peaks.

Figure 13. Probability distribution functions (PDFs) of a) observed manual snow depths in April 2015, b) UAS estimated snow depths in April 2015, c) comparing observed (red) to estimated (blue) snow depth in April 2015, d) observed manual snow depths in April 2016, e) UAS estimated snow depths in April 2016, f) comparing manual (red) to UAS (blue) snow depth in April 2016.

Table 2. Snow depth statistics for Figures 12 and 13.

<table>
<thead>
<tr>
<th></th>
<th>Snow depth</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Samples</td>
</tr>
<tr>
<td>April 2015</td>
<td>Manual (Magnaprobe)</td>
</tr>
<tr>
<td></td>
<td>UAS (scatter plot)</td>
</tr>
<tr>
<td></td>
<td>UAS (distribution)</td>
</tr>
<tr>
<td>April 2016</td>
<td>Manual (Magnaprobe)</td>
</tr>
<tr>
<td></td>
<td>UAS (scatter plot)</td>
</tr>
<tr>
<td></td>
<td>UAS (distribution)</td>
</tr>
</tbody>
</table>
The end of winter mean snow depth, standard deviation (Stdev) and CV for 2015 in Table 2 show quite different distributions for the UAS estimated depths versus the observed depths. Whereas the 2016 end of winter snow depth distributions are well represented with the UAS estimated depths (Figure 13f), with mean snow depth of 60.5 cm and 57.6 cm for observed and estimated respectively. The CV, which is a ratio of the standard deviation to the mean, is a standardized measure of variability and therefore can evaluate the distribution of snow. CV is greater in 2015 than in 2016 for both UAS estimated snow depths and observed snow depths, therefore it demonstrates that there is a greater variability in snow depths during 2015 (Table 2). The reason for this is partially due to the lack of snowfall as well as the lack of possible blowing snow events during the 2015/2016 winter.

2.4.2.2 Estimated snow densities

Method 1 uses a mean density of 0.246 g/cm$^3$ calculated from all 1390 end of winter density measurements throughout 1991 to 2016 for TVC shown in Figure 14 and Table 3. Method 2 interpolates the mean density observed at each transect over the Sikisk watershed for the two respective study years. Method 3 uses the following linear regression from the end of winter densities snow core data set (Figure 15):

$$\rho_s = 0.0011 * d_s + 0.1686$$

(4)

Where $\rho_s$ is the snow density in g/cm$^3$ and $d_s$ is the snow depth in cm. End of winter snow density distribution maps were created for the IDW interpolation method and the depth/density relationship method listed above in section 2.3.5 and are shown in Figures 16 and 17 respectively and for 2015 and 2016. Since method 1 is relying on a basin mean
density for the Siksik catchment no map was produced. Densities in Figure 16 range from 0.152 g/cm$^3$ to 0.387 g/cm$^3$, whereas Figure 17 shares the same patterns as the snow depth maps from Figure 11, where denser snowpacks are collocated with deeper snowpacks and range from 0.169 g/cm$^3$ to 0.499 g/cm$^3$. These density maps will, in turn, be used to create SWE distribution maps for the Siksik catchment by multiplying the snow depths from Figure 11 to the densities in Figures 16 and 17.

![Figure 14. End of winter snow density-depth relationship for TVC from 1991 to 2016.](image)

![Figure 15. Linear regression of the end of winter snow density/depth relationship for TVC from 1991 to 2016.](image)
Table 3. Measured snow depths, densities and SWE from snow density corers in TVC from 1991 to 2016.

<table>
<thead>
<tr>
<th>Landcover</th>
<th>Samples</th>
<th>Density Mean (g/cm²)</th>
<th>Density Stddev (g/cm²)</th>
<th>Depth Mean (cm)</th>
<th>Depth Stddev (cm)</th>
<th>SWE Mean (mm)</th>
<th>SWE Stddev (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Forest</td>
<td>167</td>
<td>0.210</td>
<td>0.049</td>
<td>66.9</td>
<td>19.9</td>
<td>140</td>
<td>67</td>
</tr>
<tr>
<td>Shrub</td>
<td>379</td>
<td>0.226</td>
<td>0.060</td>
<td>63.3</td>
<td>22.9</td>
<td>143</td>
<td>63</td>
</tr>
<tr>
<td>Tundra</td>
<td>481</td>
<td>0.222</td>
<td>0.050</td>
<td>41.9</td>
<td>13.1</td>
<td>93</td>
<td>42</td>
</tr>
<tr>
<td>Lake</td>
<td>114</td>
<td>0.260</td>
<td>0.064</td>
<td>22.5</td>
<td>10.6</td>
<td>58</td>
<td>22</td>
</tr>
<tr>
<td>Drift</td>
<td>249</td>
<td>0.340</td>
<td>0.096</td>
<td>134.6</td>
<td>77.1</td>
<td>458</td>
<td>135</td>
</tr>
<tr>
<td>All</td>
<td>1390</td>
<td>0.246</td>
<td>0.079</td>
<td>65.7</td>
<td>50.2</td>
<td>162</td>
<td>66</td>
</tr>
</tbody>
</table>

Figure 16. End of winter interpolated densities of Siksik catchment at a 1 m resolution on April 28th 2015 (left) and April 23rd 2016 (right). Applying an Inverse Distance Weighting (IDW) interpolation to densities obtained along the snow survey transects.
2.4.2.3 Estimated SWE

End of winter SWE distribution maps were created by combining the UAS estimated snow depth map (Figure 11) to each of the three estimated density methods listed above in section 2.3.5 and are shown in Figures 18, 19 and 20. The six SWE maps for 2015 and 2016 capture similar patterns over the landscape (i.e. largest SWE located in the drifts and lowest SWE located on the upland tundra). Their distributions are compared to manually observed distributions in Figure 21. Estimated SWE ranged, for both winters, from 0 mm to approximately 800 mm for method 1, 0 mm to approximately 1400 mm for method 2 and 0 mm to approximately 1000 mm for method 3. In terms of a mean distributed SWE across the catchment, we yielded quite different results between 2015
and 2016. Table 4 illustrates these differences between each year and compares each method as well as UAS derived SWE to manual snow survey SWE. In 2015, the distributed mean SWE had standard errors of 18.5%, 12.2% and 28.0% respectively for methods 1, 2 and 3. Whereas, in 2016, the distributed mean SWE standard errors were smaller, 4.7%, 7.0% and 4.9% respectively for methods 1, 2 and 3. In comparison to Siksik Creek, basin means were 129 and 123 mm in 2015 and 2016 respectively (Table 4), the UAS estimated basin SWE were all larger using all three methods. For example, using method 3, basin mean SWE was 229 and 141 mm leading to standard errors of 77% and 14% in 2015 and 2016 respectively.

Table 5 illustrates the differences of distributed mean and percentage of SWE between each landcover type for each method and the manual snow surveys within Siksik.

![Figure 18. End of winter estimated SWE of Siksik catchment at a 1 m resolution on April 28th 2015 (left) and April 23rd 2016 (right). Method 1: multiplying a basin mean density from the 1991-2016 TVC data set to the estimated snow depths produced by the UAS (Figure 11).](image)
Creek. From this table, we see that drifts (terrain dominated) account for 11 to 14% and 9.6 to 12.6% of the total basin SWE in 2015 and 2016 respectively, however they only occupy 5.1% area of the catchment. Shrubs (vegetation dominated drift) represent 22.7 to 23.8% and 19.7 to 20.7% of the total basin SWE in 2015 and 2016 respectively, with only occupying 16.2% area of the catchment.

Figure 19. End of winter estimated SWE of Siksik catchment at a 1 m resolution on April 28th 2015 (left) and April 23rd 2016 (right). Method 2: multiplying the interpolated densities to the estimated snow depths produced by the UAS (Figure 11).
Figure 20. End of winter estimated SWE of Siksik catchment at a 1 m resolution on April 28th 2015 (left) and April 23rd 2016 (right). Method 3: multiplying densities calculated from equation 4 by the estimated snow depths produced by the UAS (Figure 11).

Table 4. Distributed mean SWE for Siksik catchment using UAS and manual measurements.

<table>
<thead>
<tr>
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<th></th>
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<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>April 2015</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>mean SWE (mm)</td>
<td>195.9</td>
<td>202.1</td>
<td>228.9</td>
<td>165.3</td>
<td>180.1</td>
<td>178.8</td>
<td>129.0</td>
</tr>
<tr>
<td>std dev (mm)</td>
<td>117.2</td>
<td>132.8</td>
<td>191.2</td>
<td>93.4</td>
<td>121.0</td>
<td>137.1</td>
<td>20.0</td>
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<tr>
<td>CV</td>
<td>0.60</td>
<td>0.66</td>
<td>0.84</td>
<td>0.57</td>
<td>0.67</td>
<td>0.77</td>
<td>0.16</td>
</tr>
<tr>
<td>April 2016</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>mean SWE (mm)</td>
<td>141.8</td>
<td>128.4</td>
<td>141.5</td>
<td>148.8</td>
<td>138.0</td>
<td>148.8</td>
<td>122.6</td>
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<tr>
<td>std dev (mm)</td>
<td>65.6</td>
<td>67.1</td>
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<td>60.1</td>
<td>81.2</td>
<td>87.2</td>
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<td>CV</td>
<td>0.46</td>
<td>0.52</td>
<td>0.64</td>
<td>0.40</td>
<td>0.59</td>
<td>0.59</td>
<td>0.25</td>
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Table 5. Quantity of UAS estimated SWE captured in each landcover type of Siksik Creek

<table>
<thead>
<tr>
<th></th>
<th>Distributed mean SWE (mm)</th>
<th>SWE by volume (%)</th>
</tr>
</thead>
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<tr>
<td></td>
<td>snow survey</td>
<td>method 1</td>
</tr>
<tr>
<td>April 2015</td>
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</tr>
<tr>
<td>Drift</td>
<td>48,662</td>
<td>393.5</td>
</tr>
<tr>
<td>Shrub</td>
<td>155,047</td>
<td>328.7</td>
</tr>
<tr>
<td>Tundra</td>
<td>753,783</td>
<td>71.0</td>
</tr>
<tr>
<td>April 2016</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Drift</td>
<td>48,662</td>
<td>291.4</td>
</tr>
<tr>
<td>Shrub</td>
<td>155,047</td>
<td>205.6</td>
</tr>
<tr>
<td>Tundra</td>
<td>756,485</td>
<td>94.7</td>
</tr>
</tbody>
</table>

Figure 21. PDFs of end of winter SWE for the 3 density methods. (a) density method 1, (b) density method 2 and (e) density method 3 for April 28<sup>th</sup> 2015. (d) density method 1, (e) density method 2 and (f) density method 3 for April 23<sup>th</sup> 2016. Distribution of manual observations in red and the distribution of UAS estimated in blue.
2.5 Discussion

During the period of 1958 to 2017 at Inuvik-A, snowfall, snow on the ground and percentage of total precipitation as snowfall have declined, and the lengths of winters have become shorter. Furthermore, the Intergovernmental Panel on Climate Change Assessment Report 5 (IPCC AR5), with the use of precipitation simulations from Coupled Model Intercomparison Project Phase 5 (CMIP5), are seeing mixed changes in annual snowfall over the Northern Hemisphere (Collins et al., 2013). The IPCC AR5 states that decreases in snowfall are due to a shift in precipitation phase from solid to liquid and an earlier melt onset from consequent warming in the Arctic and that increases in snowfall are as a result of higher moisture content available from a reduced Arctic sea ice extent. Both scenarios are likely the case for the Inuvik region.

At TVC, the total corrected snowfall is consistently less than the watershed SWE at the end of winter. This is similar to Woo et al. (1983) who also found that snowfall was smaller than snow on the ground at the end of the accumulation period in the Canadian high Arctic. For various hydrological and ecological needs, this clearly indicates that there are still significant unknowns in our understanding of the Arctic snow regime. These may include the following. Firstly, the catch efficiency equation used to correct for wind induced undercatch, from Smith (2007) and Pan et al. (2016), may be underestimating the precipitation, and therefore a new equation for Arctic landscapes would be necessary. Secondly, throughout winter there may be a significant water vapour flux from the soil to the snowpack (Santeford, 1978; Woo, 1982). Such vapour fluxes are dependent on soil characteristics and fall soil moisture, in interior Alaska, Santeford (1978) measured 30 mm of SWE transferred from the soil to the snowpack representing
25 to 30% of the moisture already in the snowpack. In Resolute, NU, Woo (1982) measured an upward flux of 2 to 3 mm of SWE for a 0.5 m snowpack. This phenomenon adds additional SWE to the snowpack without a snowfall source and needs further monitoring. Thirdly, sublimation occurs throughout winter, in greater amounts during blowing snow events (Pomeroy et al., 1997), reducing the quantity of snow on the ground by approximately 20% (Pomeroy et al., 1997). It is still difficult to capture sublimation from eddy covariance systems running all winter long at a very high frequency due to the lack of power during the long dark winter periods. Fourthly, snow surveys may over/under estimate snow on the ground. Unfortunately, our understanding of the magnitude, and in some cases direction of the errors, of each of the above is poorly understood. This is a significant problem greatly limiting both our understanding of the current snow environment of the Arctic and ability to predict future changes to the snow environment.

With the use of UAS, it was confirmed that traditional snow surveys provide reasonable estimates of basin average SWE, this will provide us some confidence in the long-term record of snow surveys. UAS estimates may not be better than traditional snow surveys if all that is needed is a basin average SWE. However, UAS estimates provide a better way for mapping SWE across the landscape at fine resolution, which is very important for hydrological model testing and development. In terms of UAS repeatability, the end of winter 2015 UAS snow depths weren’t significantly representative of the observations, while the 2016 UAS snow depths showed greater representation of the observations. This was shown by the overall snow depth RMSE of 49.0 cm and 27.6 cm for 2015 and 2016 respectively. In comparison to other studies using UAS
photogrammetry, Vander Jagt et al. (2015), De Michele et al. (2016) and Harder et al. (2016) have obtained RMSEs of 9.6 cm, 14.3 cm and 13.7 cm respectively. However, these three studies were conducted on smaller scales, 0.007 km$^2$, 0.3 km$^2$ and 0.65 km$^2$ respectively, whereas our site is approximately 1 km$^2$ in area. In addition, Vander Jagt et al. (2015) flew a multi-rotor UAS, meanwhile, De Michele et al. (2016) and Harder et al. (2016) both flew a fixed-wing UAS.

Obtaining spatial estimates of density at a high resolution on Arctic landscapes is quite difficult. Firstly, snow coring bulk density data is labour intensive and requires a great amount of time to complete (DeBeer and Pomeroy, 2009; Harder et al., 2016), secondly, numerical modelling could be a possibility, however they don’t work well in depth hoar rich snow conditions (Barman and Jain, 2016). Therefore, the three snow density distribution methods that were chosen illustrate different techniques at obtaining densities on a spatial scale. While we have only tested three methods of spatially distributing densities across the landscape, each one has its advantages and disadvantages. Method 1 is simplistic and would work well at a larger scale, however because it uses a single mean density, it misrepresents the shallow and deeper snowpacks and therefore would not be great for high resolution scales. Method 2 did not represent the density spatially well even though the densities were from the 2015 and 2016, the IDW interpolation was not great due to the few number of densities taken over the catchment. Although method 3 is more complex, it performs well at representing the densities of shallow and deeper snowpacks, and seems to be the best method for high resolution mapping. Others have used computer density models to densify the snow throughout the accumulation season to acquire distributed spring densities (Jonas et al., 2009; Bormann
et al., 2013). This technique could be useful in the future for it is less labour intensive than the collection of snow densities every year, however the computer density models do not perform very well on the Arctic landscape because of the vast amount of depth hoar overlying the hummock terrain (Barrere et al., 2017). These three methods were primarily used to acquire SWE distributions by converting snow depth estimates to SWE with each distributed density method.

With the use of UAS estimated SWE maps and traditional snow surveys it was shown that terrain dominated drifts can accumulate almost three times their area and vegetation dominated drifts almost one and a half times their area as seen in Table 6. Together they can account for 39 to 56% of the total basin SWE, for only occupying a combined area of approximately 21%. The results show terrain and vegetation (shrubs) dominated drifts on this landscape can hold 5 and 4 times more SWE than the tundra respectively in terms of equivalent area (Table 6). Terrain dominated drifts in Siksik Creek captured 14% and 17% of total winter precipitation. Sturm et al. (2001b) found similar results, as much as 15% of the winter precipitation is captured in Alaskan drifts. Moreover, terrain dominated drifts, vegetation dominated drifts and tundra accumulated 282%, 209% and 80.5% of total snowfall on average respectively for the 2-year study in Siksik Creek.

Finally, by comparing the two study winters we see that 2014/15 had more possible blowing snow events throughout the winter than 2015/16, due to a greater number of wind speeds larger than the 7m/s blowing snow threshold (Li and Pomeroy, 1997). This reduced the amount of SWE held in tundra and increased the amount of SWE held in drifts and shrubs, resulting in larger CV. In contrast, the winter of 2015/16 saw
less total snowfall, however the tundra gained SWE over the previous winter. Therefore, fewer blowing snow events must have occurred in 2015/16, causing a reduction in wind scoured tundra (lighter winds seen in 2015/16; Figure 8c and d) and consequently less SWE deposited into drifts and shrubs.

2.5.1 Limitations in snowfall

Measuring snowfall is very difficult since snow does not always fall during similar weather conditions (i.e. temperature, atmospheric pressure, wind speed and direction) and must be corrected for wind-induced undercatch in this harsh climate. However, anemometers tend to slow or even completely stop when ice builds on the

### Table 6. Landcover and snow characteristics summary for Siksik Creek.

<table>
<thead>
<tr>
<th></th>
<th>Drift</th>
<th>Shrub</th>
<th>Tundra</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slope (°)</td>
<td>8 - 15</td>
<td>1 - 6</td>
<td>1 - 4</td>
</tr>
<tr>
<td>Aspect</td>
<td>E - NE</td>
<td>E - SE</td>
<td>all</td>
</tr>
<tr>
<td>Area (m²)</td>
<td>48662</td>
<td>155047</td>
<td>756485</td>
</tr>
<tr>
<td>Area percentage (%)</td>
<td>5.1</td>
<td>16.1</td>
<td>78.8</td>
</tr>
<tr>
<td>Shrub species</td>
<td>Alnus</td>
<td>Alnus &amp; Salix</td>
<td>Lichen &amp; Betula</td>
</tr>
<tr>
<td>Shrub height (m)</td>
<td>1 - 1.5</td>
<td>0.8 - 2.5</td>
<td>0 - 0.5</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>April 2015</th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Depth (cm)</td>
<td>110.6</td>
<td>101.2</td>
</tr>
<tr>
<td>Density (g/cm³)</td>
<td>0.354</td>
<td>0.323</td>
</tr>
<tr>
<td>SWE (mm)</td>
<td>393.5</td>
<td>328.7</td>
</tr>
<tr>
<td>SWE (m²)</td>
<td>19148</td>
<td>50964</td>
</tr>
<tr>
<td>SWE (% by vol.)</td>
<td>15.5</td>
<td>41.2</td>
</tr>
<tr>
<td>SWE ratio to tundra</td>
<td>5.54</td>
<td>4.63</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>April 2016</th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Depth (cm)</td>
<td>93.2</td>
<td>83.2</td>
</tr>
<tr>
<td>Density (g/cm³)</td>
<td>0.300</td>
<td>0.245</td>
</tr>
<tr>
<td>SWE (mm)</td>
<td>291.4</td>
<td>205.6</td>
</tr>
<tr>
<td>SWE (m²)</td>
<td>14180</td>
<td>31878</td>
</tr>
<tr>
<td>SWE (% by vol.)</td>
<td>12.0</td>
<td>27.1</td>
</tr>
<tr>
<td>SWE ratio to tundra</td>
<td>3.08</td>
<td>2.17</td>
</tr>
</tbody>
</table>
rotating cups or propeller as seen in Figure 22 and from data in Figure 8 from January 8\textsuperscript{th} 2016 to February 17\textsuperscript{th} 2016, therefore our wind speed readings may in fact be much larger in reality. For this period with lack of measured wind speeds due to frozen anemometers from TMM and surrounding stations, wind speeds from Inuvik, NWT were used to correct snowfall. Other errors observed came from the permafrost thaw under the Geonor gauge at TMM during heavy rainfall events of August and September 2015. This tilted the gauge to approximately 10 degrees which increased the amplitude of the signal noise in the raw data, making it much harder to pick out snowfall events to wind pressure error. A combination of ice-build up and drifting snow accumulated on the rim of the gauge and partially covered the orifice of the Geonor gauge (Figure 23) causing accumulation of precipitation to decrease or to completely cease, this is seen in the data at the forest site (Figure 8g and h).

2.5.2 Limitations in end of winter snow cover

Measuring snow depth, density and SWE is no easy task across a large area. Sampling these three snow parameters is affected by various
snowpack properties and underlying vegetation properties (i.e. ice layers, depth hoar, organic material; Berezovskaya and Kane, 2007). Berezovskaya and Kane (2007) showed that when probing the snowpack an overestimation of 11 to 31% in snow depth occurs due to penetration of the probe into the low density organic snow substrate (moss and lichens) overlaying the impermeable frozen ground. They also found overestimates of up to 20% when measuring SWE. This overestimation error is difficult to correct given that the error varies with the multiple personnel sampling the snowpack (Berezovskaya and Kane, 2007).

2.5.3 Limitations in spatial variations of end of winter SWE

The photogrammetric method for extracting snow depth and characterizing SWE, based on snow depths retrieved and a variety of methods to spatially distribute density, have many strengths and weaknesses. Most studies extract a UAS snow-free DSM from their UAS snow-covered DSM to obtain snow depths, whereas this study uses a snow-free bare ground LiDAR instead of the UAS snow-free DSM. This was done to be able to rule out most of the errors that would come from vegetation on the landscape; however, there are still errors due to minimal vegetation above the snow surface during the UAS snow-covered flight. The main sources of error in the UAS derived snow maps are from the vertical resolution of the LiDAR DEM and UAS DSMs of approximately 0.05 m each overtop of the GCPs. Errors between the UAS derived snow depths and the observed MagnaProbe snow depths can be quite large as seen as depth profiles in Figure 24. This error is due to the MagnaProbe’s GPS horizontal accuracy of 5 m (Nolan et al., 2015), while the UAS DSM has been georeferenced with RTK and has a much higher horizontal
accuracy of just under 0.05 m. Other factors that may have influenced the UAS derived snow depths consist of the type of UAS and weather conditions. As previously mentioned, the UAS used in this study was a fixed-wing device, which allows for larger areal extent mapping, however is more susceptible to instability in stronger winds than the multi-rotor devices. Sun angle also played an important role in capturing the snow surface because shadows were cast on the snow surface behind steep slopes or within snowmobile tracks.

Figure 24. End of winter snow survey depth profiles of three sites in April 2016. Comparing manual observed snow depth (red) to UAS estimated snow depth (black).
2.6 Conclusion

To conclude, changes in snowfall, end of winter snow cover and spatial variability have been explored. Over the past 60 years, snowfall has been decreasing in Inuvik, NWT, similarly over the past 26 years, end of winter SWE has also been decreasing in TVC. However, these two parameters hold many limitations in methodology. This thesis focussed on a solution to measuring snow depth, density and SWE at a high resolution, in turn, to quantify the importance of snow drifts on total watershed snow cover.

The research results indicate that UAS photogrammetry can be used to acquire accurate snow depth estimates and spatial variability of snow at high spatial and temporal resolutions. Here, a small fixed-wing UAS created DSMs and snow depth maps through SfM techniques of photographs of the Siksik catchment and was tested in its ability to estimate snow depth and spatial variability over an Arctic tundra landscape. Furthermore, the estimated snow depths from the UAS were combined with three spatially defining snow density methods to acquire spatial variability in SWE across the Siksik catchment at a high resolution. This new method at obtaining spatial variation of SWE proves to be helpful in understanding local scale snow processes and can allow us to monitor SWE spatially throughout the melt period with multiple UAS flights over the catchment. These distributions of snow depth and SWE are necessary for validation of hydrological models, climate models and remote sensing techniques.

The resulting UAS SWE maps illustrated that drifts (both terrain and vegetation controlled) can account for 39 to 56% of the total basin SWE, for only occupying a combined area of approximately 21%. The observations presented in this study indicate that fixed controls, topography and vegetation, are key factors in the quantity of SWE that
is redistributed and deposited in depressions or trapped in ground cover (Rees et al., 2014). On another note, traditional snow surveys do provide reasonable estimates of basin average SWE, however UAS imagery provide a better way of mapping SWE at fine resolutions and relatively low costs. The next step for future studies is to quantify the snow cover variability between the land classes throughout the entire winter instead of only at the peak accumulation. This is necessary to document snow cover properties and to understand how the different drift features form.
Chapter 3: Conclusion

3.1 Summary

In this thesis, changes in the snow environment and the role of snow drifts in the Trail Valley Creek watershed were investigated. Using new remote sensing technology, a small fixed-wing UAS, DSMs and snow depth maps were created with the use of SfM techniques. The UAS estimated snow depth maps were then combined with spatially defining snow density methods across a sub-catchment to acquire spatial variability of SWE. This new method of obtaining spatial variation of SWE proves to be helpful in the understanding of local scale snow processes. These distributions of snow depth density and SWE are necessary for validation of hydrological models, climate models and remote sensing techniques. A description of the main results of this thesis follows.

In chapter 2, the results from changes in snowfall, end of winter snow cover and spatial variability have been explored. It is shown that in the previous 60 years there has been a decline in snowfall and in the percentage of yearly precipitation fallen as snow in Inuvik, NWT. A decline in maximum annual snow depth and a reduction in the length of winters in Inuvik are also present during this period. Furthermore, this study shows that there is a slight decrease in end of winter SWE stored in the TVC watershed.

Within the forest-tundra ecotone, the distribution of vegetation and topography play a dominant role in defining the snow distribution patterns of the landscape, to the extent that terrain and vegetation dominated drifts can hold up to 5 and 4 times more SWE than tundra respectively in terms of equivalent area. This end of winter snow distribution pattern is dependent on wind speed, direction and weather patterns throughout the winter months. For this study, generation of snow depth and SWE maps,
with the use of UAS imagery, was necessary to locate and calculate the area and quantity of SWE stored in these drift features. This, in turn, was used to illustrate the importance of drifts on Arctic landscapes.

3.2 Suggestions for future work

Regardless of the changes in snowfall, spatial variability and end of winter snow cover documented in this thesis, some aspects affecting the latter remain to be fully explored. For instance, the ability to measure sublimation in Arctic environments at high frequencies using eddy covariance systems and to fully understand the soil to snow vapour flux throughout the full winter period. These two processes could influence the end of winter snow cover significantly and, with the addition of a blowing snow model, a snow mass balance for the basin would be feasible.

Additional spatial densification methods could be tested, namely using numerical model densification to obtain spatial variability of density. This method can then be compared to the methods applied in this thesis. Furthermore, with the implementation of these spatial densification methods and the application of UAS, new studies are now possible to capture changes in snow covered area, SWE and basin water storage across the snowmelt period at higher spatial and temporal resolutions. Another possible practical use for UASs would be to measure normalized difference vegetation index (NDVI) with different wavelength band cameras. Not only are UASs very efficient at capturing these changes but they are also a very cost-effective tool with a comparable accuracy to traditional measurements.
A future approach to build on this study would be to quantify the snow cover variability between the land classes throughout the entire winter instead of at the peak accumulation at the end of winter. This is necessary to document snow cover properties and to understand how and when drift features form.

At a larger timescale, the impacts of climate change on the Arctic snow cover are of great concern. Will the role of snow drifts and their holding capacity increase or decrease in a warmer environment? Will decreases in snowfall necessarily lead to decreases in drift size? Many uncertainties in the topic of snow drifts and spatial variability still exist and they present great research opportunity for the hydrometeorological community.
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