THESIS

# THE FATE OF DINWOODY GLACIER: PRESENT STATE OF MASS BALANCE AND DOWNSTREAM IMPACTS OF GLACIER RUNOFF.

Submitted by

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# ABSTRACT


THE FATE OF DINWOODY GLACIER: PRESENT STATE OF MASS BALANCE AND DOWNSTREAM IMPACTS OF GLACIER RUNOFF.

The Wind River Range in Wyoming supports many of the few remaining continental glaciers of the North American Rocky Mountains; the glacier meltwater runoff feeds four major river systems within the U.S. West. Runoff from glaciers affects downstream ecosystems by influencing the quantity, seasonality, and chemistry of the water. We describe the present state of Dinwoody Glacier, the fourth largest glacier in the Wind River Range. We utilize photogrammetry, snow depth measurements, and ablation measurements to characterize surface mass balance for summer of 2017. Localized and nearby stream gauge measurements help to quantify glacial meltwater runoff inputs to Dinwoody Creek. Both of these methods allowed us to put the changes of the Dinwoody Glacier into the broader context of the Missouri River Watershed. If melted, Dinwoody Glacier would no longer provide a reliable source of melt water for thousands of people living in the Missouri River Watershed. Understanding how shrinking glaciers and decreasing melt-water runoff will impact communities and ecosystems downstream is critical for effective environmental management. The response of the Wind River glaciers to future climate is uncertain; however, past research has shown declines in glacial mass, snow cover, snowmelt timing and stream power. The data we collected in the summer of 2017 tells the story of a quickly diminishing and critical resource despite 2017 being a uniquely wet and cold year. While glacier meltwater runoff contributions to Dinwoody Creek were above average, the Accumulation Area Ratio for Dinwoody Glacier in 2017 was 21% suggesting a glacier in severe recession.

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# Chapter 1

# Introduction

Runoff of melt water from glaciers is an important source of water and nutrients for ecosystems and people living downstream of glaciers (Kaser et al., 2010). Mountain glaciers across the world are expected to lose mass this century. Glaciers in Western Canada and the continental U.S. are expected to lose 85% of their volume by 2100 (Radic et al., 2014) and as a result, glacial runoff in the region will decrease (Bliss et al., 2014; Marks et al., 2015). Recent research on glacial volume and runoff decline has focused on larger areas, but runoff projections for individual glaciers and watersheds are scarce (Casassa et al., 2009). This precision in climate forecasting is critical for making informed and sustainable water management decisions on a local level (Bell et al., 2012).

The Wind River Range in Wyoming contains the largest concentration of glacial mass in the Rocky Mountains of the contiguous USA (Figure 1.1) with 116 glaciers and ice patches (RGI Consortium, 2017). The range is oriented northwest to southeast and makes up 160 km (Marks et al., 2015) of the Continental Divide. The region is characterized as a dry continental climate with annual precipitation ranging from 0.20 m at the base to 1 m at its peaks (Martner, 1986). During the winter, orographically-lifted air masses provide a mean annual precipitation as snow of 1.50-5.1 m (Vanlooy et al., 2014). These numbers have changed over time with the recent upward trend in average temperatures; the average annual temperature in the last 30 years has increased by three degrees since the 1970s (Cayan, 1996). Wind River glaciers are at a higher elevation than most other glaciers in the lower 48 States and generally receive less precipitation, making them more vulnerable to changes in temperature (Marston et al., 1991).

The focus of this research is Dinwoody Creek Watershed, it contains two of the largest remaining glaciers in the U.S. Rocky Mountains: Gannett Glacier and Dinwoody Glacier. Dinwoody Creek is the smallest watershed (228 $km^2$) with a stream gauge in Wind River Range, but is the most heavily glaciated (DeVisser and Fountain, 2015; Thompson et al., 2011). The waters from Dinwoody Creek are allocated through the Prior Appropriation Doctrine that allocates water on a

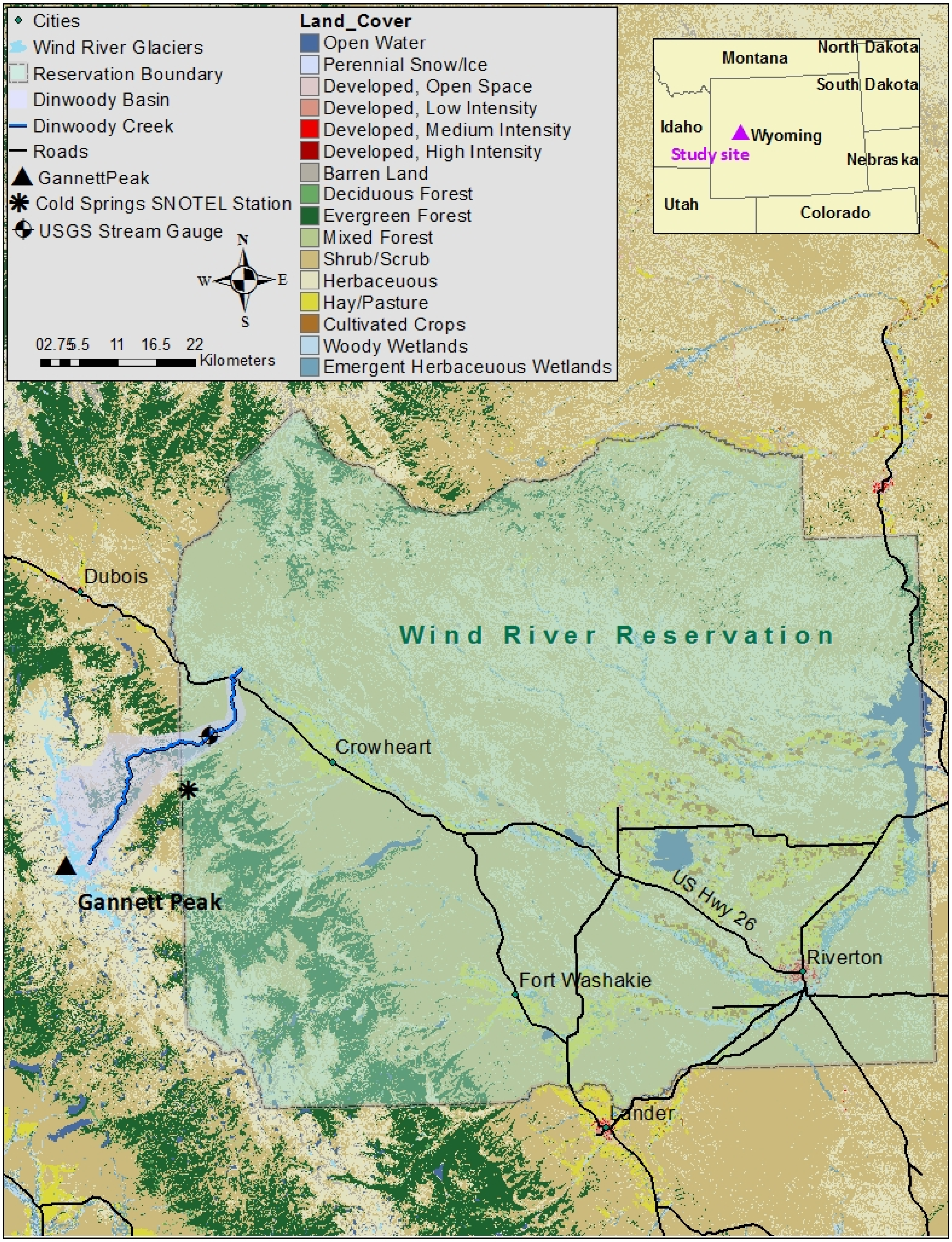



**Figure 1.1:** Land use map of the Wind River Range.

"first in time, first in right" basis (Johnson, 2015). In this case, "first in time" refers to the Eastern Shoshone and Northern Arapaho Tribal members that were forced onto the Wind River Indian Reservation in the late 1800s and early 1900s (Kinney, 1993). Today, tribal members and the adjacent ecosystems use the water from Dinwoody Creek Watershed to maintain crops, and supply for fisheries and hunting (Flanagan and Laituri, 2004). However, recent and rapid climate change has altered the reliability of the Reservation's water resources; warmer temperatures, and extreme droughts in the area have compromised the livelihoods of tribal members and local ranchers (McNeeley et al., 2017; Ojima, 2015).

The runoff from Dinwoody Glacier and its neighbors flows to the Missouri River Basin to the east, whereas the western slope of the Wind River Range flows into the Green River and the Colorado River basin (Pochop et al., 1990). Seventy-seven percent of the glacial area in the Wind River Range is located on the eastern side of the divide (Bell et al., 2012). Consequently, most of the glacial meltwater drains into the Wind River Basin which then flows into the Missouri-Mississippi River system. The Wind River Indian Reservation is the first water-right holder to the runoff; the tribal members of this reservation reside closest to the headwaters of Dinwoody Creek, in the grasslands of the eastern slope of the range (Robison, 2015).

The reliability of the Reservation's water resources largely hinge on water stored in the form of snow or ice at the highest points along the Range. Glacial bodies are often conceptualized as reservoirs that store water in the form of ice and release it at future times (Fountain et al., 1997; Jansson et al., 2003; Pochop et al., 1990). Watersheds that contain glaciers and sustain a perennial snowpack typically reach their maximum discharge rates in late summer, when downstream domestic and agricultural needs are greatest (Casassa et al., 2009; Johnson, 2015). This unique annual storage and runoff pattern acts to buffer streamflow throughout the melt season (Fountain et al., 1997; Meier and Post, 1962). Thus, watersheds containing glaciers provide a more stable source of water than non-glaciated watersheds (Braithwaite and Zhang, 1999). Glaciated watersheds have approximately 50% greater summer runoff and less runoff variability than non-glaciated basins (Fountain et al., 1997). Glacial meltwater in the Wind River Range is a critical component of the annual quantity and quality of streamflow that water managers have relied on for over a century. The cycle of water in this basin is replenished each year with winter precipitation and depleted throughout the warm months of summer. Once annual snowpack melts out, the exposed glacial ice releases its ancient air bubbles as the meltwater runs off into the alpine portion of Dinwoody Watershed; it cascades and meanders through various ecotones, and finally to water users. Though the reliability of the cycle from headwaters to tapwater has decreased since 1850, the approximate end of the Little Ice Age (Martner, 1986; Meier, 1951), rapid glacier recession in this region is a response to increased temperatures since the 1930s (Fountain et al., 2017; Kaser et al., 2010). The loss of these local glacial meltwater reserves has and will continue to cause water resource management issues on the Wind River Indian Reservation.

To improve land management practices and drought resilience in this area, regional-scale modeling is being utilized to better predict annual quantities of water storage in the upper reaches of the watershed. The accuracy of regional-scale models is dependent on a thorough understanding of the limitations of model assumptions and data inputs. One limitation to robust climate/glacial response models is the proximity and accuracy of meteorological stations. Climatic conditions such as snowfall are difficult to estimate on a regional-scale because mountainous regions have complex weather. Thus, patterns of glacial accumulation or ablation are difficult to predict even from one basin to the next (Brown et al., 2010). Owing to these variable conditions, Dinwoody Glacier may not disappear as quickly as previously anticipated.

To understand the ecological complexity and water resource management challenges of this region, this thesis will elucidate the nuances of the region's unique past. The Wind River Range is of great interest to anthropologists, historians, geologists, geographers, sociologists, paleoclimatologists and glaciologists alike because of its unique climate and near 10,000 year old history of human habitation. The research we completed remotely gives a broad understanding of the history of glacial advance and recession in this region. The data gathered in the field helps to elucidate our main research questions:

1. What is the predominant climatic trend of this region?

2. What is the present mass balance of Dinwoody Glacier?

3. How much does glacial melt water runoff contribute to stream power and stream quality in the Dinwoody Creek watershed?

This information will be used to assess the hydrologic resilience of Dinwoody Creek Watershed. The literature review will delve into the rich history of glacial advancement and recession, as well as describe the prehistoric peopling of Eastern and Western Flanks of the Continental Divide up to the Pre-Columbian Era and then into the Reservation Era. The literature review will also dive deeper into the ways in which glacier meltwater has historically flowed from the glaciers, and its importance to the settlement of the U.S. West.

# Chapter 2

# Literature review

## 2.1 Historic Climate and Peoples

It has been conjectured that Shoshone peoples have been living in the Greater Yellowstone Ecosystem anywhere from 500 years (Hultkrantz, 1987) to several thousands of years (Husted and Edgar, 2002). Climate, food-supply, and territorial contractions/expansions likely drove immigration through this region (Aikens and Witherspoon, 1986).

Regional groups of the Shoshone were named for the primary source of food in their diet (Adam et al., 2009); there were Bison-eaters, Sheep-eaters, and Root-eating Shoshone that existed in the Greater Yellowstone Ecosystem (Hultkrantz et al., 1981). It was the Wyoming Sheep-eaters that spent their summers in the Wind River Range and Absaroka Range (Hultkrantz et al., 1981; Shimkin, 1942). "Sheepeater" is an anglicized version of the Shoshone term "Tukudika". Modern descendants of Sheep-eaters, the Eastern Shoshone, steward the Wind River Indian Reservation along with members of Northern Arapaho Tribe. This section will describe the ways in which the climate and historic peoples interacted to form lifeways in and around the Wind River Range.

Climate influences geographic boundaries of human existence (Morgan et al., 2012). Important climatic fluxes in this region are marked by the Last Glacial Maximum and the Little Ice Age. Following these documented cool-periods, Euro-American migration to the Western U.S. became a predominant force of geographic boundaries for Native Americans.

#### Last Glacial Maximum

Former glacial extents as well as erosional pathways are important components to understanding modern ecosystem structure (Whitlock, 1993). In Earth's climatic history, the Last Glacial Maximum (∼21,500 years ago) is the last period of the most recent Ice Age where ice sheets were at their maximal extent (Mithen, 2004). The vast expanses of ice sheets during this time profoundly affected Earth's climate, causing drought, desertification, and a drop in sea levels (Mithen, 2004).

In the U.S. West, decreased temperatures aided the advancement of glaciers along the continental divide. The Wind River Range was upwind of the former Laurentide Ice Sheet, thus minimizing the regional climatic influence of the giant ice sheet on its glaciers (Whitlock, 1993). However, cooler temperatures and increased precipitation fostered the growth of local glaciers. Major valleys on both flanks of the Wind River Range supported large valley glaciers, some extending >50 km in length with a minimum thickness of 600 m (Blackwelder). Glacier lobes flowed from the peaks of the Wind River Range and down into the mountain forelands as piedmont lobes (Hostetler and Clark, 1997). Oscillations of these piedmont glaciers produced well-spaced belts of moraine systems that today serve as a topographic template of the landscape (Figure 2.1).

The topographic template carved by the Last Glacial Maximum served as the physical bounds of human existence in the Wind River Range however, archaeologists have argued that humans affected waterways, plant regimes, wildlife populations in ways that also shaped their geographic boundaries. Evidence of humans in this area most reliably dates back 2800-1500 BP (∼3000 years ago) (Morgan et al., 2012). The Rocky Mountains were seasonally productive habitats that produced essential plant resources like camas and pine nuts and more importantly, forage and browse for the large ungulates upon which mobile hunter-gatherers relied upon (Black, 1991).

**Little Ice Age**

The most recent advance of the Wind River Range glaciers coincided with a documented cool period, the Little Ice Age, starting locally, circa 1745 and ending in 1850 (Naftz et al., 1996; Pochop et al., 1990). A reconstruction of glacier equilibrium-line altitudes estimated a 0.2°C depression of summer temperatures and an increase in precipitation in the region (Zielinski and Thompson, 1987). One study estimated equilibrium line altitudes in the region advanced 60 m downslope during the Little Ice Age (Albanese and Frison, 1995). The Wind River Range at this time was experiencing the influx of Euro-American explorers.

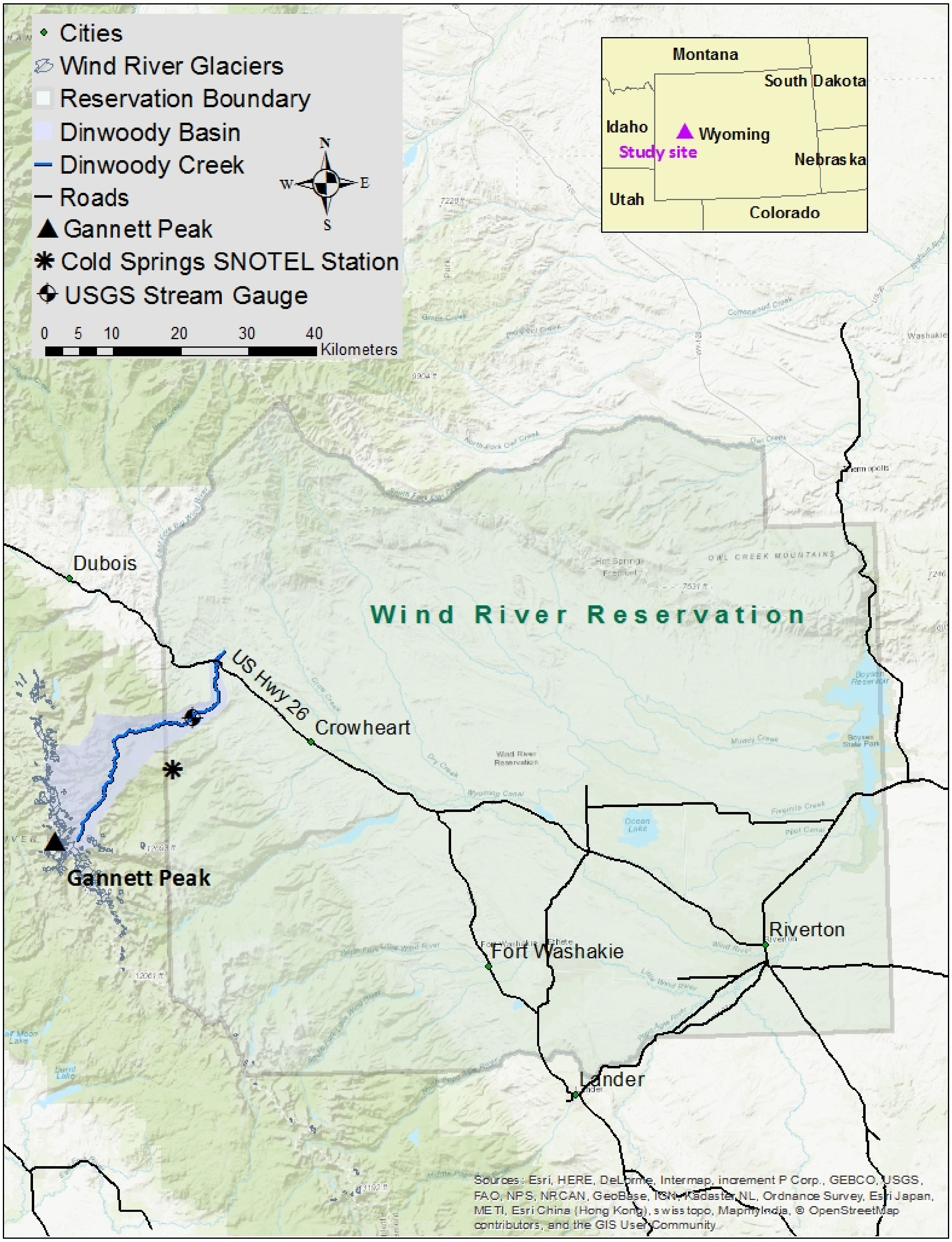



**Figure 2.1:** Topographic area map of the Wind River Range including a state-level inset map.

**Euro-Americans and the Reservation Era**

In 1805, Lewis and Clark's expedition encountered Shoshone tribes (Loendorf and Stone, 2006), the ancestors of the Eastern Shoshone tribe of the modern-day Wind River Indian Reservation. Lewis's diary describes the tribal members as, "frank, communicative, fair in dealing, generous with the little they possess, extremely honest and by no means beggarly." The explorers traded goods such as jackets and knives for three horses of the Shoshone (Loendorf and Stone, 2006).

Exploration of the U.S. West by Euro-Americans was for the purposes of western expansion (Devisser, 2011) with greater hopes of creation of commerce on western trails (The National Archives, 1843). The earliest recorded explorations were for scouting routes across the Continental Divide. In 1843, one such scouter, John Fremont and a crew of 20 men including Kit Carson, came home without a route across the continental divide but reported on the lush botany and relatively moderate climate of the West. This report changed the paradigm of many unwilling emigrants who considered western expansion impractical (Jones, 1873). During the Fremont expedition, John C. Fremont, climbed and named the third highest peak in all of Wyoming (7 km north of Dinwoody Glacier) for himself (Figure 2.2). Today the glacier on the north-side of this peak, just east of Dinwoody Glacier, also holds this Euro-American's name despite the fact that neither the glacier of this peak nor any other glacier in the region was mentioned in his personal journals (Sprague, 1964).

Western expansion reshaped the Wind River Range landscape by using new technologies, exhausting resources, and altering habitats (Hodge, 2013). The Shoshone peoples who lived on this landscape for centuries, were faced with an altered landscape that they were unable to survive off anymore (Loendorf and Stone, 2006).

In 1860, Yellowstone National Park authorities posed a different depiction of the Shoshone than that of Lewis and Clark, describing them as, "objects of pity, wretched, destitute, and feeble in mind and stature" (Jones, 1873). The descriptions given to the Shoshone by each Euro-American entity are subjective truths and should be examined with context to the environment. In the 1860's,


**Figure 2.2:** Illustration by Charles Preuss, reprinted from *A Report of an Exploration of the Country Lying between the Missouri River and the Rocky Mountains on the Line of the Kansas and Great Platte Rivers*, printed by order of the United States Senate, 1843. Source: The National Archives (1843)

a half of a century of westward movement of Euro-Americans and the discovery of gold had significantly altered Shoshone lifeways (Albanese and Frison, 1995). Larger pressure on game and fish populations, polluted streams and deforestation all contributed to an altered survival strategy. In addition, waterways were dammed and diverted, forever changing the ecosystems the Shoshone members understood (Loendorf and Stone, 2006).

In 1868, the traditional territory of the Shoshone, the Wind River Basin was established as a Reservation under the Fort Bridger Treaty of 1868. Less than ten years later, the Northern Arapahoe, a traditionally "warring tribe to the Shoshone" were forcibly settled on the reservation as well (Flanagan and Laituri, 2004).

The Northern Arapahoe are an indigeneous people of the central plains area of North America (Anderson, 2001; Flanagan and Laituri, 2004). Evidence of their history west of the Missouri River is scarce prior to the 17th century (Toll, 1962). In the 1870s, the U.S. Government considered the members of the Northern Arapahoe as neither having been "aggressively hostile or generously aiding" to Euro-American presence and settlement of the West (Anderson, 2011). Due to this impartiality, the Northern Arapahoe were not awarded their own Reservation by treaty and were forced to share a Reservation with another Tribe. The Northern Arapahoe resisted placement in low-lying territory due to visions of great floods by the most respected elder of the tribe and cooperated with the U.S. Military for placement at the Wind River Indian Reservation (Anderson, 2001).

### The Discovery of Glaciers

The first written record of the glaciers of the Wind River Range is credited to a group of geologists working for the U.S. Geological Survey, led by F.V. Hayden in 1877 (Hayden, 1878). The first published scientific observations of the glaciers of Dinwoody Creek Watershed was in 1931 by Wentworth and Delo (1931); their work described Dinwoody, Gooseneck and Gannett Glaciers. In 1951, Meier (1951), a group of geologists from the University of Iowa, attempted to inventory the glaciers of the Fremont Peak area (Figure 2.3) utilizing maps that they conjectured to "omit many and locate glaciers, some in improbable places. The generalized mapping is most inaccurate west of the continental divide."

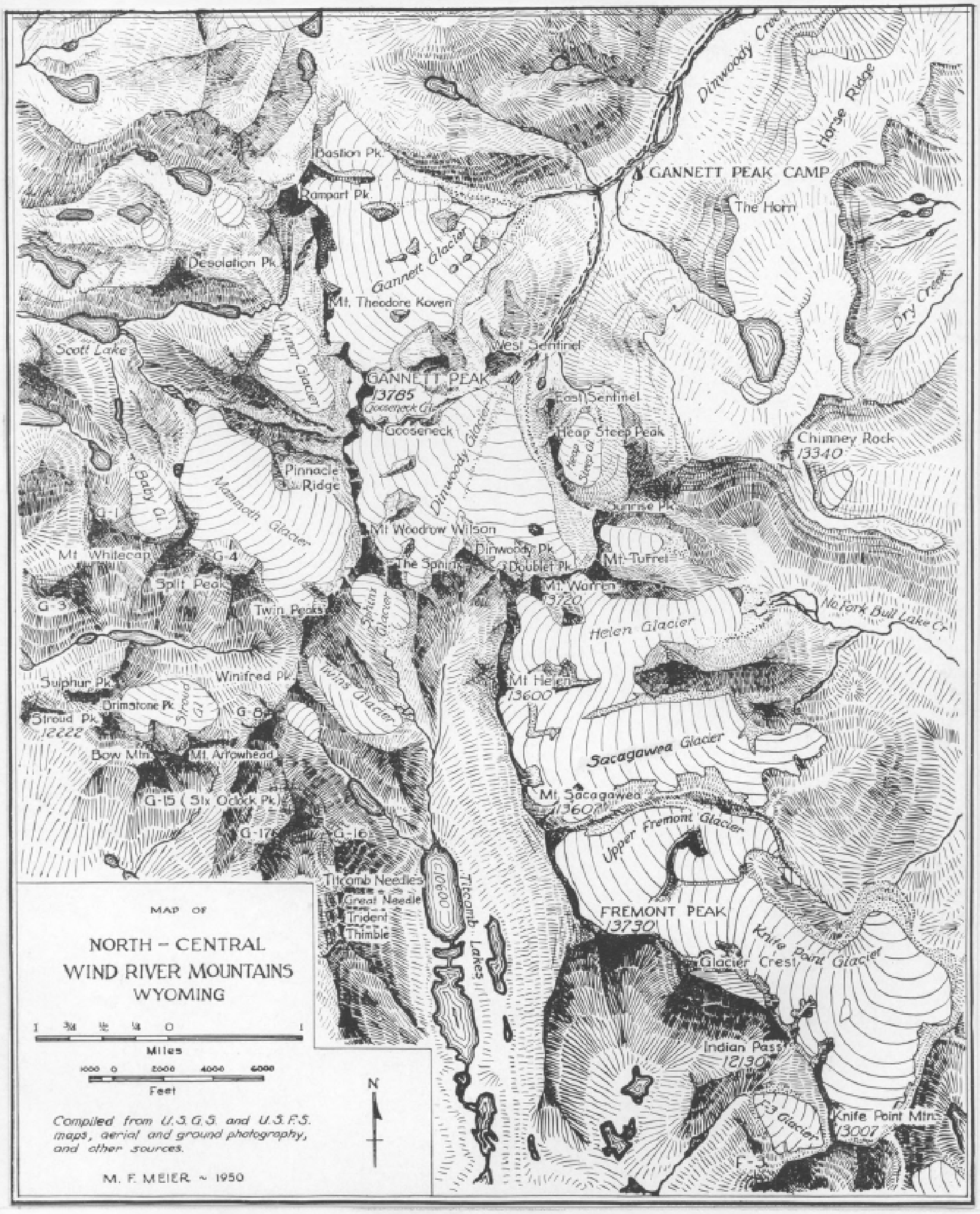
Modernly, the Wind River Range is a highly managed landscape with large populations dependent on its natural resources, namely, water (Tribal Water Engineer, 2017). The next section will describe how the resources provided by this landscape have been utilized to quench the thirst of its many downstream users.

## 2.2 Glacial ablation

Ablation is the removal of snow and ice from a glacier through melt or evaporation; it is a function of temperature, amount of radiation inputs, slope, shading and surface roughness (Fountain et al., 1997; Slaymaker, 1991). The historic seasonal trends of this region illustrate a landscape that is dependent on runoff from snowpack and glacial meltwater runoff. For Dinwoody Creek Watershed, glacial melt water is a large determinant of water availability (Hunter et al., 2006; McNeeley and Beeton, 2017); in years with higher temperatures, glacial melt water runoff increases (Hall et al., 2015; Jost et al., 2012). Ablation and the subsequent runoff from the surface and interior of the glacier, evaporation of liquid water from the glacier surface, and sublimation all lead to decreased mass of the glacier.

## 2.3 Water management on the Wind River Indian Reservation

When the Wind River Reservation was first established in the 1868, the US government's intention was to turn the nomadic tribes into farmers (Edwards and Wyoming public radio., 2014). Through the Winter's Doctrine (1908), a U.S. Supreme Court ruling, Reservations were appropriated water based on their treaty date (Shimkin, 1942). This ruling gave the Eastern Shoshone and Northern Arapaho the most senior rights to the water; however, their right remains tied to agricultural purposes (Getches, 1997). This relic of the Reservation Era means that the tribes cannot use the water for instream flows, traditional and cultural uses, riparian ecosystems, fisheries or various wildlife species (Robison, 2015). In addition, water rights do not extend to ground water. The




map courtesy The Iowa Climber

Fig. 2

**Figure 2.3:** Hand drawn map of known glaciers as of 1950 (Meier, 1951).

legacy of these water-use rights severely limits the ability of this region's tribes to adapt to changes in climate (McNeeley et al., 2017).

Three of the most pivotal legal arrangements in Wind River Reservation history are the Indian Removal Act of 1830, the Dawes Act of 1887 and the McCarren Amendment of 1952. The Indian Removal Act gave the U.S. president authority to exchange Indian land, and forced removal of tribes to the U.S. West. The Dawes Act provided legal authority to the U.S. Government to insidiously subdivide reservation land and allowed parcels of Reservations to be sold to Euro-Americans (Ford and Giles, 2015) (Figure 2.4). Finally, the McCarren Amendment allowed for States to adjudicate federally reserved water rights. The result of these legal arrangements is a patchwork of ownership for both Reservation land and resources (Flanagan and Laituri, 2004).

The legal rights to distribution of the water of this region belong to governing bodies outside Indian governance (Flanagan and Laituri, 2004). Specifically, the State of Wyoming, the Bureau of Reclamation, and the Bureau of Indian Affairs have ultimate decision-making power of water administration (Thorson et al., 2005). The governing body with the greatest ability to influence administration of water is the Bureau of Indian Affairs; however, the Federal Indian Code they follow contradicts the prior appropriation doctrine in one important way: the Federal Indian Code requires that "during times of drought, the deficits are experienced and shared between users equitably" (Burton, 1991). Throughout history, the Federal Indian Code has been overturned in U.S. courts of law in favor of U.S. Government policies.

### Legal frameworks of water rights on the Wind River Indian Reservation

There are three types of water rights in this region (Robison, 2015):

1. Tribal reserved water rights with a priority date of 1868,
2. Walton rights and
3. State based rights

Tribal rights are the most senior; however, a Walton right may also carry a priority date of 1868 without the necessity of being a tribal member. State-based rights are junior water rights that have

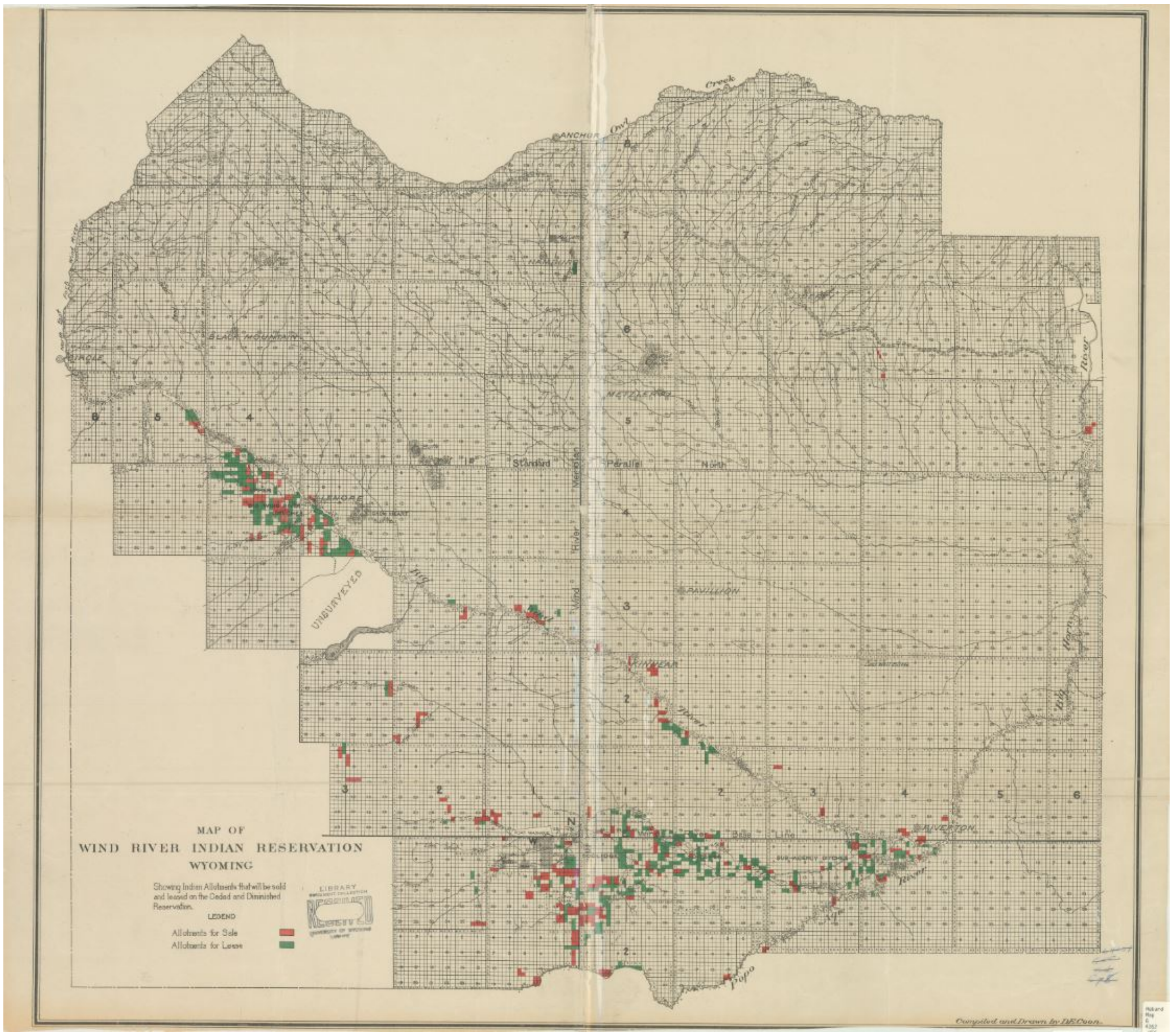



**Figure 2.4:** Map of Wind River Indian Reservation allotments to be sold or leased after the Dawes Act of 1887 was ratified in 1887

priority dates ranging from 1905 to 2012 (McNeeley and Beeton, 2017). The Wind River Indian Reservation is afforded approximately 499,862 acre-feet to be utilized on tribal lands (US Army Corps of Engineers, 2004).

The adjudication determining water availability in this region is: The Bighorn General Stream Adjudication, alterations of which in 1977 placed the legal responsibility in administering tribal and non-tribal water rights on the Wind River Indian Reservation in the hands of The State (Robison, 2015). Currently, The State allows the Tribal Water Engineers' Office to assess the hydrological conditions for the irrigation season and will deliver water according to prior appropriation. A limit to the Tribal Water Engineer however, is that they deliver water within Bureau of Indian Affair managed projects and it is they, not the Tribal Water Engineer, that ultimately determines when to release water (McNeeley and Beeton, 2017). The nuances of water administration in this region determine the annual water availability regardless of the annual quantity.

**Barriers to water security on the Wind River Indian Reservation**

Several adverse factors converge to contribute to inadequacies in the physical infrastructure of the Bureau of Indian Affairs managed Wind River Irrigation Project; they are significant barriers to effective water management in this region (Edwards and Wyoming public radio., 2014). Infrastructure for water conveyance and delivery was built in the 1920s and today, the estimated replacement and repair costs are upwards of $93 million (US EPA). Other issues include: variable reservoir capacity, rising fees for Operation and Management, leaky canal systems and a lack of monitoring systems at head gates (McNeeley and Beeton, 2017). All issues with infrastructure contribute to inefficiencies in the system (Edwards and Wyoming public radio., 2014).

# Chapter 3

# Study site

## 3.1 Overview

Dinwoody Glacier sits on the rain-shadow side of the Wind River Range, just below Gannett Peak (4208 m), Wyoming's highest point (Figure 3.1); it ranks as the fourth largest glacier in the American Rocky Mountains (Wolken, 2000). Four separate archean gneiss and granitic cirques flank the upper reaches of Dinwoody Glacier (Oswald et al., 2015). The glaciers of these cirques flow into a gently sloping central cirque basin with the total elevational gradient of 700 m from peaks to terminus (Wolken, 2000).

In contrast to much of the world's rapid melt of mountain glaciers, small cirque glaciers are uniquely resilient (Haeberli et al., 2007). Cirque glaciers, such as Dinwoody Glacier, are often protected by large rock headwalls that reduce the amount of direct solar radiation, and often receive extra snow accumulation through drifting (Kuhn, 1996). Previous investigators have predicted the demise of Dinwoody Glacier by 2018 (Marston et al., 1991). Yet today, the glacier remains. What Marston et al. (1991) failed to intimate is that Dinwoody Glacier is a north-facing cirque-glacier, increasingly shielded from direct solar radiation as it recedes (Brown et al., 2010).

During the melt season, the snow and glacial meltwater runoff from the basin of cirque glaciers, flows from the terminus of Dinwoody Glacier and then through a steep-walled U-shaped valley. Several other glacial streams join with Dinwoody Creek down-valley where it becomes a tributary of the Wind River. As you move down river from the headwaters, the ecotone transitions from periglacial to alpine to upland range. When the gradient lessens, the meanders of the Wind River travel through shrublands, grasslands, and finally, desert. Wind River Range plant types, wildlife and decomposers are all greatly dictated by their elevational gradient and disturbance regime (Chapin III et al., 2017). The vegetation of the montane and subalpine zones of the range is dominated by coniferous forests and large treeless "parks" consisting of sagebrush, grasses and

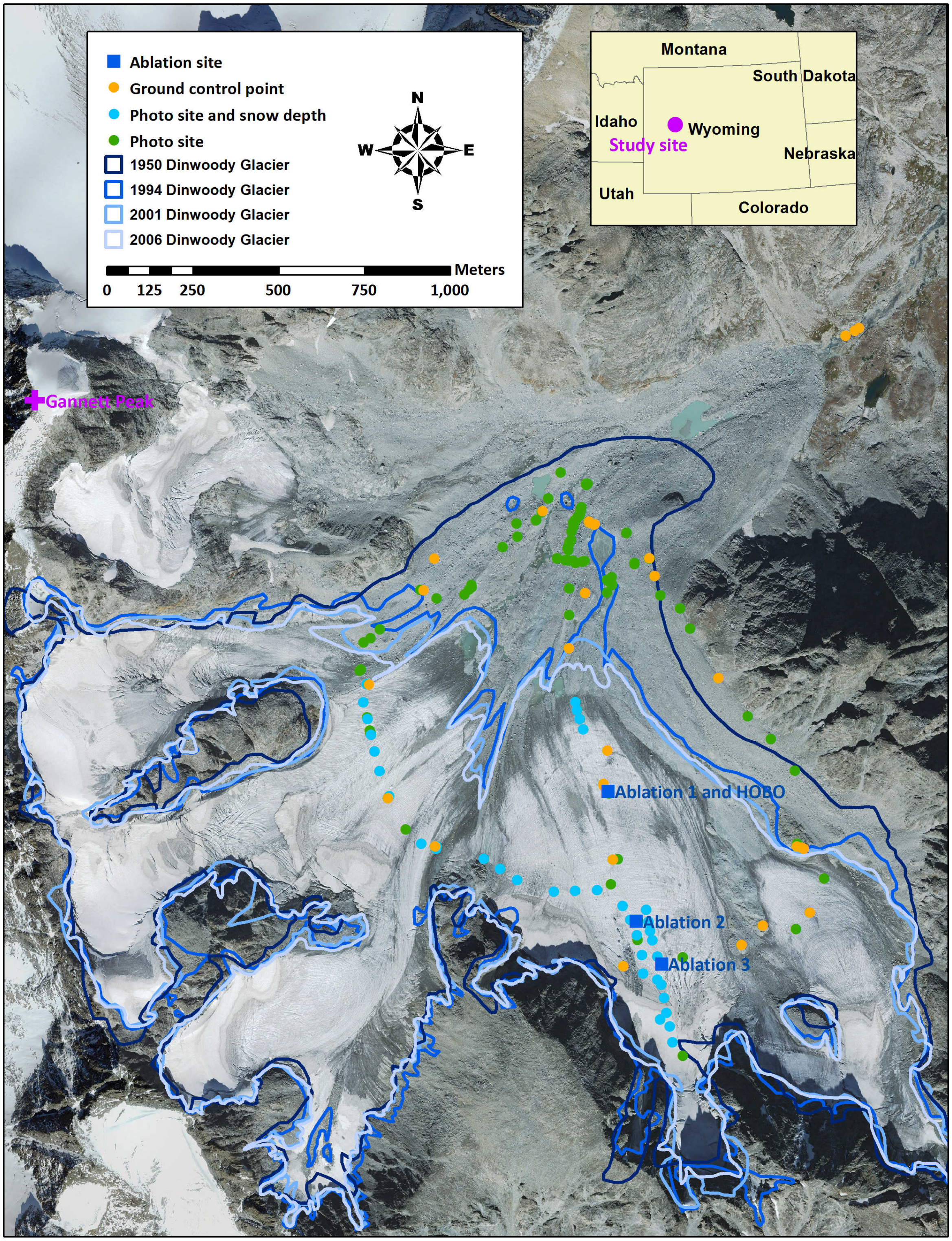



**Figure 3.1:** Map of Dinwoody Glacier and our measurement locations. Background image from National Agricultural Imagery Program (NAIP) 2015/9/10.

forbs (Lynch, 1998). The vegetation of the lower elevations are highly dependent on the amount of precipitation each year. All weather measurements in the Range are recorded in subalpine zones (Figure 3.2).

### 3.1.1 Study site climate

Historically, the climate of the Wind River Basin has been characterized as cold and arid, with some areas of cold desert (Koppen and Gieger, 1954). Precipitation in this region is generated by orographic uplift of mid-latitude transient storm systems (Whitlock, 1993). This orographic uplift results in snowfall-dominated upper reaches of the watershed and rainfall-dependent ecosystems at lower elevations. The water equivalent of the amount of snow that falls per season is recorded at 17 SNOTEL sites across the Range (Figures 3.2, 3.3, and 3.4). Annual variability of precipitation is attributed to variable storm tracks that are "driven by large-scale state factors like the temperature of the Pacific Ocean" (Cayan et al., 2016). For example, El Niño events tend to manifest as wetter than normal cold-seasons over the region (Trenberth and Hurrell, 1994).

### 3.1.2 Climate change in the Wind River Range

The historic seasonal trends of this region illustrate a landscape that is dependent on runoff from the snowpack and glacial melt. For Dinwoody Creek Watershed, glacial melt water is a large determinant of water availability (Hunter et al., 2006; McNeeley and Beeton, 2017); in years with higher temperatures, glacial melt water runoff increases (Hall et al., 2015; Jost et al., 2012). The increased glacial melt water runoff sustains stream flow through the hottest months of the year but at the cost of decreased glacial mass overtime (Cheesbrough et al., 2009).

Regional climate projections suggest increased temperature and evapotranspiration as well as altered magnitude and timing of precipitation events (Rice et al., 2012). The most comprehensive regional climate model including the Wind River Range was completed by Rice et al. (2012). The climate model captured the entire Shoshone National Forest and then was subdivided into three drainages (Figure 3.5) that contribute to the Missouri River Basin Watershed: Yellowstone River Drainage, Big Horn Drainage and Wind River Drainage. Climate data from 6 climate stations

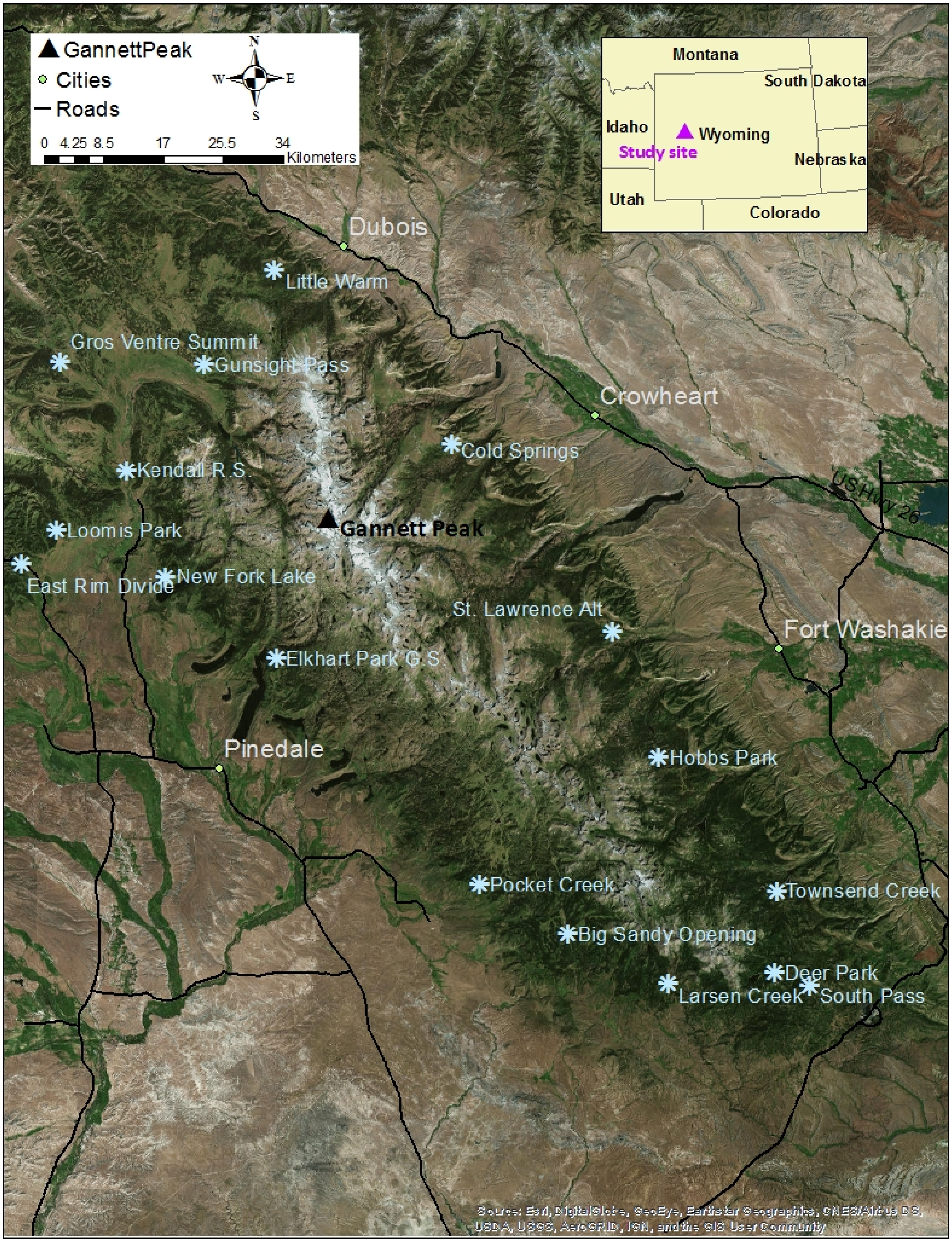



**Figure 3.2:** Map of 17 Snotel Sites surrounding the Wind River Range.

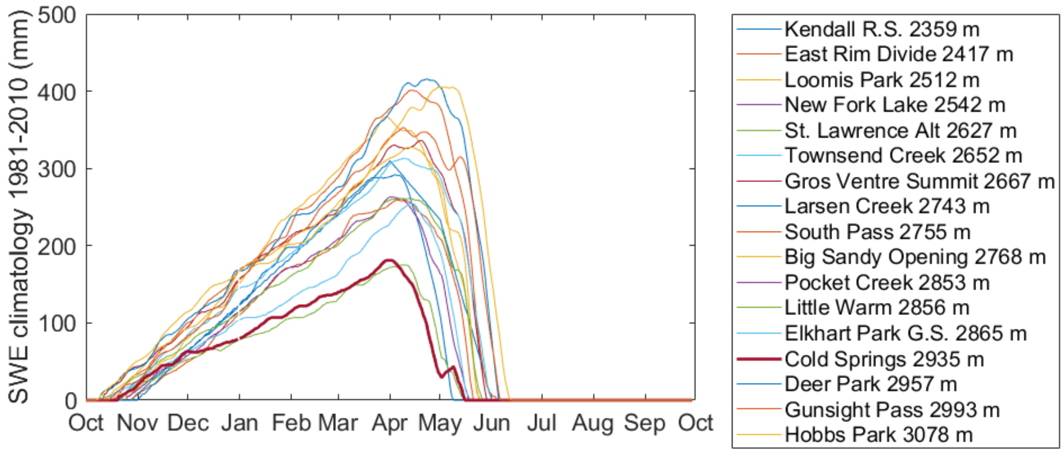



**Figure 3.3:** Seasonal patterns of snow water equivalent for all SNOTEL sites across the Wind River Range from 1981-2010.

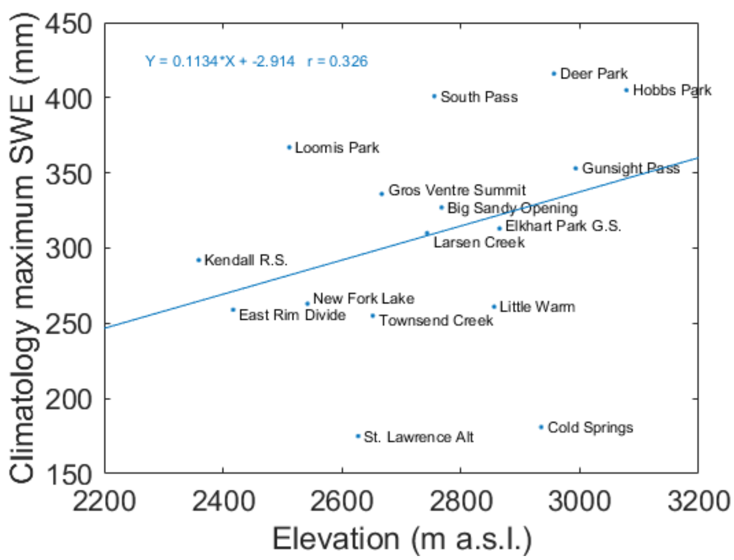



**Figure 3.4:** The amount and type of precipitation recorded at each SNOTEL site across the Wind River Range is correlated with the elevation of the SNOTEL site (r= 0.326). SNOTEL sites at higher elevations generally have higher snow-water equivalents however, the East side of the Range is considered to be a rain-shadowed climate, which results in lower snow w.e. measurements.

managed by the Western Regional Climate Center (1986-present), and 20 snowpack telemetry (SNOTEL) stations were used to identify finer spatial patterns and compute 30-year averages over the Twentieth Century (1901-1930 to 1971-2000). Rice et al. (2012) found that temperatures increased 0.6°C across each drainage, however, precipitation patterns varied widely, reflecting the topographic differences as well as their relationship to summer-wet and winter-wet regimes (Rice et al., 2012). For the purposes of this research, we will only describe the data associated with the Wind River Drainage- of which, the 30-year precipitation averages for most months decreased, with the largest decreases seen in early fall and late spring.

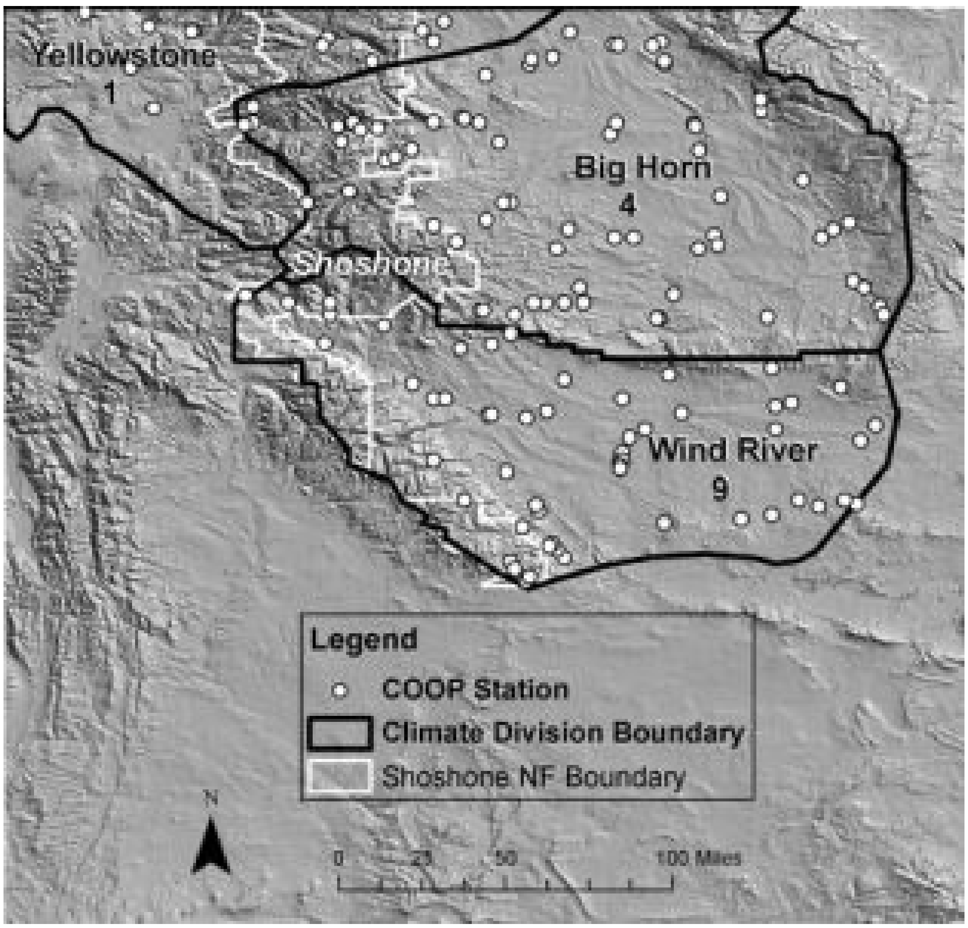



**Figure 3.5:** Three watershed drainages contributing to the Missouri Watershed. Reproduced from Rice et al. (2012).

The climate data we focused on was solely from the Cold Springs SNOTEL site because it was the closest SNOTEL site (19 km northeast of Dinwoody Glacier) to Dinwoody Glacier with a long-term data set. This data set is not perfect, with the most glaring issue being that it lies 500 m below the glacier. In addition, research by Oyler et al. (2015) suggests that systematic errors with instrumentation at many SNOTEL sites across the U.S. Mountain West has led to artificial amplification of warming trends. We have not corrected for those effects here. Air temperature at the Cold Springs SNOTEL site has warmed about three degrees over the period of 1980 to 2010 (Figure 3.6). Average annual temperature for this time period was 2°C with a range of daily temperatures between -33°C and 20°C.

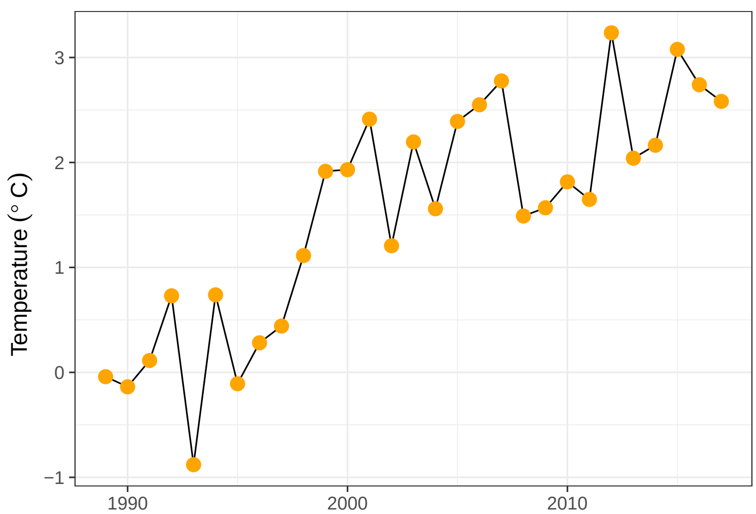



**Figure 3.6:** Temperature recorded from the Cold Springs SNOTEL site (19 km northeast of the glacier, 2935 m a.s.l.) has warmed about 3 degrees in the last 30 years.

Interestingly, the recorded precipitation at Cold Springs had no significant trend up or down (Figure 3.7). Average precipitation from 1990-2017 was 48 mm (NRCS). It should be noted that the numbers we utilized were the raw data with very minimal data points removed because they were outliers. In addition, this data is not corrected for errors associated with precipitation measurements.

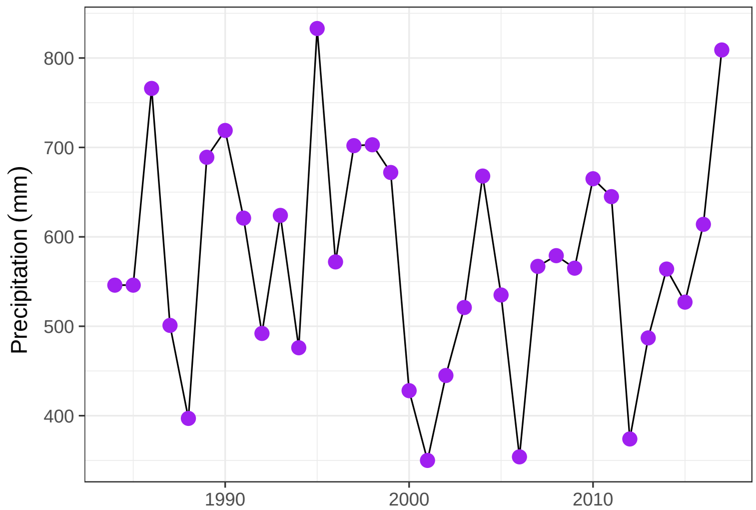



**Figure 3.7:** Precipitation recorded from the Cold Springs SNOTEL site (19 km northeast of the glacier, 2935 m a.s.l.) has no significant trend up or down. It is worth noting that 2017, the year of our field measurements, was a very high snow year.

The annual maximum snow water equivalent (Figure 3.8) was well-correlated with annual precipitation (r=0.61). Like precipitation, it was quite variable from year to year, ranging from 117 mm w.e. in 2001 to 498 mm w.e. in 1999. The mean over the interval 1984 to 2017 was 224 mm w.e.

The day of year when the maximum SWE is reached (Figure 3.9) varied from early March (day 68) to mid-May (day 141) with a mean of April 9th (day 100) over the period 1984 to 2017.

Despite the orographic influence of precipitation in this region, drought is a natural part of the climate's regime (McNeeley and Beeton, 2017). Recently however, this region has experienced more frequent and more severe droughts (Johnson, 2015). From 2000-2016, Wyoming experienced three "extreme to exceptional droughts" (2002, 2006, 2012-2013) (Tribal Water Engineer, 2017). The recorded increases in temperature correlate with less snow accumulation and earlier peak runoff (one to three weeks earlier) for local streams (Bell et al., 2012; Hall et al., 2012). Recent changes in weather and seasonality have impacted the amount and timing of discharge in the Wind River Range area (Cayan et al., 2016; Hall et al., 2015). The percentage of annual streamflow that can be derived directly from snowmelt has decreased by 10-25% since the 1950s (Cayan et al.,

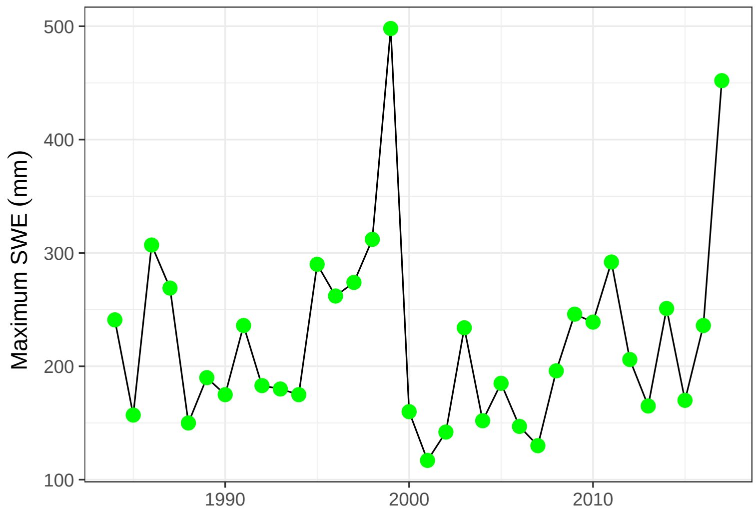



**Figure 3.8:** Annual maximum snow water equivalent recorded from the Cold Springs SNOTEL site (19 km northeast of the glacier, 2935 m a.s.l.) It is worth noting that 2017, the year of our field measurements, was a very high snow year.

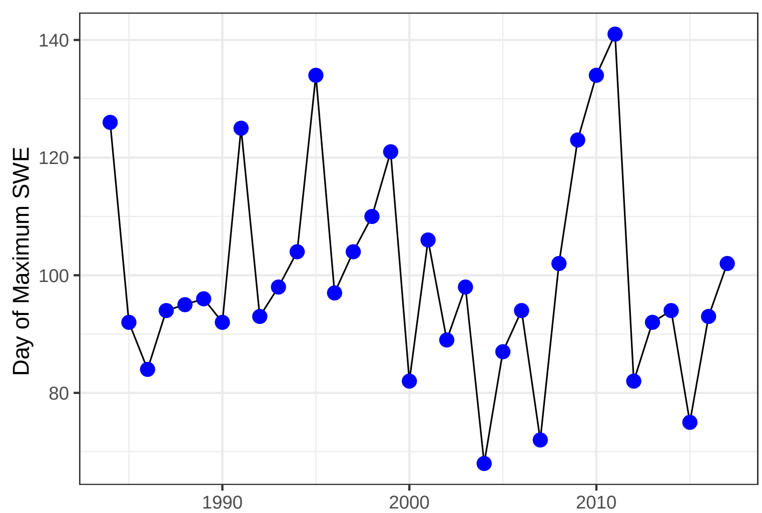



**Figure 3.9:** Day of maximum snow water equivalent recorded from the Cold Springs SNOTEL site (19 km northeast of the glacier, 2935 m a.s.l.).

2016). Decreased inputs of glacial melt water has been linked to higher stream temperatures and altered stream chemistry (Cable et al., 2011; Jost et al., 2012; Naftz et al., 2002).

Variable timing of weather has significantly impacted how agriculture and ranching is done in this region (Cheesbrough et al., 2009). Years with insufficient snow-pack coincide with decreased forage production and water availability. Local farmers and ranchers within the last two decades have had to let land fallow, alter grazing strategies and sell their cattle to minimize herd size during times of drought (McNeeley and Beeton, 2017). This region is now seeing a shift towards drought-resistant plant species that generally have less leaf area per unit than do drought-sensitive species (Chapin III et al., 2017). As a result, ranchers are having to purchase more hay for cows that can't find as much palatable forage (McNeeley and Beeton, 2017). Shifts in plant communities have been recorded on long term scales (30+ years) as well as, on shorter temporal scales following periods of drought (Barnhart et al., 2016; Harpold et al., 2012; McNeeley and Beeton, 2017). Plant community shifts are evident across the Wind River Range with large succession shifts towards invasive exotic grasses. The invasion of exotic grasses has increased fire frequency and decreased resource supply, trophic interactions, and the rates of most ecosystem processes (Hooper and Vitousek, 1997; Loope and Gruell, 1973).

The consequences of decreased water availability and forage production also negatively impact native wildlife such as elk, deer, antelope, and sage grouse in this region. In years with insufficient snowpack, forage may not completely mature resulting in decreased annual production (McNeeley, 2017). The decreased forage availability then impacts ecosystem structure and function, as they are important actors in transfers of energy and materials within ecosystems (Chapin III et al., 2017). In addition, wildlife are culturally important to the tribal members living on the Wind River Indian Reservation.

Fisheries in this region have also seen changes in abundance and distribution of Flathead Chub, burbot, and sauger (Underwood et al., 2016). The decline of fisheries in this region is a result of poorly constructed dams and fish passages, as well as sustained droughts that reduce stream flow.

The combination of these two factors increases stream temperatures and reduces safe habitat for fish to spawn and lay eggs (Wyoming Water Science Center).

Together, these factors create challenges for land managers on long and short term scales. "Stationarity", the idea that ecosystems can be managed the same way as they were in the past, is no longer the paradigm of water management practices in this region (McNeeley and Beeton, 2017). The changes occurring to the water cycle at the Wind River Indian Reservation have been difficult to plan for and respond to as managers year after year have dealt with uncertainty and novelty in weather patterns.

### 3.1.3 Uniqueness of 2017 weather

The year that our research was performed was anomalous (Figure 3.3); 2017 was a cold and wet year across the Wind River Range. It was the seventh coldest winter for Riverton, WY since 1963, with an average temperature 9°C below normal (Tribal Water Engineer, 2017). Precipitation and snow-water-equivalent recorded across the Eastern and Western flanks of the Wind River Range exceeded 200% of the median amount (Figures 3.10, 3.11, 3.12, 3.13). Dubois, WY, the jumping off point for our field research, received 2.07 m of snow, which was 439% higher than its average of 0.37 m (USGS).

In the following section, the hydrologic character of Dinwoody Creek Watershed will be discussed.

## 3.2 Hydrology

Dinwoody Glacier acts as a reservoir containing solid water until hot summer temperatures transform the water's solid phase to liquid. Past research conducted on Dinwoody Glacier focused on glacial melt water contributions to stream flow (Bell et al., 2012; Cable et al., 2011; Cheesbrough et al., 2009; Hall et al., 2015). Forecasting the flow of glacier-fed streams requires an understanding of accumulation and ablation processes and of glacier hydrology (Paterson, 1981). The annual accumulation of snow on Dinwoody Glacier has been estimated via satellite imagery

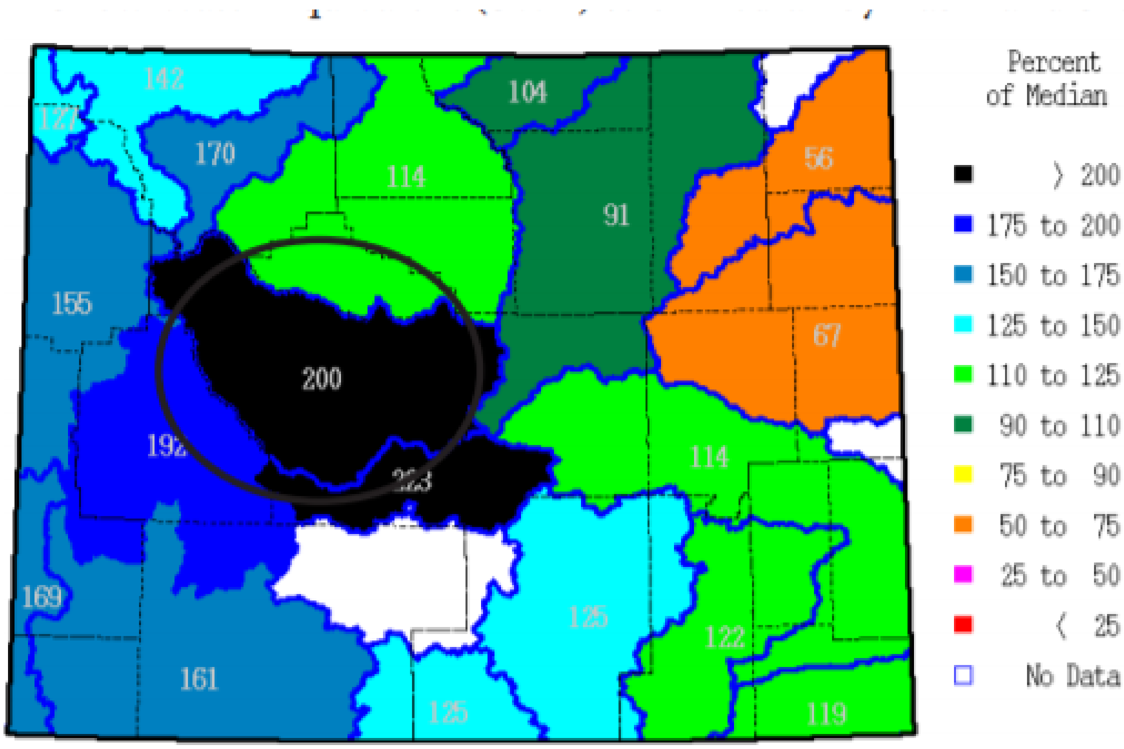



**Figure 3.10:** Snow Water Equivalent (SWE) Percent of Median by Basin and SNOTEL Site (Wyoming) in February of 2017 (Tribal Water Engineer, 2017)). The reference period for the median is 1981-2010.

by various researchers (DeVisser and Fountain, 2015; Hall et al., 2012; Thompson et al., 2011). However, field-based measurements are difficult to obtain due to the area's rugged and remote setting. For satellite derived snow-cover extents, the annual accumulation is determined from snow-cover maps collected at the transition of winter to spring, late March or early-April. Similarly, ablation is calculated with snow-cover maps collected before snow starts to accumulate, late July or early August. The combination of high resolution satellite imagery, local stream gauge data, and meteorological data can portray an accurate depiction of glacial mass balance (Paterson, 1981). In addition, field based measurements are critical for calibrating remotely sensed data. Glacial surface melt water is the most important source of water in temperate glaciers (DeVisser and Fountain, 2015). During the summer, melt water meanders across the glacier's surface forming braided networks of flowing water that resemble river systems.

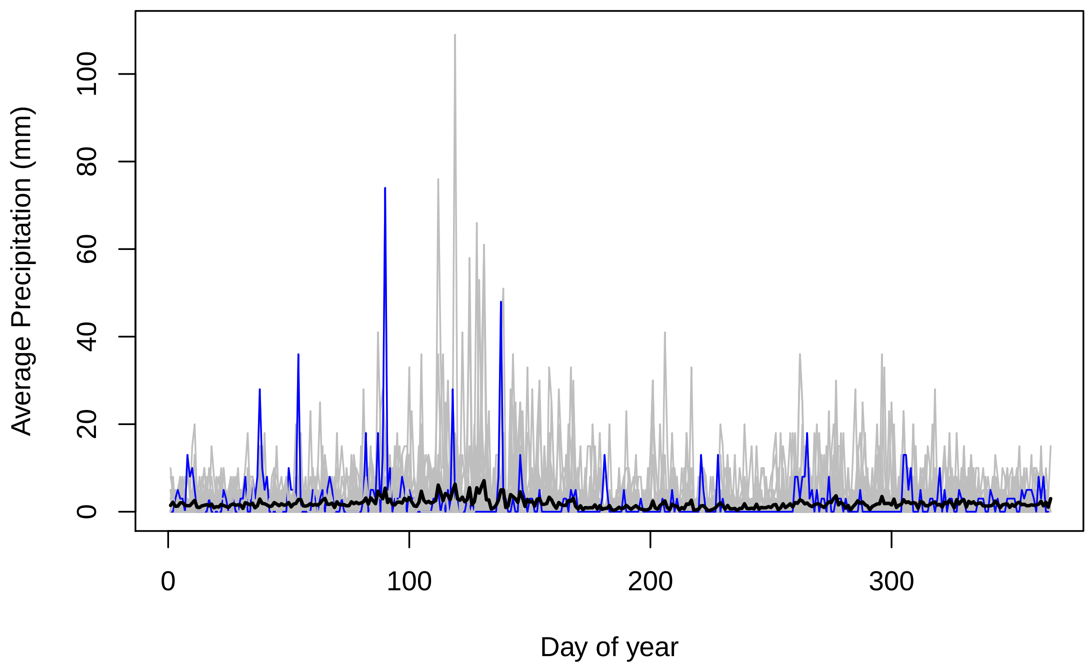



**Figure 3.11:** Daily precipitation at the Cold Springs SNOTEL site for 1990-2017 (gray lines, with 2017 highlighted in blue). The mean for each day of the year is plotted in black.

Surface melt-water's fate is the glacier's terminus but the path there often includes subsurface or even basal flow via shallow or deep crevasses (Paterson, 1981). Eventually, the meltwater reaches the mouth of Dinwoody Glacier and enters Dinwoody Creek.

In the last 30 years, peak accumulation of perennial snow has decreased and ablation (melt) is occurring earlier in the summer as a result of increased temperatures, reducing stream flow when water demands downstream are highest (Cheesbrough et al., 2009; Hall et al., 2012). The percentage of annual streamflow that can be derived directly from snowmelt has decreased by 10-25% since the 1950s (Cayan et al., 2016). Altered stream flow due to decreased inputs of glacial melt water has been linked to higher stream temperatures and altered stream chemistry (Cable et al., 2011; Jost et al., 2012; Naftz et al., 2002). In a study conducted from 1994-2001, glacial meltwater contributed 4.4–14.1% of the flow in Dinwoody Creek (DeVisser and Fountain, 2015). River discharge at the USGS gauge Dinwoody Creek Above Lakes Near Burris, WY was high in

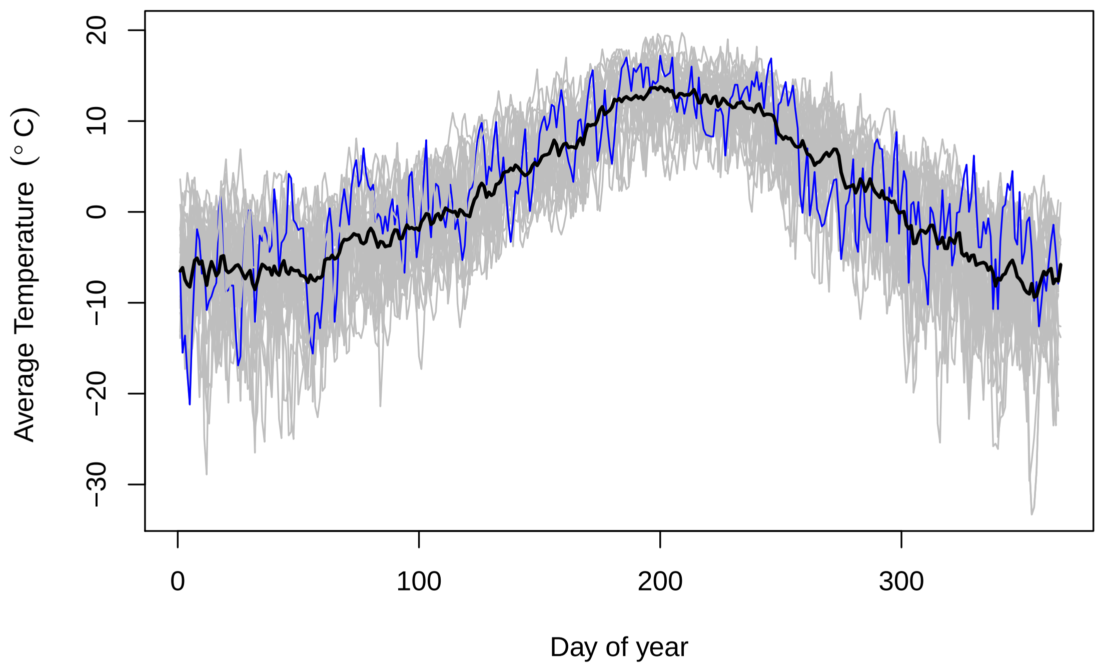



**Figure 3.12:** Daily temperature at the Cold Springs SNOTEL site for 1990-2017 (gray lines, with 2017 highlighted in blue). The annual mean was 2°C with minimum and maximum temperatures of -33°C and 20°C. The mean for each day of the year is plotted in black.

2017 (Figures 3.14 and 3.15. Annual discharge in 2017 (0.19 $km^3$) was more than three standard deviations above the mean annual discharge (0.13 $km^3$).

## 3.3 Recent glacier changes

Increased temperatures and decreased snowfall amounts have led to glacier recession in the Wind River Range (Johnson and Ohara, 2018; Plummer et al., 2006; Pochop et al., 1990), (Figure 3.1). Although temperature and precipitation are the predominant factors controlling glacier mass balance, glacier change can also be affected by natural phenomena and local factors such as glacier exposure, snow accumulation conditions, and glacier morphology (Bell et al., 2012; Cayan et al., 2016; Thompson et al., 2011).

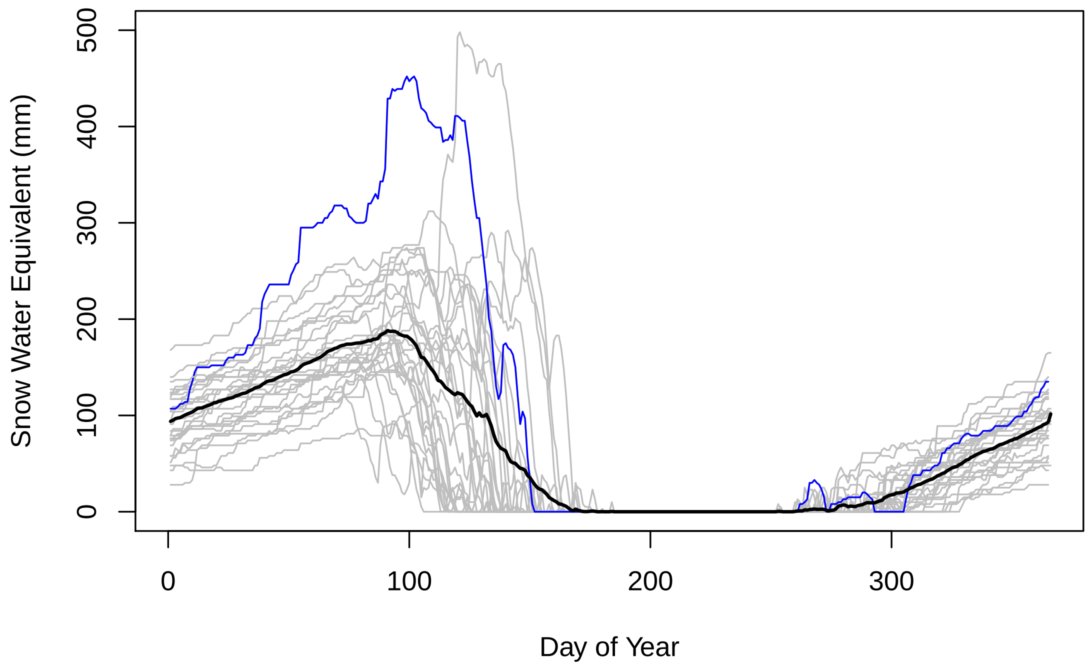



**Figure 3.13:** Snow water equivalent for the Cold Springs SNOTEL site for 1990-2017 (gray lines, with 2017 highlighted in blue). The mean for each day of the year is plotted in black.

Changes in glacier area are commonly dominated by changes in terminus position (Cheesbrough, 2007). Research comparing the Little Ice Age glacial extents to the modern terminal moraines estimate glacier area decreases since the Little Ice Age (∼ 1900 to 2006) of about 44%, with half of the mass loss occurring in the past 35 years (DeVisser and Fountain, 2015; Thompson et al., 2011). It should be noted here that ice core records collected by Love and Thompson (1987) showed evidence that many of the glaciers of the Wind River Range had re-advanced to their Little Ice Age (1850) position by 1930, which may be supported by a documented cool period between 1915 and 1930 (Naftz et al., 2002). The significant recent changes near the terminus of the glaciers are due to warmer temperatures at lower elevations as well as lower albedo (the fraction of incoming solar radiation that is reflected by a surface). This is due in part to the litter of rocks and dirt on the snow and ice near the terminus lowering the albedo and increasing the absorption of solar radiation (Cheesbrough et al., 2009). A 2015 investigation of Dinwoody Glacier using georeferenced

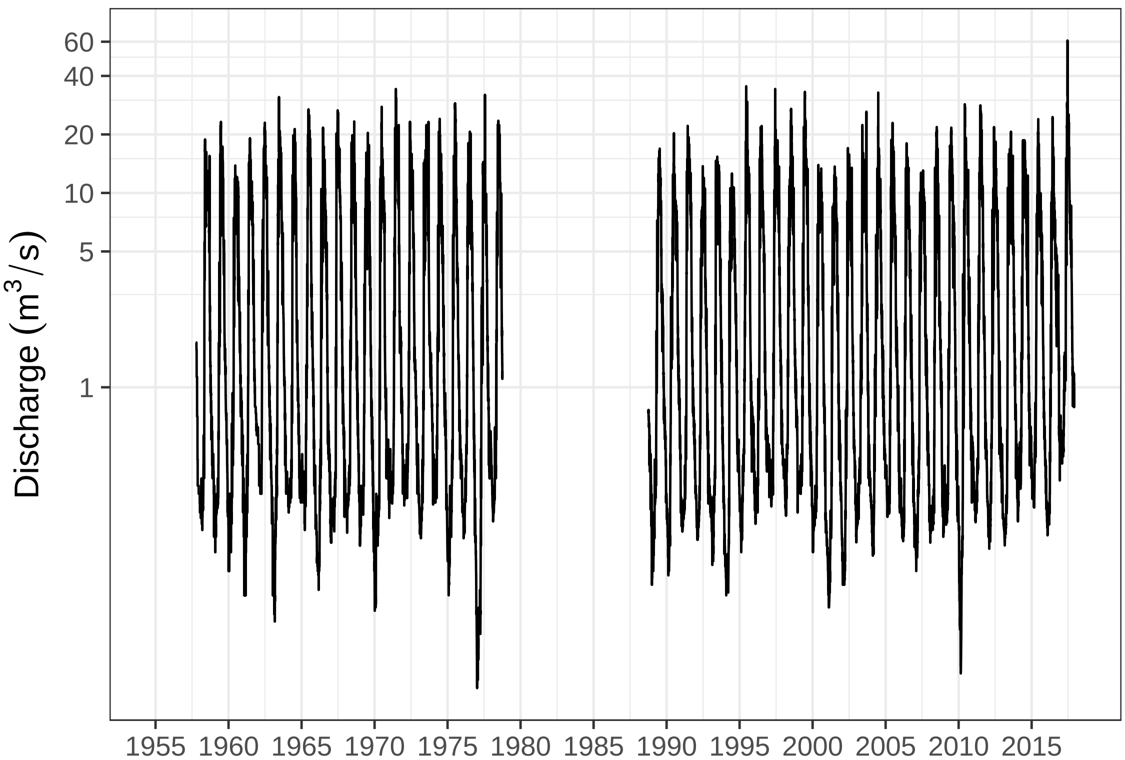



**Figure 3.14:** Stream discharge from the USGS gauge 35km north-east of the glacier (Dinwoody Creek Above Lakes Near Burris, WY, 1981 m a.s.l) captures the hydrologic inputs from a drainage area of about 228 $km^2$.

maps at 1m resolution from 1900-2006 found a surface area loss of 2.38 $km^2$, an overall reduction of surface area of 35% (**?**). The recession of Dinwoody Glacier is correlated (among other factors), to decreased snow cover; on average, snow cover in this region is melting 16$\pm$10 days earlier in the 2000s compared to the time period of 1972-1999 (Hall et al., 2015). The relationship between percentage of basin-wide snow cover and timing of snowmelt are primary drivers of stream discharge. Thus, understanding the relationship between snow cover and stream flow is important for water managers in this region (Hall et al., 2012).

SNOTEL sites distributed across the Wind River Range have recorded temperature increases over the last 30 years (Figure 3.6); however, no meteorological sites exist at the elevation at which the glaciers reside. As a result, temperature proxies are utilized or temperature is estimated via correlation with lower elevation sites. A study done on the Upper Fremont Glacier (7 km north

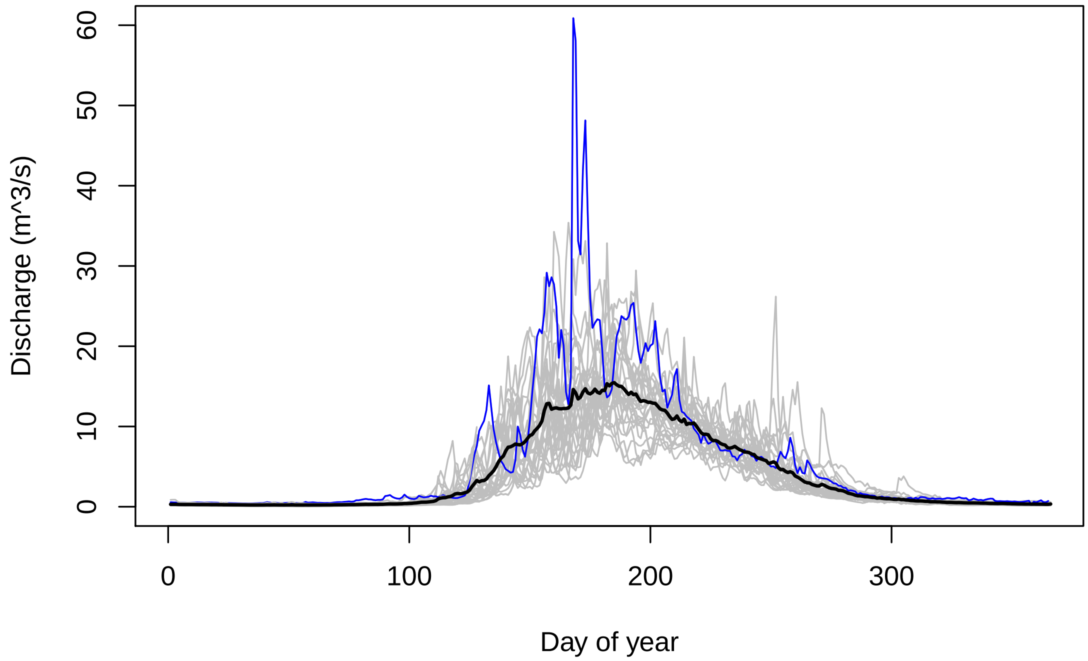



**Figure 3.15:** Dinwoody Creek discharge climatology for 1990-2017 (gray lines, with 2017 highlighted in blue). The mean for each day of the year is plotted in black.

of Dinwoody Glacier) by Naftz et al. (2002) indicated a 3.5°C increase in temperature since the 1960's based on a strong correlation between $\delta^{18}O$ levels in snowfall and ice cores. Further more, Thompson et al. (2011) examined temperature increases using local SNOTEL sites and found that for eight sites around the Wind River Range, average temperature increased in July and August, 3°C± from 1986 to 2008.

The following section will discuss our research methodologies and findings.

# Chapter 4

# Field methods, data, and results

Fieldwork was an essential part of this research as it allowed us to gather in situ ground measurements that ensured a more precise calibration of remotely sensed data and remote meteorological stations. Our fieldwork provided measurements (quantitative) and descriptions (qualitative) of variables that will be used in future work to validate physically-based numerical models of hydrologic processes. Information gathered in the field related to rates of process operation, as well as, the size and status of Dinwoody Glacier ($\sim$3 km$^2$, 3300-4000 m above sea level).

We collected data on Dinwoody Glacier 2017/08/05-08 including glacier photogrammetry, ablation rates, snow-line elevations, localized weather observations and stream flow. The data gathered in the field helped to elucidate our main research questions:

1. What is the predominant climatic trend of this region?
2. What is the present mass balance of Dinwoody Glacier?
3. How much does glacial melt water runoff contribute to stream power and stream quality in the Dinwoody Creek watershed?

To answer these questions, we sought to measure the mass balance of Dinwoody Glacier. The concept behind measuring a glacier's mass balance is quite simple, it is the sum of mass losses and gains over a year (Ahlmann, 2018); however, in practice, accurately gathering the data is often complicated by various physical barriers. In general, the primary sources of mass input are snow accumulation from snowfall, or redistributed snow from wind and avalanches (Fountain et al., 1997). Other sources of mass input include the freezing of rain within the snow, or condensation of water vapor. Processes of ablation (all forms of mass loss) include melt and subsequent runoff from the surface and interior of the glacier, evaporation of liquid water from the glacier surface, and sublimation.

## 4.1 Glacier-wide photogrammetry

We constructed a three-dimensional model of the surface of Dinwoody Glacier to track changes in glacier volume over time and to map the snow line from ground-based photos. On 2017/08/07-2017/08/08 we took nearly 7000 ground-based photographs from the locations shown in Figure 3.1. We used a Nikon D80 DSLR camera with a 50 mm fixed lens (35 mm equivalent focal length = 75 mm). We used natural features for ground control points: unique, prominent, and large boulders that protruded from or were adjacent to Dinwoody Glacier. A total of 20 ground control points were recorded. Locations of the ground control points were measured with a geodetic-quality Trimble Geo 7x receiver. Post processing of the GPS points gave an estimated average horizontal precision of 0.1 m.

We used AgiSoft Photoscan software to process an ad hoc selection of the nearly 7000 images. We manually identified each ground control point on at least 3 photos and added the GPS location information. Then Photoscan generated a dense 3D point cloud through a process of stereomatching which detects homologous pixel pairs on two pictures using feature-based or area-based algorithms (Dissard and Jamet, 1995). Photoscan also solves for camera positions and optical characteristics of the lens. From the 3D point cloud, Photoscan generated an orthophoto and a DEM.

Given the number of ground control points and their spatial distribution, we are confident that the distribution of collected points ensures that data coverage accurately reflects the topography of Dinwoody Glacier, though error inevitably still exists.

**Results of glacier-wide photogrammetry**

Photoscan generated a dense 3D point cloud with more than 250 million points in Dinwoody Glacier's basin (Figure 4.1). The orthophoto generated from the photogrammetry showed a clear contrast between areas of the glacier that were covered by snow and areas that were bare glacier ice. Using ArcGIS, we digitized those areas and found that snow covered 71% of the glacier on 2017/8/8. We expect that the percentage of snow cover at the end of summer 2017 was significantly less than 71%, though likely higher than other years due to the high snowfall in the 2016/2017 winter.

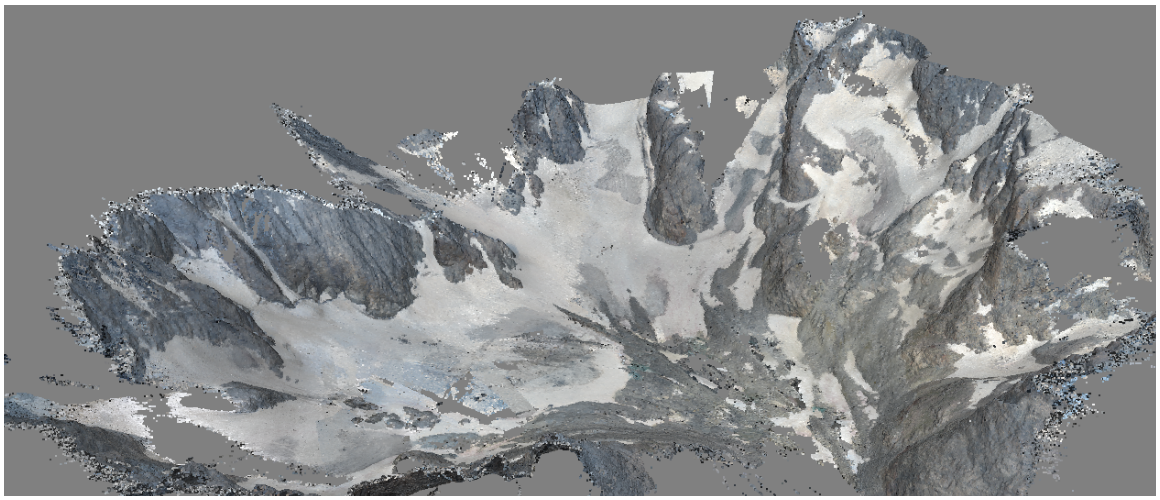


**Figure 4.1:** Perspective view of the photogrammetric reconstruction of the Dinwoody Glacier surface. The view is from the east towards Gannett Peak (just out of view in the upper right). Generated from photos taken on 2017/8/7-8.

Future work will compare the elevations derived from the photogrammetry with older digital elevation models to calculate elevation change and mass loss over time.

## 4.2 Summer snow conditions

Utilizing a 270 cm avalanche probe, we measured the depth of seasonal snow that remained on 2017/8/7-8 at roughly 20 m intervals from west to east across Dinwoody Glacier at a consistent elevation (see Figure 4.2). The chosen elevation and contoured path was located in the ablation zone in normal snow years (such as 2015, Figure 3.1), so we expected the probe to reach the solid glacial surface in most locations. We also did a longitudinal profile of snow depth going up-glacier from Ablation 2. The purpose of gathering the depth of snow measurements for our research was to glean information on the previous winter's snow accumulation.

Three depths were recorded at each site, horizontally spaced apart by approximately 1 m. The three depths from each site were later averaged to make the map in Figure 4.2. If the first three

**Table 4.1:** Density measurements from the snow pit at Ablation 2, 2017/8/7. Sample depth is measured from the surface.

| Sample depth (cm) | Density ($kg/m^3$) |
|---|---|
| 0-6 | 587 |
| 10-16 | 653 |
| 20-26 | 623 |
| 30-36 | 614 |
| 40-46 | 571 |

probed depths differed by more than 50 cm, we probed until three consistent depths were measured. Probe depths may have been inconsistent because of ice layers (lateral or vertical), buried crevasses, or a non-uniform snowpack. When calculating the average depth for each location, we discounted the inconsistent depths. The methodology utilized for snow depth measurements in this study maximized our safety while striving for a sample that was representative of the current snowpack. That is to say, our snow depth measurements have error associated with them due to the quantity of measurements taken per location. Research performed by Fassnacht et al. (2018) suggests that variability in snowpack cannot be calculated to a modeled mean within 5% error by utilizing three probed depths per location.

In addition to probed snow depths, we dug a 50 cm deep snow pit (see Figure 4.3) near the Ablation 2 plot (approximately 3567 m above sea level). We collected snow samples using a plastic tube with rectangular cross-section, thickness of 6 cm, and volume of 438 $cm^3$. The weight of each sample was measured on an Etekcity model HY2000 scale. We then calculated snow density.

### Results of summer snow conditions

For the lower glacier in early August 2017, snow depth ranged from 0 to more than 270 cm. The average snow depth recorded was 220 cm. The snow depths that were deeper than 270 cm were scaled by averaging the mean snow depth and the minimum snow depth at the specific measurement location. Areas that were shaded likely experienced less melt since the end of winter than adjacent sunny areas. Some of the probed locations near the cirque headwall likely received extra

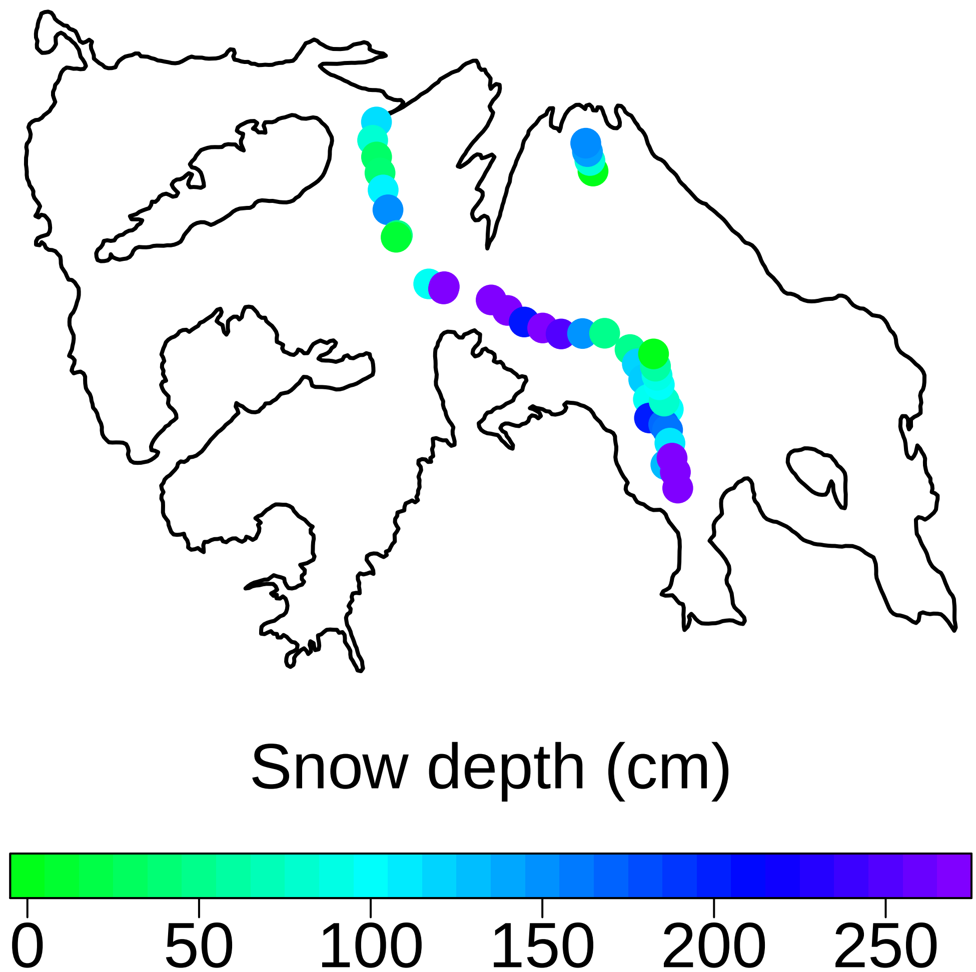



**Figure 4.2:** Averaged snow depth per location on the glacier (2017/8/7-8) varied from 0 to over 3 m. We found deep snow on north facing slopes beneath avalanche chutes.


**Figure 4.3:** The snow pit we dug at the Ablation 2 site on 2017/8/7 contained layers of different densities.

accumulation due to wind-redistribution of snow or avalanches. There were few surface remnants of avalanches on 2017/8/07, though the slope of the upper-reaches of the glacier are above 40° and therefore, able to generate them.

Calculated densities in the snow pit we dug varied from 571 kg/m$^3$ to 653 kg/m$^3$ (Table 4.1) near the Ablation 2 plot (approximately 3567 m above sea level).

## 4.3 Estimated glacier snow accumulation via snowline tracking

The equilibrium-line altitude of a glacier is the average elevation at which accumulation of snow is exactly balanced by ablation (Benn and Lehmkuhl, 2000). The instantaneous position of the boundary between snow-covered glacier above and bare glacier ice below is known as the snow line. Throughout the accumulation and melt seasons, the position of the snow line is transient. These transient snow-lines are used to estimate glacier mass balance (Meier and Post, 1962) and several studies have shown that the snow line altitude at the end of the hydrological year is a good indicator of the equilibrium-line altitude for mid-latitude glaciers (Rabatel et al., 2012).

Snow line altitude can be reconstructed from satellite imagery or ground based photos, among other remote sensing techniques. Photogrammetry is an ideal technique for the collection of topographic data in glaciated areas because it allows temporally comparable measurements from year to year (Fox and Nuttall, 1997). It also gives a fine-scale spatial and temporal representation of the changing glacier surface.

We utilized ground-based photogrammetric techniques to acquire the elevation of the snow line in the fall of 2016 (Figure 4.4). The photos were taken each hour during daylight by a Wingscapes TimelapseCam Pro camera. Unfortunately, the time lapse camera tilted upward after a heavy snowfall in late September of 2016. The hour of day that had the least amount of shadows and most spatially consistent lighting across Dinwoody Glacier was the 10 a.m. shot, so we used these for our analysis. Using ArcGIS software (ESRI, 2018), we registered to a common base image the 46 photos taken at 10 a.m. starting with 2016/7/26 and ending with 2016/9/17. Then for each image we digitized the outline of snow on the glacier's surface. The digitization resulted in a raster

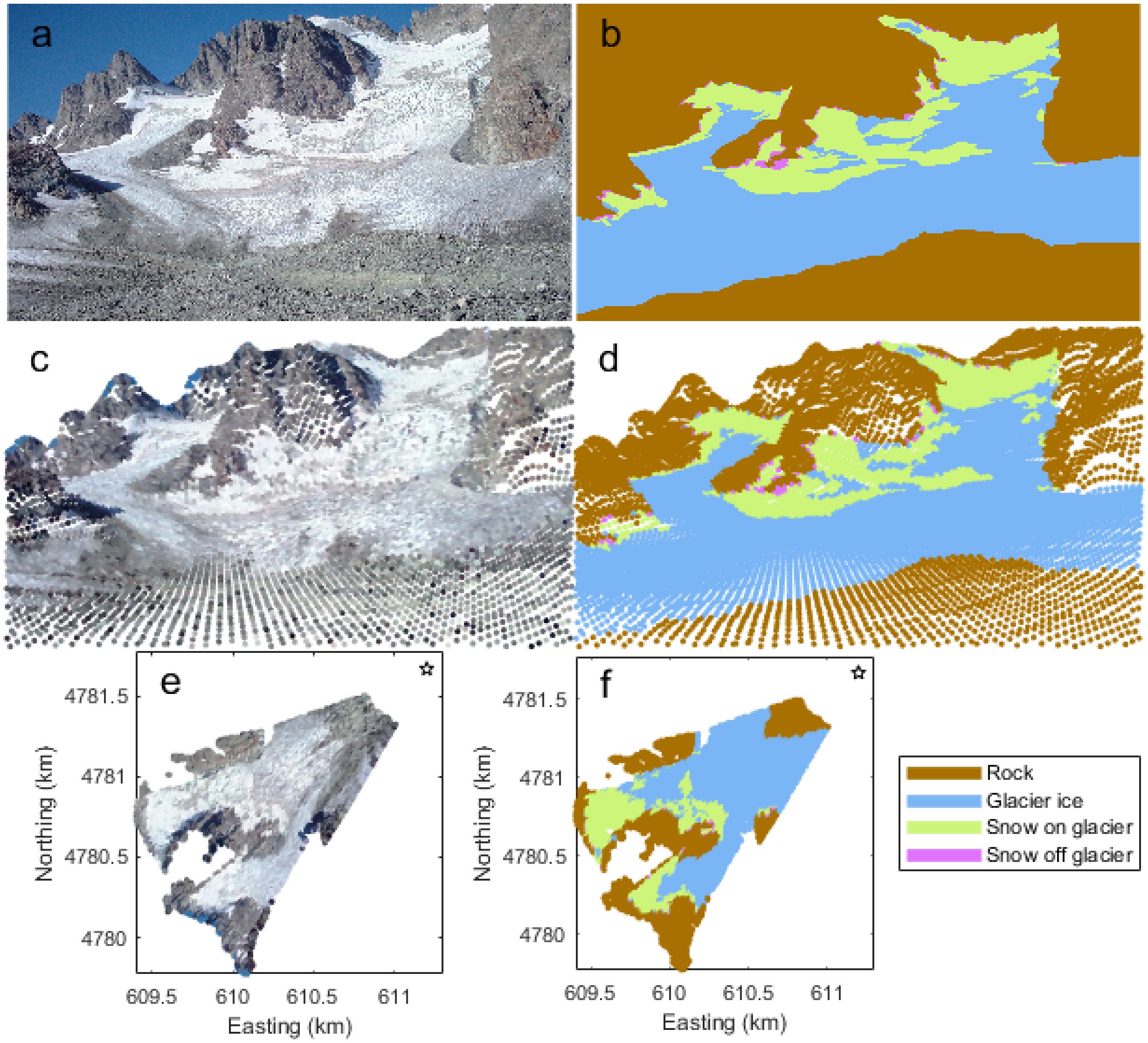



**Figure 4.4:** Illustration of the method to produce snowline maps. (a) A photo from 2016/8/29 10:00 a.m. (b) The manual classification of the photo. (c) RGB values extracted from the photo for the center of each DEM grid cell. (d) Classification from the center of each DEM grid cell. (e) The photo, reprojected as an orthophoto. The star shows the camera location. (f) The classified image, projected as an ortho map.

for each day with 4 classes: rock and sky, glacier ice, snow on glacier, and snow off glacier. The same person did all the outlines for the snowlines to keep the analysis as consistent as possible.

We projected each of the photos and rasters onto the DEM using software called PRACTISE, written with Matlab (Härer et al., 2016). This allowed us to reproject the photos and classifications onto a vertical map view where we could accurately calculate areas belonging to each class on each day (Figure 4.4).

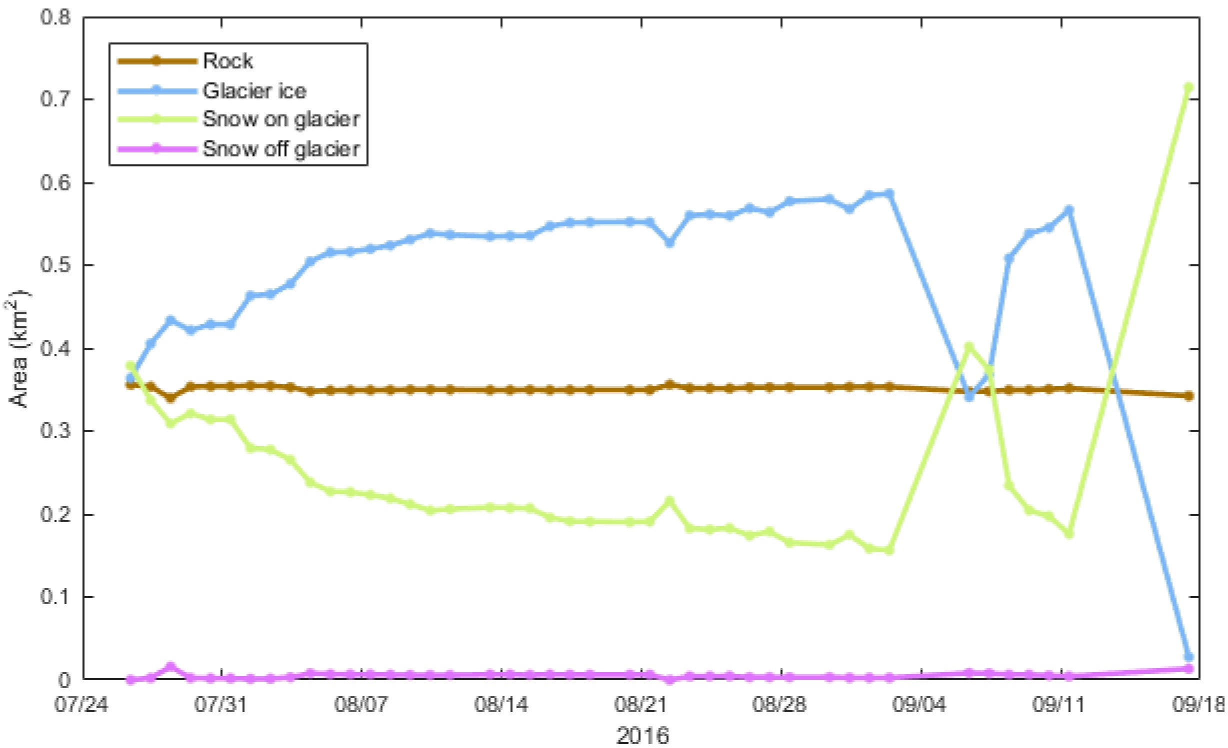



**Figure 4.5:** The area covered by different surface types in late summer of 2016, within view of the camera.

### Results of snowline tracking

The total area of the glacier within view of the camera was 0.74 km$^2$ or about one third of the entire glacier surface. The area of snow on the glacier (within view) gradually decreased from 0.38 km$^2$ on 2016/7/26 to 0.16 km$^2$ on 2016/9/2 (Figure 4.5). A storm dropped snow across the glacier between 2016/9/3 and 2016/9/6, most of which melted again by 2016/9/11. A storm which started on 2016/9/12 covered the entire area visible to the camera with snow and between the 18th and

19th the camera tilted to the sky. The minimum accumulation area ratio (AAR) for the two lobes of the Dinwoody Glacier that we can see in the time lapse imagery was 21% on 2016/9/2. Since typical equilibrium values of AAR are between 40 and 80% (Dyurgerov et al., 2009), it is likely that Dinwoody Glacier in 2016 had a negative mass balance.

## 4.4 Ablation measurements

Ablation is the removal of snow through melt or evaporation; it is a function of temperature, amount of radiation inputs, slope, shading and surface roughness (Fountain et al., 1997; Slaymaker, 1991). We measured ablation at three sites on Dinwoody Glacier (Ablation 1, 2, and 3, Figure 3.1). Ablation 1 was the lowest elevation site at 3509 m and a slope angle of 7.5°. Ablation 2 was our mid-elevation site at 3567 m and slope angle of 11°. Ablation 3 was the highest elevation we could access safely on the east lobe of the glacier at 3606 m with a slope of 20°. At the time of installation, the weather and time of day was recorded. Each site maintains a north facing aspect; Ablation 1 faces 338°, Ablation 2 is due-north, Ablation 3 faces 350°. Variable shading at each ablation site contributed to the differences in incoming solar radiation at the individual sites and therefore overall melt.

At each site we installed three ablation stakes. For each stake, we drilled a hole about 0.3 m deep in the ice or snow and placed a stake made of white PVC pipe in the hole. The initial height of the PVC pipe was recorded as it protruded from the ice. On three consecutive days (2017/08/06-08), the height of each stake was measured on the upslope and downslope sides. Ablation 1 was installed on the evening of the 2017/08/05 and has one more data point than the rest of the sites.

#### Results of ablation experiment

The site Ablation 1 experienced 7.2 cm of melt over the period 2017/8/5-8 (Table 4.2 and Figure 4.6). From 2017/8/6-8 Ablation 2 lost 3.7 cm, and Ablation 3 lost 4.4 cm. At Ablation 1, average ablation for the interval 2017/8/6-8 was 2.23 cm/day. At Ablation 2 and 3, average ablation was 2.05 cm/day, and 2.46 cm/day respectively.

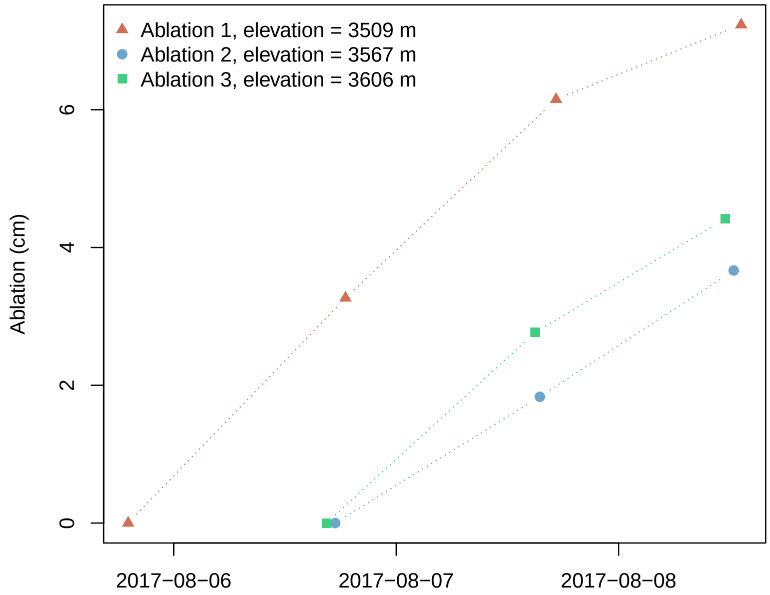



**Figure 4.6:** Ablation rates of three ablation experiment sites 2017/08/06-2017/08/08.

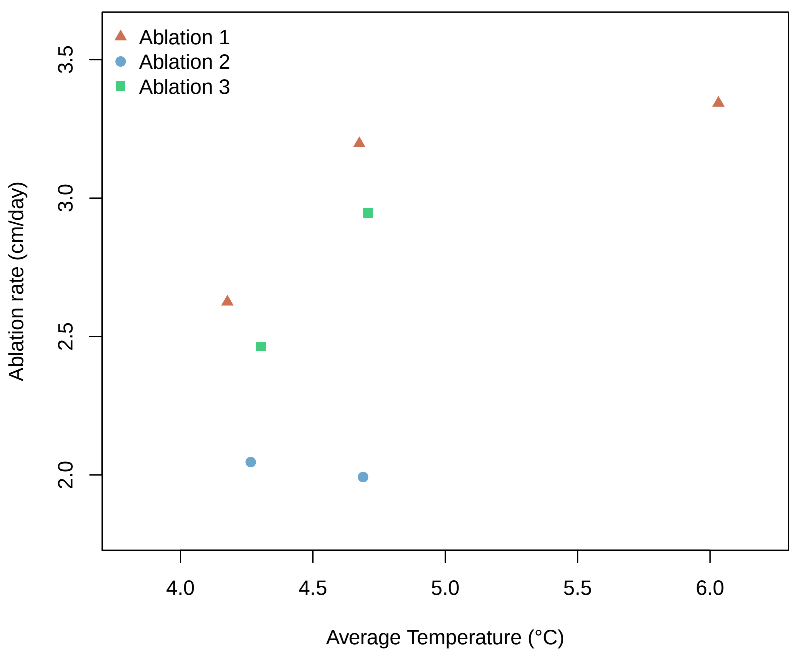



**Figure 4.7:** Daily ablation rates of three ablation experiment sites 2017/08/06-08 plotted against average temperature measured at Ablation 1. Ablation rates correlate with temperature, however, shading, surface type (ice or snow), and surface roughness influence melt too.

Table 4.2: Ablation measurements for the period 2017/8/5-8.

| Site | Ablation 1 | Ablation 2 | Ablation 3 |
|---|---|---|---|
| Elevation (m) | 3509 | 3567 | 3606 |
| Slope (°) | 7.5 | 11 | 20 |
| Aspect (°) | 338 | 0 | 350 |
| Surface cover | Ice | Snow | Snow |
| Dates | 2017/8/5-8 (6-8) | 2017/8/6-8 | 2017/8/6-8 |
| Ablation (cm) | 7.2 (4.0) | 3.7 | 4.4 |
| Ablation rate (cm/day) | 2.63 (2.23) | 2.05 | 2.46 |

Our ablation experiment found the highest ablation rate correlated with the highest temperatures (Figure 4.7). Differing rates of ablation can be attributed to temperature, surface type (snow or ice), surface roughness, shading, and the elevation gradient.

**Results of glacier-wide runoff calculation**

By extrapolating from the ablation rate at Ablation 1 (2.63 cm/day 2017/8/5-8/8) to the whole 2.3 $km^2$ glacier area, we estimated the total ablation of Dinwoody Glacier to be 0.63 $m^3/s$ for early August 2017. If we use the mean of all three ablation sites (2.25 cm/day 2017/8/6-8/8) the total ablation would be 0.54 $m^3/s$. For our research, we assumed mean ablation for all three ablation sites best represents glacier-wide runoff totals.

## 4.5 Ablation plot photogrammetry

In addition to glacier-wide photogrammetry, we collected photogrammetric images of each ablation site with a Garmin Virb camera at the beginning and end of the ablation study. The camera was mounted on an avalanche probe and held approximately three meters above the glacier surface. The camera took photos at three second intervals as the researcher walked a square perimeter of the site at a distance of three to four meters from the triangle of ablation stakes. The ablation stakes were used as reference points to establish the scale of the photos and to register the before and after images together. We measured the distance between each stake with a standard tape measure at the time of the photos to establish scale in the photogrammetric model. The photos were processed using AgiSoft Photoscan (see method in section 4.1), which generated a 3D point cloud, an orthophoto, and a DEM for each measurement time.

We also collected plot scale (approximately 5 $m^2$) photos of the ablation sites using 8 to 18 images taken with a Nikon D80 DSLR camera with a zoom lens set to 18 mm (35 mm equivalent focal length = 27 mm). Images were processed in Agisoft Photoscan with a method similar to the whole-glacier photogrammetry described above.

With the before and after DEMs, we calculated the difference (ablation) in the vicinity of the ablation stakes for comparison with the stake measurements.

**Results of ablation plot photogrammetry**

For Ablation 1, the surface melt for the 65.3 hour period from 2017/8/5 7:40 pm to 8/8 1:00 pm can be estimated by differencing the two DEMs. The height difference in the vicinity of the ablation stakes (see rectangle in Figure 4.8e) was 7.6±1.5 cm (mean ± standard deviation). The average ablation rate calculated with this method was 2.8 cm/day.

At Ablation 2, the height difference was 3.7±3.8 cm (mean±standard deviation). Time difference 1.8 days (2017/8/6 5:40 pm to 2017/8/8 12:30 pm). Average ablation rate was 2.1 cm/day.

At Ablation 3, the height difference was 2.5±2.8 cm (mean±standard deviation). Time difference 1.8 days (2017/8/6 4:40 pm to 2017/8/8 11:40 am). Average ablation rate was 1.0 cm/day.

The photogrammetry-based ablation rates (for the area within the rectangles shown in the figures) compare favorably with the ablation rates calculated from the stake measurements themselves. The photogrammetry shows a richer picture of ablation rather than just a set of 3 point measurements. Future attempts at using this method should try to establish a more-stable coordinate reference system. Our PVC system was not as stable as we would have liked and this introduced errors in the photogrammetry-based calculation that are not as apparent in the stake measurements.

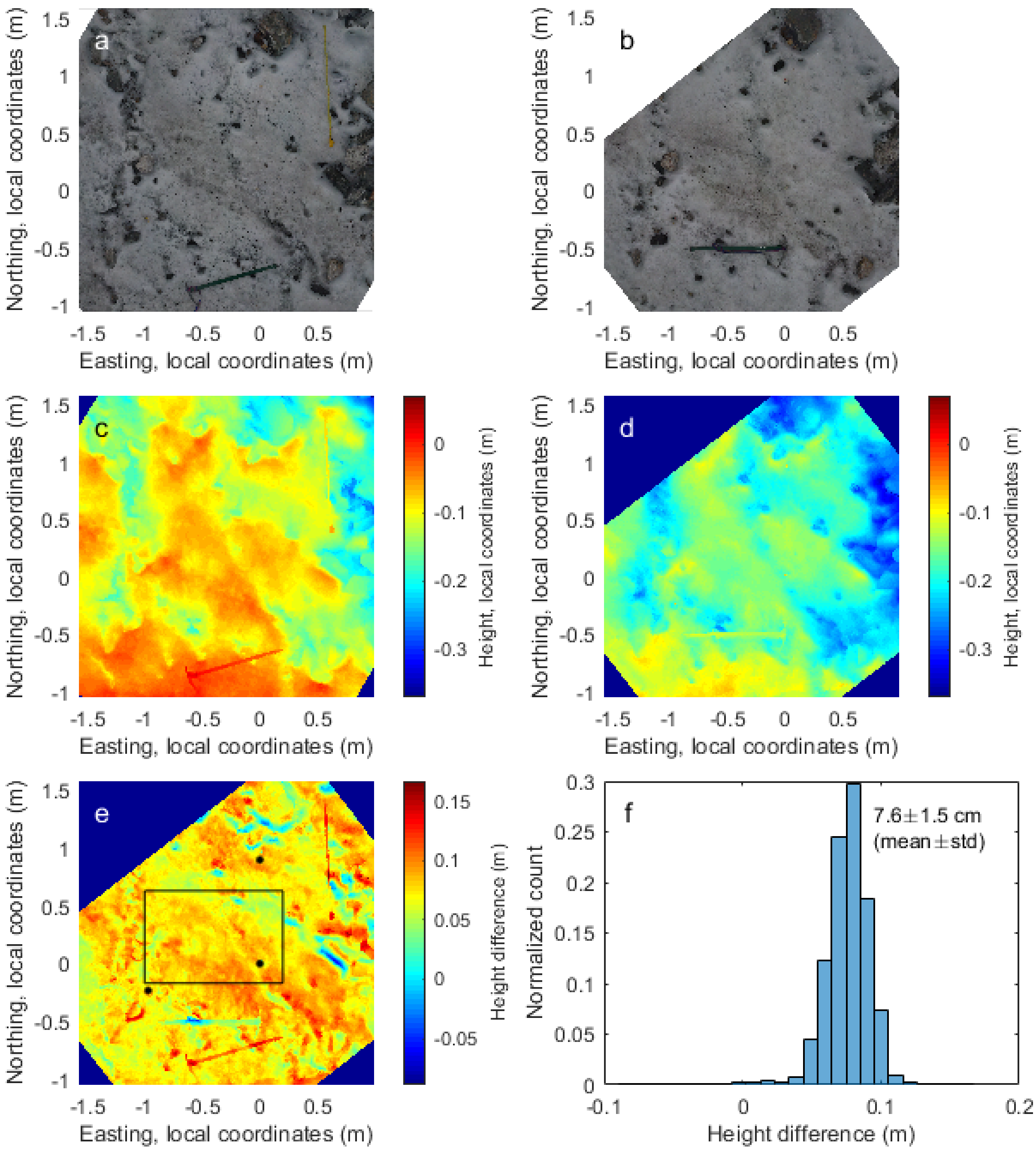



**Figure 4.8:** Photogrammetry of Ablation 1. (a) Orthophoto from 2017/8/5 7:40 pm. (b) Orthophoto from 2017/8/8 1:00 pm. (c) DEM from 2017/8/5. (d) DEM from 2017/8/8. (e) DEM difference showing the area used for statistics (black rectangle) and 3 ablation stake locations (black dots). (f) Histogram of the height difference caused by melt over the period 2017/8/5-8.

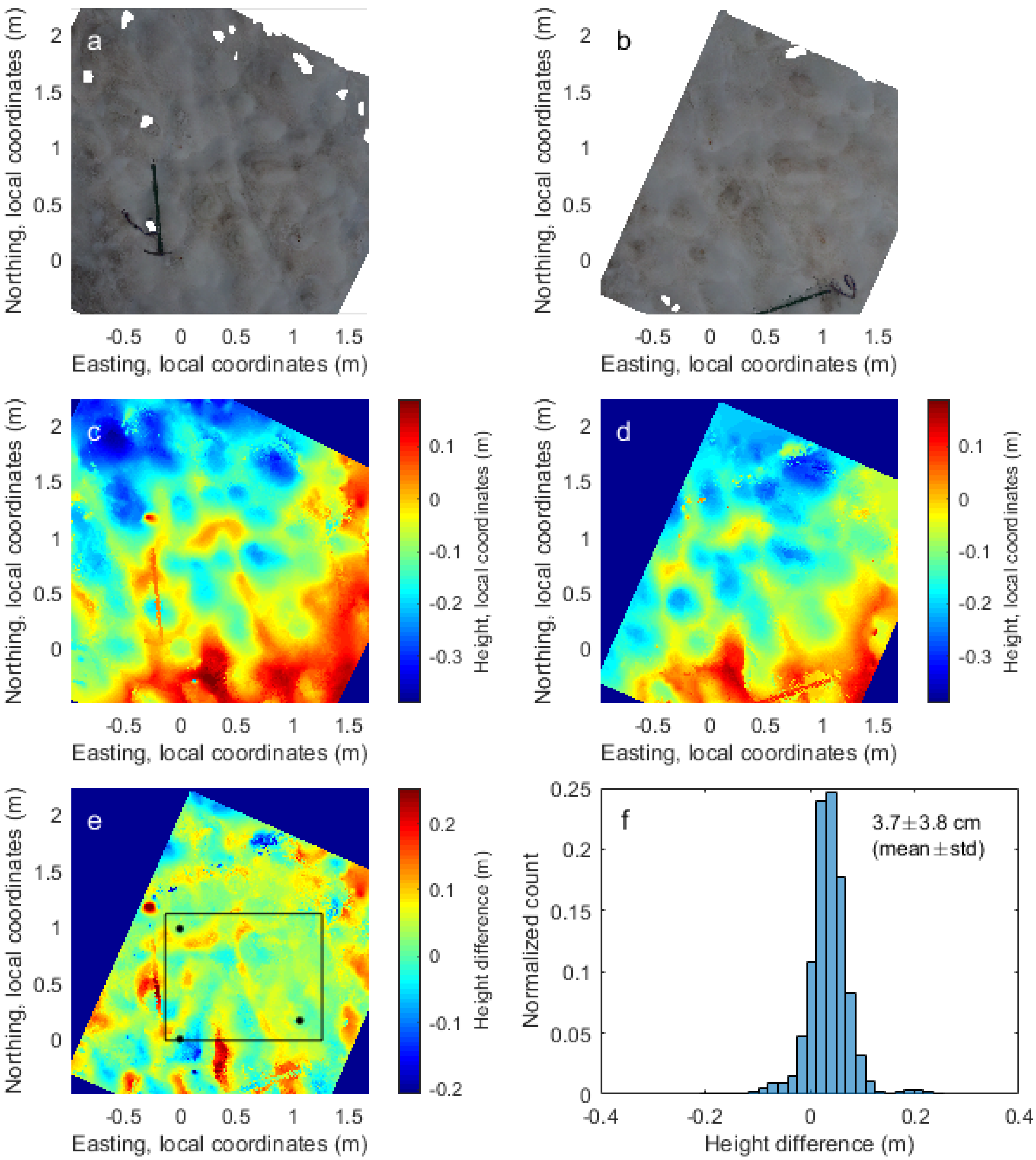



**Figure 4.9:** Photogrammetry of Ablation 2. (a) Orthophoto from 2017/8/6 5:40 pm. (b) Orthophoto from 2017/8/8 12:30 pm. (c) DEM from 2017/8/6. (d) DEM from 2017/8/8. (e) DEM difference showing the area used for statistics (black rectangle) and 3 ablation stake locations (black dots). (f) Histogram of the height difference caused by melt over the period 2017/8/6-8.

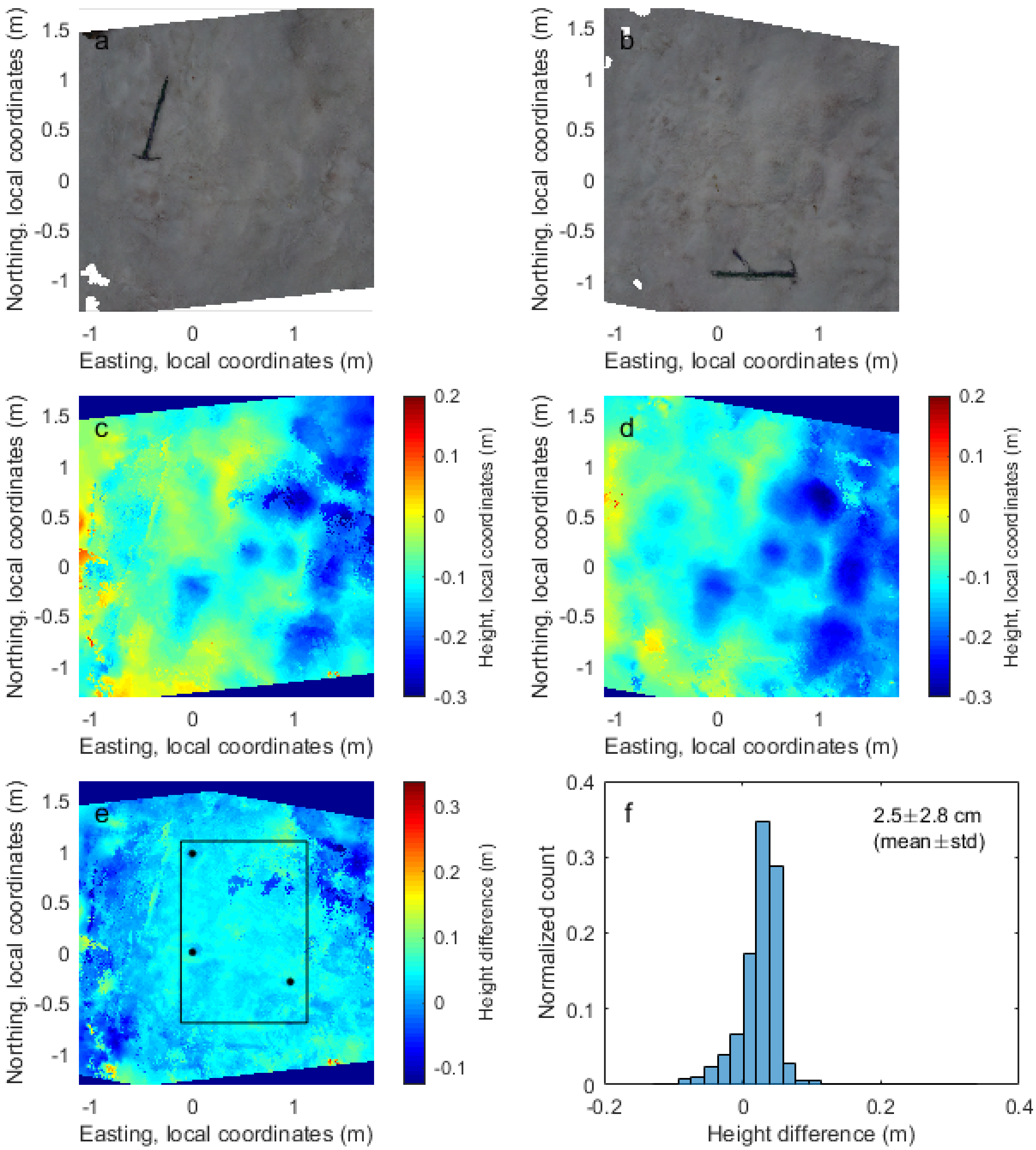



**Figure 4.10:** Photogrammetry of Ablation 3. (a) Orthophoto from 2017/8/6 4:40 pm. (b) Orthophoto from 2017/8/8 11:40 am. (c) DEM from 2017/8/6. (d) DEM from 2017/8/8. (e) DEM difference showing the area used for statistics (black rectangle) and 3 ablation stake locations (black dots). (f) Histogram of the height difference caused by melt over the period 2017/8/6-8.

**Table 4.3:** Weather measurements were recorded over the period 2017/8/5-8 at various times, generally in tandem with field measurements.

| Date | Time | T (°C) | RH (%) | Wind (m/s) | Direction | Cloud fraction | Notes |
|---|---|---|---|---|---|---|---|
| 2017/8/5 | 5:30 PM | | | | | 0.7 | light rain |
| 2017/8/6 | 12:45 PM | | | | | 0.5 | |
| 2017/8/6 | 2:00 PM | | | | | 0.6 | |
| 2017/8/6 | 2:23 PM | 11 | 45 | 2.4 | down glacier | | |
| 2017/8/6 | 3:18 PM | 9.9 | 38.6 | 0.6 | up glacier | 0.4 | |
| 2017/8/6 | 4:40 PM | | | | | 0.8 | |
| 2017/8/6 | 5:25 PM | 5.8 | 78.7 | 2.1 | up glacier | 0.6 | |
| 2017/8/6 | 5:25 PM | | | | | 0.6 | |
| 2017/8/6 | 6:25 PM | | | | | 0.5 | |
| 2017/8/6 | 6:32 PM | 6.7 | 82.3 | 1.7 | | 0.5 | |
| 2017/8/7 | 11:33 AM | | | | | 0.7 | |
| 2017/8/7 | 12:00 PM | | | | | 1 | |
| 2017/8/7 | 1:00 PM | | | | | 0.9 | |
| 2017/8/7 | 1:50 PM | | | | | 0.9 | sparse 2mm hail |
| 2017/8/7 | 3:00 PM | | | | | 1 | hail more frequent and bigger |
| 2017/8/7 | 4:42 PM | 5.4 | 77.6 | 0.9 | up glacier | 0.9 | |
| 2017/8/7 | 5:15 PM | 5.6 | 89 | 1.6 | up glacier | 1 | light rain |
| 2017/8/7 | 5:15 PM | | | | | 1 | light rain |
| 2017/8/8 | 10:08 AM | | | | | 0.5 | thin high clouds |
| 2017/8/8 | 12:25 PM | | | | | 1 | light rain and thunder |
| 2017/8/8 | 1:30 PM | | | | | 0.9 | |
| 2017/8/8 | 1:54 PM | | | | | 1 | |

## 4.6 In Situ Weather observations

Temperature and relative humidity are indicative and measurable variables of local weather and climate (Benn and Lehmkuhl, 2000). We collected in situ weather measurements with a hand-held weather meter (Kestral 5500 Weather Meter) and a fixed station (HOBO U23 Pro v2 Temperature/Relative Humidity Data Logger) at Ablation 1 for comparison to nearby long-term SNOTEL sites. Having a precise record of in situ weather gives us the ability to accurately determine what the weather was on the glacier, 574 m higher in elevation than the closest SNOTEL site.

### 4.6.1 Kestral

Weather measurements on Dinwoody Glacier collected with the Kestral 5500 Weather Meter included temperature (°C), relative humidity (%), wind speed (m/s) and wind direction (Table 4.3). In addition, we visually estimated the fraction of the sky covered by clouds during each weather measurement. These measurements supplement the station we set up at Ablation 1.

### 4.6.2 Temperature and relative humidity at Ablation 1

While in the field, we set up a HOBO U23 Pro v2 Temperature/Relative Humidity Data Logger at the Ablation 1 site (elevation: 3509m). The station recorded a sample every half hour.

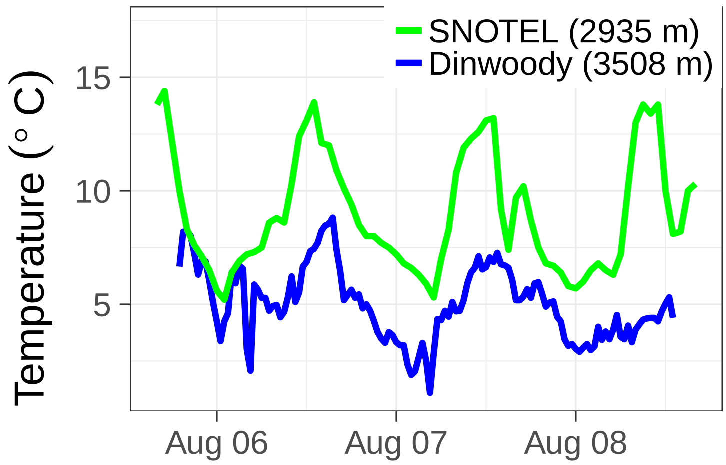



**Figure 4.11:** Temperature measured on the glacier was consistently colder than at the Cold Springs SNOTEL site from 2017/08/06-08.

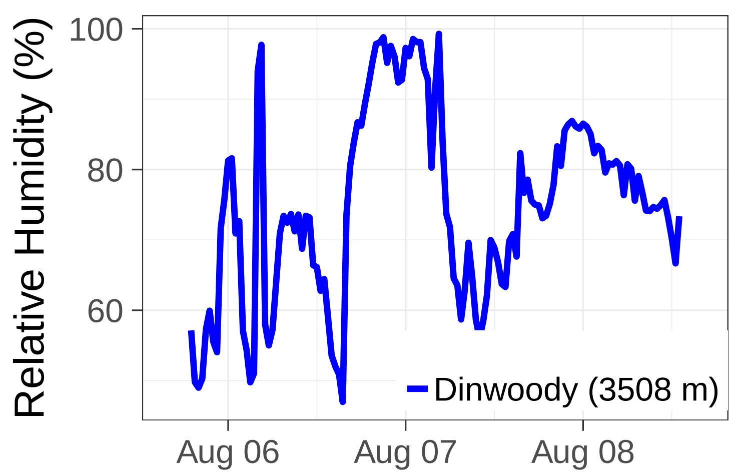



**Figure 4.12:** Relative humidity was recorded by the HOBO sensor (elevation: 3509 m) on the glacier.

**Results of in-situ weather measurements**

As compared to the Cold Springs SNOTEL site, Kestral and HOBO temperature measurements on the glacier were consistently cooler (Figure 4.11 and Table 4.3). The average temperature collected by the HOBO data logger for 2017/8/6-8 was 5 °C. The average temperature at the Cold Springs SNOTEL Site (elevation 2935 m) over the same interval was 8.67 °C. Most of the difference can be explained by the 574 m difference in elevation. The temperature difference was largest during the day time and the difference was smaller at night. This is consistent with the ice reducing the temperature of the air above the ice, an effect that is stronger when the temperature gradient is stronger (e.g. during the day when air temperatures are warmer). The mean relative humidity collected by the HOBO was 75% (Figure 4.12). Relative humidity was nearly 100% when we recorded precipitation (Table 4.3).

## 4.7 Local stream discharge

The contribution of glacial meltwater runoff to Dinwoody Creek stream discharge (see location on the basemap: Figure 3.1) was measured via a salt dilution gauging experiment. In general, dilution experiments are used to study water discharge, time-of-travel for solutes, diffusion and dispersion (Day, 1976). The theory backing "tracer methods" is that an instantaneous injection of a known volume of tracer into a flowing stream with unknown discharge can, when measured, provide a mean motion of the tracer particles which is indicative of overall stream discharge. For our purposes, local stream discharge was measured to quantify the hydrologic inputs of Dinwoody Glacier meltwater runoff into Dinwoody Creek Watershed (257 $km^2$). Discharge ($Q$) is calculated utilizing equation 4.1 (Hudson and Fraser, 2005), where $M$ is the mass of salt in grams and $A$ is the area under the curve of the graph of concentration over time (Figure 4.13).

$$Q = M/A \tag{4.1}$$

Quantity $A$ in equation 4.1 can be calculated as:

$$A = \Sigma c_t \cdot t_{int} \tag{4.2}$$

The quantity $c_t$ is the concentration of salt at time $t$, and $t_{int}$ is the time interval between data points. $c_t$ is measured as electrical conductivity ($EC$) within the stream and is calculated with equation 4.3.

$$c_t = (EC_t - EC_0) \cdot CF \tag{4.3}$$

where $EC_0$ is the background electrical conductivity and $EC_t$ is measured at time $t$. $CF$ is an empirical concentration factor, set to 1 in our case, because we used a temperature-compensated logger.

The experiment began at 4:30 pm on 2017/08/08. First, we measured the background concentration of total dissolved solids in the stream using an Etekcity TDS conductivity sensor. The background concentration was 3 ppm. Then 100 mg of salt was diluted in a bucket and the solution was dumped into Dinwoody Creek below the Little Ice Age moraine and above a section of rapids that filtered into a single constriction of the stream (Figure 4.14). The concentration was recorded at a 10 second interval for the first test, and at five second intervals for the second and third tests. For each test, the concentration peaked and then once the saline solution passed, the conductivity meter returned to the background concentration (Figure 4.13). The measurements for the second and third tests were taken 10 meters farther downstream to allow complete mixing of the salt upstream from the sensor. Due to the incomplete mixing in the first test, we disregarded it in our analysis.

**Results of local stream discharge**

The watershed area draining to the location of our discharge measurements was 7.0 km$^2$ and included Dinwoody Glacier, Gooseneck Glacier, and rocky areas with very little vegetation. The mean discharge of the stream measured with the salt dilution gauging experiment was 0.89 m$^3$/s.

The ratio of glacier-wide runoff (section 4.4) to local stream discharge was 0.54 / 0.89 or 61%.

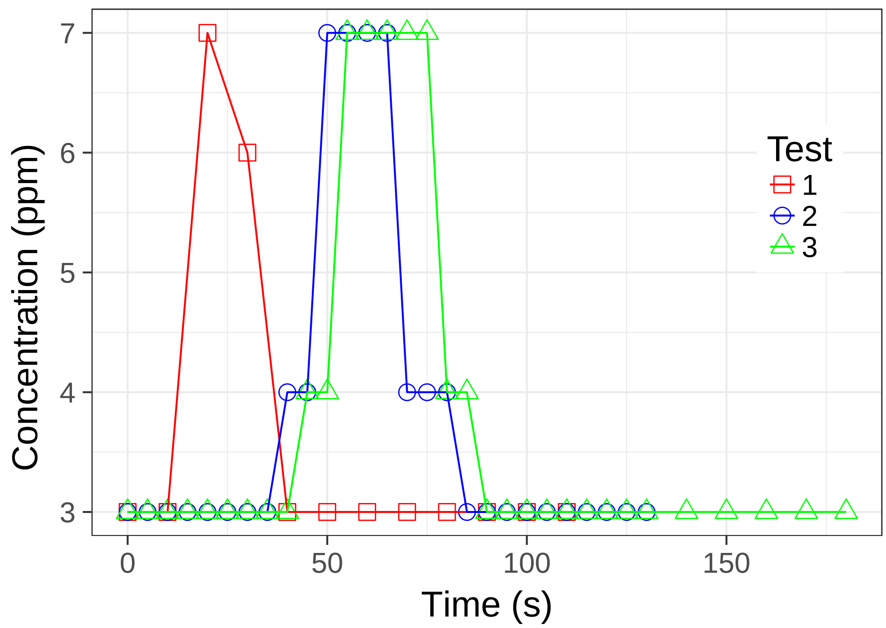



**Figure 4.13:** The salt-dilution experiment measured total dissolved solids through time after injection of a know quantity of salt into the stream. From this, we quantified the contribution of glacial meltwater runoff to Dinwoody Creek stream discharge.


**Figure 4.14:** Salt dilution gauging experimental setup. The salt injection site was just behind the photographer and the concentration measurements were at the labeled locations.

A USGS gauge (Figure 3.14) located 51 river kilometers below the glacier (Dinwoody Creek Above Lakes Near Burris WY, 1981 m a.s.l., drainage area = 228 $km^2$), reported a mean discharge of 9.0 $m^3/s$ for that day (Wyoming Water Science Center), meaning the upper Dinwoody basin contributed about 10% of the total runoff from just 3.1% of the total area.

# Chapter 5

# Discussion

## 5.1 Comparison of glacier melt and stream discharge

Glacial meltwater contributions to stream discharge have been decreasing steadily since the 1950's (Cable et al., 2011; Cheesbrough et al., 2009; DeVisser and Fountain, 2015). Utilizing our observed ablation rate on 2018-08-08, we calculated the amount of stream discharge at the terminus of Dinwoody Glacier that could be attributed to glacial meltwater input to Dinwoody Creek to be 61% (drainage area: 7.0 $km^2$). On the same day, we observed the discharge recorded at the associated USGS stream gauge (Dinwoody Creek Above Lakes Near Burris WY, 1981 m a.s.l.), 35 km NE of the glacier and determined that 10% of the total runoff that day could be attributed to glacial meltwater runoff (drainage area: 228 $km^2$). This means that the upper Dinwoody basin contributed about 10% of the total runoff from just 3.1% of the total area. The importance of this comparison is that, the presence of glaciers in a watershed can alter the streamflow power (Maloof et al., 2014).

Our calculations are consistent with the finding of Maloof et al. (2014), whose research determined 70% of streamflow in the Wind River Range is coming from snow and glacial melt (Figure 3.14). Isotope analysis of Dinwoody Creek by Cable et al. (2011) found that glacial meltwater contributed (53 – 59%) to streamflow in a low-flow year (2007). The difference between Cable et al. (2011) and our work is that they differentiated between snowmelt and glacial meltwater runoff. This is a more precise way to understand how snowmelt and glacial meltwater contribute to Dinwoody Creek at different times in the season. DeVisser and Fountain (2015) found glacial meltwater contributions to total discharge measured at the associated USGS stream gauge (Dinwoody Creek Above Lakes Near Burris WY, 1981 m a.s.l.), 35 km NE of the glacier contributed 3.45-10.94%, again, lining up with our findings.

The consistency of reported decreases in stream discharge have led us to the conclusion that glacial meltwater runoff is a critical contribution to overall steam discharge in Dinwoody Creek Watershed.

## 5.2 Comparison of glacier climate to SNOTEL site climate

Temperature measured on the glacier was consistently colder than at the Cold Springs site while we were in the field. The 575 m elevation difference explains a large portion of the difference, though nighttime temperatures on the glacier were relatively warm (Figure 4.11). In situ weather measurements are important for calculating average lapse rates for temperature, as well as calibration of temperature inputs to hydrology models we'd like to create in the future. Other differences in the glacier's climate to the SNOTEL data are likely due to topographic influences such as storms occurring at high elevations and not at the elevation of Cold Springs SNOTEL site. DeVisser and Fountain (2015) found that elevation and temperature were the two variables most correlated to total surface area of glaciers and perennial snow. As was noted in their work, elevation does not "drive" the formation of glaciers rather, it sets the stage for where they can exist. And if elevation sets the stage, increasing temperatures at lower elevations create the physical extent or boundary of where glacier ice remains frozen. Lower elevations receive less precipitation as snow, less snow accumulation and increasingly more winter rain, yielding decreased snow persistence (McCabe et al., 2007). Increased temperatures across the Wind River Watershed, especially in the spring and autumn have changed the dynamics of ecosystem's function across the elevation gradient. At lower elevations, where the Wind River Indian Reservation lies, interviews with Tribal Members have noted "changes in timing and seasonality of berry harvest and migration patterns of wildlife" within the lifetime of some tribal elders (McNeeley and Beeton, 2017). Up in the alpine zone, higher temperatures lengthen the ablation season (DeVisser and Fountain, 2015). The extent of Dinwoody Glacier has decreased by 34.7% since 1950, therefore, it is highly likely that temperature is increasing at the elevation of its terminus.

A study done on the Upper Fremont Glacier (7 km north of Dinwoody Glacier) by Naftz et al. (2002) indicated a 3.5°C increase in temperature since the 1960's based on a strong correlation between $\delta^{18}$O levels in snowfall and ice cores. Furthermore, Thompson et al. (2011) examined temperature increases using local SNOTEL sites and found that for eight sites around the Wind River Range, average temperature increased in July and August by 3°C± from 1986 to 2008. Patterns of precipitation recorded at individual SNOTEL sites don't have a distinct trend up or down. However, studies examining the Wind River Range as a whole find decreases in total precipitation with the largest monthly decreases occurring in late fall and early spring (Rice et al., 2012).

# Chapter 6

# Conclusions

For both riparian ecosystems and water users dependent on Dinwoody Glacier's contributions to water supply, the rapid decline in glacier mass will continue to yield diminishing meltwater runoff (Barnett et al., 2005; Kehrwald et al., 2008). Climatic trends display a rapid climate change that has altered the reliability of water resources within Dinwoody Creek, Wind River, and Missouri River watersheds. Water resources are particularly vulnerable as warmer temperatures are projected to reduce snowpack, increase evaporation, lengthen summer seasons, and start spring runoff earlier (Rice et al., 2012). Today, warmer temperatures, and extreme droughts in the area compromise the livelihoods of tribal members and local ranchers (McNeeley et al., 2017; Ojima, 2015); while the paradigm of "stationarity" within water management practices has needed to change with altered predominant climatic trends in the region.

Our research found 2017 to be a particularly heavy snow year, both at the Cold Springs SNOTEL station and on Dinwoody Glacier. We found ablation rates on the glacier between 2.05 cm/day and 2.46 cm/day in early August 2017 using traditional ablation stakes and novel photogrammetric techniques. Meltwater runoff from Dinwoody Glacier contributed 61% of the flow of Dinwoody Creek in the upper Dinwoody basin in early August 2017. In turn, runoff from the upper Dinwoody basin contributed 10% of the water within Dinwoody Creek Watershed though it is a mere 3.1% of the total watershed area.

Modernly and looking into the future, warmer temperatures are likely to lead to reduced stream flows, which are critical to habitat and reservoir storage for agricultural and human uses. The rapid recession of Dinwoody Glacier as told by the 2016 accumulation area ratio of 21% is of the same magnitude seen across the continental U.S.: the Colorado Front range has seen a 34% decrease in glacial and perennial snow surface area (Hoffman et al., 2007) and the Sierra Nevada has experienced a 55% decrease (Basagic and Fountain, 2011) in the 20th century. The anticipated decline of Dinwoody Glacier, and others like it, will increase drought-stress downstream, directly impacting Wind River Indian Reservation tribal members and other water users of the Missouri River Watershed.

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