The Effects of a Climate Manipulation Experiment on Snow Properties, Snow Surface Energy Balance, and Soil Temperature and Moisture Along an Elevational Gradient on Niwot Ridge, Colorado

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THE EFFECTS OF A CLIMATE MANIPULATION EXPERIMENT ON SNOW PROPERTIES, SNOW SURFACE ENERGY BALANCE, AND SOIL TEMPERATURE AND MOISTURE ALONG AN ELEVATIONAL GRADIENT ON NIWOT RIDGE, COLORADO

By JENNIFER PETRZELKA

B.A., Geography, Texas State University, 2005

A Thesis submitted to the Faculty of the Graduate School of the University of Colorado in partial fulfillment of the requirement for the degree of Master of Arts

Department of Geography 2011
This thesis entitled:

THE EFFECTS OF A CLIMATE MANIPULATION EXPERIMENT ON SNOW PROPERTIES, SNOW SURFACE ENERGY BALANCE, AND SOIL TEMPERATURE AND MOISTURE ALONG AN ELEVATIONAL GRADIENT ON NIWOT RIDGE, COLORADO

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The final copy of this thesis has been examined by the signatories, and we find that both the content and the form meet acceptable presentation standards of scholarly work in the above mentioned discipline.
ABSTRACT

Petrzelka, Jennifer (M.A. Geography)

The Effects of a Climate Manipulation Experiment on Snow Properties, Snow Surface Energy Balance, and Soil Temperature and Moisture Along an Elevational Gradient on Niwot Ridge, Colorado

Thesis directed by Professor Mark W. Williams

This research investigated the influence of near-infrared heaters used in climate manipulation experiments on snow properties (e.g. snow depth, grain size, grain shape, density, temperature), snow surface energy balance, and soil temperature and moisture at three sites along an elevational gradient on Niwot Ridge, CO. At the lower subalpine site (LSA), heated plots experienced an ephemeral snow cover, never reaching more than 45 cm of snow. Snow depths between heated and control plots at the upper subalpine (USA) and alpine (ALP) were similar but snow disappearance occurred 0 to 18 days earlier in heated plots relative to controls. Heated plots in all three sites experienced warmer soil temperatures and higher soil moisture during the winter relative to control plots. Overall, the effect of the heaters on snow properties, soil temperature, and soil moisture decreased with increasing elevation.

Heaters altered the snow surface energy balance by increasing incoming longwave radiation (LW_{in}). In order to derive estimates of energy and mass balance exchange at the snow surface in heated and control plots, the one-dimensional, physically based snowmelt model SNOWPACK was used. In heated plots at the LSA, net radiation accounted for 80 to 100 % of the energy available to melt snow compared to 35% in control plots. In heated plots at the USA/ALP, net radiation accounted for 100% of the energy available for melt when snow depth exceeded heater height. However, when snow depth is below the heaters (1.2 m), only 5% of the
energy for snowmelt comes from net radiation. Model results illustrate greater mass losses to
sublimation/evaporation (54 to 83% of total SWE) in heated plots compared to control plots (6 to
38% of total SWE). The results of this study will aid in the interpretation of warming
experiments, as well as develop a better understanding of the interactions between climate,
hydrology, and ecological processes.
ACKNOWLEDGEMENTS

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Influence of near-infrared heaters on snowpack properties and soil temperature and moisture along an elevational gradient on Niwot Ridge, Colorado
Abstract

This research investigated the influence of near-infrared heaters used in climate manipulation experiments on snow properties (e.g. snow depth, grain size, grain shape, density, temperature), snow surface energy balance, and soil temperature and moisture at three sites along an elevational gradient on Niwot Ridge, CO. At the lower subalpine site (LSA), heated plots experienced an ephemeral snow cover, never reaching more than 45 cm of snow and quickly melting after snow events. At the USA and ALP sites caves formed beneath the snow surface once snow depth exceeded the height of the heaters (1.2 m) which altered the thermal regime of the snowpack. Snow depths between heated and control plots at the upper subalpine (USA) and alpine (ALP) were similar but snow disappearance occurred 0 to 18 days earlier in heated plots relative to controls. All three sites experienced episodic snowmelt throughout the winter resulting in warmer soil temperatures and higher soil moisture relative to control plots. Overall, the effect of the heaters on snow properties, soil temperature, and soil moisture decreased with increasing elevation which suggests that as air temperatures increase with climate change, these variables will be affected more greatly at lower elevations than high elevations. Furthermore, the results of this study will aid in the interpretation of warming experiments, as well as develop a better understanding of the interactions between climate, hydrology, and ecological processes.
1.1 Introduction

Human-driven changes to the global environment – including those in climate, atmospheric composition, nutrient cycles, hydrologic cycling, and ecosystem structure – are now pervasive throughout the world, and continue to accelerate (e.g. Steffen et al. 2007; IPCC 2007; Galloway et al. 2008; Röckstrom et al. 2009). While such changes have brought substantial benefits to humanity (e.g. Smil 2001, Kareiva et al. 2007, Townsend and Howarth 2010), they increasingly cause detrimental outcomes for both people and ecosystems (Galloway et al. 2008; Carpenter 2009), including those in seasonally snow-covered systems (Bowman and Steltzer, 1998, Williams and Tonnesson, 2000; Bowman et al. 2006). Indeed, while some portions of high-elevation ecosystems are often free of direct transformation via land use change (Bourgeron et al. 2009), taken as a whole, alpine tundra and montane forests have been identified as particularly sensitive to the suite of human-induced environmental changes that currently challenge society (IPCC 2007; Williams et al. 2002).

The harsh environmental conditions characteristic of these environments suggest that organisms in montane and alpine ecosystems are on the edge of tolerance (Williams et al. 1998). Consequently, organisms and the biogeochemical processes mediated by them in high-elevation catchments are notably sensitive to small changes in climate and other environmental parameters (Williams and Tonnesson, 2000; Bowman et al. 2006). Recent climate analyses have shown widespread declines in the winter snowpack of mountain ecosystems in the western USA and Europe that are coupled to positive temperature anomalies (Laternser and Schneebeli, 2003; Scheerrer et al. 2004; Mote et al. 2005). Research at the Saddle site at Niwot Ridge in the Colorado Front Range has shown that changes in the timing, duration, and extent of snow cover affects soil moisture (Taylor and Seastedt 1994), litter decay rates (O’Lear and Seastedt 1994),
plant productivity (Walker et al. 1994), organic matter accumulation (Burns and Tonkin 1982), species diversity (Litaor et al. 2008), and fluxes of trace gases such as CO2, N2O, and CH4 (Brooks et al. 1996, 1997; West et al. 1999, Liptzen et al. 2009; Filippa et al. 2009). It also governs the microbial processes, which control gross N mineralization and N immobilization among plant communities (Fisk et al. 1998, Litaor et al., 2002). Monson et al. (2006) used a six-year record of net ecosystem carbon dioxide exchange in the subalpine forest at C1 on Niwot Ridge to show that years with a reduced winter snowpack are accompanied by significantly lower winter rates of soil respiration. Furthermore, they show that the cause of the high sensitivity of soil respiration rate to changes in snow depth is a unique soil microbial community that exhibits exponential growth and high rates of substrate utilization at the cold temperatures that exist beneath the snow. These observations suggest that small changes in the duration, depth, and timing of an insulating snow cover may cause large changes in the ecosystem processes of montane and alpine ecosystems.

Several researchers have tried to simulate the effect of a warmer climate on the seasonal snow-cover and to derive quantitative effects on biological compounds in soil. Snow-removal, as a simulation of a lack of snowcover, has been carried out world-wide in the last decade (Groffman et al. 2001; Decker et al. 2003; Freppaz et al. 2008), while other researchers have used snow fences to experimentally manipulate snow accumulation (Williams et al. 1998; Nobrega and Grogan 2007), or grooming to change its density (Rixen et al. 2008). A decreasing of winter precipitation may result in shorter winter seasons, in more pronounced and more frequent freeze/thaw cycles, and in more days with the soil temperature well below 0°C during winter. The experiments sometimes indicate as a consequence of these phenomena a faster mineralization of N (Panikov and Dedysh 2000; Grogan et al. 2004; Freppaz et al. 2007a), higher
N$_2$O emissions related to freeze thaw cycles (Sharma et al. 2006), and a reduction of respiration rates (Mariko et al. 1994; Melloh and Crill 1996; Brooks et al. 1997; Welker et al. 2000; Nobrega and Grogan 2007). At the same time, a lower mineralization of N (Walker et al. 1999; Schimel et al. 2004), lower N$_2$O emissions (Goldberg et al. 2008), an increase in respiration rates (Goldberg et al. 2008), were also indicated as the consequence of comparable experimental simulations, making it considerably difficult to derive any general conclusion.

Currently, the use of thermal radiation via overhead infrared (IR) heaters to warm plots is an appealing method because it most closely simulates global warming and appears to have the fewest limitations (Harte et al., 1995; Kimball et al., 2008; Amthor et al., 2010). Many researchers have implemented IR heaters to investigate various ecosystem responses to global warming (e.g. Harte and Shaw, 1995; Kockelbergh et al., 2000; Price and Wasser, 2000; Wan, 2002; Adler et al., 2007; Wall et al., 2011). While the dynamics of soil and plant energy exchanges in a warming climate is well documented, there has been little research on how such warming experiments affect snowpack properties, in particular the energy balance of the snow surface.

Various processes affect the energy balance of the snow surface. Incoming shortwave (solar) radiation, outgoing (reflected) shortwave radiation, incoming longwave radiation, and outgoing (emitted) longwave radiation, as well as sensible and latent heat fluxes are avenues of energy lost or gained from the snowpack. The sum of these parameters determines the total amount of energy available for snowmelt. The exchange of radiative and turbulent energy between the snow and atmosphere has been well studied (Cline et al., 1997a,b; Link and Marks, 1999; Pohl et al. 2006a,b; Molotch et al., 2009). However, the complexity of processes involved in the energy exchange between the snow surface and atmosphere complicates quantification of
these fluxes. What remains unknown is how the additional $LW_{in}$ from the heaters alters energy exchange at the snow surface and subsequently, snow depth, density, grain morphology, and temperature.

Knowledge of these snow properties provides important information on the hydrology in each plot to gain further understanding on the impact a changing climate may have on the ecology and biology of these environments. An important aspect in interpreting these experiments, as well as developing a better understanding of interactions between climate, hydrology, and ecological processes is how the snowpack responds to radiative forcing caused by IR climate manipulation experiments. At Niwot Ridge, Kueppers and colleagues have initiated a global warming experiment using IR heaters, at three sites along an elevational gradient.

This research aims to address the influence that these IR heaters may or may not have on snowpack properties. Research questions are:

- How do the warming experiments affect the duration, timing, accumulation and depth of the snowpack?
- How does snow water equivalent in the warming plots differ from control plots at each site?
- How do the warming experiments change snowpack stratigraphy, e.g. grain size, grain type, more or less ice lenses, etc?
- How does warming affect soil microclimate (temperature and moisture) relative to control plots?
• How do these same variables change with elevation? E.g., does the same amount of climate forcing (warming experiments) have a greater effect at lower elevations than higher elevations, or vice versa?
1.2 Background

For over two decades scientists have been implementing and improving upon various climate manipulation experiments to study the responses of ecosystems to global warming. According to the IPCC, global warming may correspond to an increase in radiative forcing of 4-9 Wm$^{-2}$ by the end of the century resulting in an increase in global mean temperatures of 1.5-6 °C. Additionally, absolute humidity will increase, while relative humidity is to remain constant (IPCC, 2007). Therefore, experiments should warm air an amount representative of global warming predictions, as well as humidify air to maintain constant humidity (Kimball et al., 2005).

Various researchers have implemented passive experiments that manipulate the local environment such as field greenhouses (Shaver et al., 1998), open-top chambers (Marion et al., 1997), and passive nighttime warming via shading (Emmett et al., 2004) to study the effects of climate change on ecological systems. These methods work by retaining heat to increase air temperatures relative to control plots but have the side effect of enriching the air with CO$_2$. There is no direct control on energy flux often resulting in large differences in the range and variability of temperature between heated and control plots (Marion et al., 1997). Another limitation includes the ability to allow shortwave radiation in, but prevent IR from escaping (Amthor et al., 2010; Kimball, 2005). Thus, difficulty arises in controlling gas concentrations while also minimizing increasing air temperatures (Kennedy, 1995). Other limitations include the altering of wind, precipitation, and humidity regimes due to the enclosed structures (Shaver et al., 2000).

On the other hand, ‘active’ methods supply the system with external heat at or above the soil, thus controlling the energy flux (Kimball et al., 2008). Sensitive temperature regulation is
required in order to maintain a constant temperature difference between the heated and control plots (Peterjohn et al., 1994). Heated cables/wires (Van Cleve et al., 1990; Peterjohn et al., 1994) and fluid heated pipes (Chapin and Bloom, 1976) have been placed above or buried beneath soil. Because these methods only heat the soil, they limit the ability to study the relationships between air temperature and soil temperature effects on an ecosystem (Aronson and McNulty, 2009; Kimball, 2010).

The use of thermal radiation via overhead IR heaters to warm plots appears to have the fewest limitations and most closely simulates climate change predictions by increasing the amount of incoming longwave radiation (Shaver et al., 2000; Kimball, 2010; Aronson, 2010). As a result, air temperatures rise increasing the temperatures of the surface (plants, soil, and snow) via conduction and convection (Amthor et al., 2010). Conversely, thermal (longwave) radiation from the heaters does not directly heat the air temperature above the plots, rather the energy that the leaves and soils absorb is transferred to the canopy air via sensible/latent heat transfer (Shaver et al., 2000). Thus, ambient air temperatures are very little affected by IR heating (Kimball, 2010). As a result, experiments must supply significantly more incoming longwave radiation than the projected 4-9 Wm\(^{-2}\) in order to increase the temperature of plants and soils by the amount expected with global warming. Furthermore, artifacts from the interaction of thermal radiation with wind, radiation, or precipitation are reduced in comparison to open or closed top chambers (Harte et al., 1995) making overhead IR heaters a viable means to study the effects of global warming on ecosystems.

Harte and Shaw (1995) first deployed overhead infrared heaters at the plot scale using a "constant flux" approach whereby heaters were suspended a certain distance above the plot and operated at full power to maintain an energy flux of 22Wm\(^{-2}\). Subsequent studies also used this
approach (Harte et al., 1995; Wan et al., 2002). However, using the theoretical analysis of Kimball et al. (2005), Wan et al. (2002) found that warming of plots was much greater at night due to more turbulent daytime conditions. Further analysis showed that the amount of energy flux needed to increase daytime temperatures to the desired level resulted in nighttime temperatures that were 10-20°C higher. In contrast, techniques developed by Nijs et al. (1996) and improved upon by several others (Nijs et al., 2000; Kimball, 2005; Kimball et al., 2008) uses a “constant temperature rise” approach. This system works by controlling the energy output of the heaters to maintain the temperature rise of the heated plots by a set amount above that of the corresponding reference plots, thus more closely simulates global warming predictions. To accomplish this, measurements of leaf (Nijs et al., 1996; 2000) and canopy (Kimball, 2005; Kimball et al., 2008) temperature in feedback control systems are necessary. Using this method, Nijs et al. (2000) successfully maintained a 2.5°C difference in surface temperature between heated and control plots in an Arizona agricultural site by adding temporally varying amounts of IR to the plots (often IR input is really high). Such a constant temperature increase approach is more challenging to implement in highly variable natural systems, without uniform canopies.

There are many complexities involved in predicting how ecosystems may respond to climate change. No one method perfectly simulates the expected climatic changes, but with cautious interpretation can reveal similarities, differences, and potential responses to a changing climate. While the effects of experimental warming on microclimate and ecological response have been extensively studied for the growing season (Harte and Shaw, 1995; Hart et al., 1995; Kockelbergh et al., 2000; Price and Wasser, 2000; Wan, 2002; Dunne et al., 2004; Harte et al., 2006; Adler et al., 2007; Wall et al., 2011), the influence of warming on snow properties and soil microclimate during the winter is severely lacking. During the winter months, snow generally
covers the ground affecting the energy exchanges between the heaters, snow surface, and underlying vegetation and soil. Thus, knowledge of how the snowpack and soil responds to the radiative forcing imposed by IR heaters will help facilitate interpretation of warming experiments, as well as develop a deeper understanding of how global warming may effect ecological and hydrological systems.
1.3 Site Description

Research was conducted during the winter and spring of 2010 on Niwot Ridge, located in the Colorado Front Range of the Rocky Mountains about 5 km east of the Continental Divide (40°03’ N, 105°35’ W). This site is an UNESCO Biosphere Reserve and the location of the Niwot Ridge (NWT) Long-Term Ecological Research (LTER) site. Warming experiments were located at three sites located along an elevation gradient within the Niwot Ridge LTER research area, the Lower Alpine Site (LSA; 3048 m), the Upper Subalpine (USA; 3367 m), and the Alpine site (ALP; 3517 m) (Figure 1.1).

Figure 1.1: View of Niwot Ridge and locations of LSA, USA, and ALP sites. LTER managed sites are also shown (C1, Saddle, and D1). Inset shows the layout of sensors and equipment for each site. Each heated plot contains 6 heaters placed in a hexagonal pattern. Control plots are located within ~3 meters of the associated heated plot. Snow depth sensors were suspended over the center of each plot by a horizontal boom. Soil moisture and temperature sensors at 15-20 cm depth were located in the center of each plot. Soil temperature and moisture sensors at 5-10 cm depth were located in each quadrant of the plot, ~80 cm from the corner.
The LSA is located within a closed-canopy subalpine forest (forest height about 20 m). Less than a half km from LSA, the NWTLTER collects numerous measurements at the C-1 site, including meteorological data, the SnoTel network site NIWOT 663, and soil moisture and temperature. This site has a mean annual temperature of 1.5 °C and receives about 800 mm of precipitation annually, with 60% as snow, and 40% as rain.

The USA site is located near treeline in an opening within a stand of krummholz consisting of Engelmann Spruce (Picea engelmannii) and Subalpine Fir (Abies lasiocarpa); maximum tree height is 3 meters. The ALP site is characterized by alpine tundra, located approximately a one km to the west of USA and is located in a moist meadow plant community with vegetation height generally less than 20 cm. Near both the USA and ALP sites, the NWTLTER program maintains instrumentation similar to C1 at The Saddle (40°03’17”N; 105°35’ 21”W; 3528m), which is located in alpine tundra. The Saddle maintains a meteorological tower capable of closing the energy balance, a snow lysimeter array, an index snowpit, and a National Atmospheric Deposition Program (NAPD) precipitation collector (site CO02). The climate of the Saddle area consists of long, cool winters and a short growing season of one to three months. Mean annual temperature is -3.7 °C and annual precipitation is 1000 mm (Williams et al., 1996) with approximately 80% of the annual precipitation falling as snow (Caine, 1995). Snow depth accumulation is extremely variable due to high winds and topography. Snow cover generally lasts October to June. The alpine portion of Niwot Ridge experiences strong westerly winds that commonly redistribute snow (Erickson et al., 2005).
1.4 Methods

1.4.1 Experiment Design

Within each UC Merced site were replicated heated and control plots separated by about 3 m, with each pair of controls and heated plots considered a treatment (Figure 1.1). Each plot was 3 m in diameter (7.07 m²). Heater arrays were designed according to Kimball et al. (2008). Each heater was mounted 1.2 m above the ground. There were 6 heaters per heated plot arranged in a hexagonal pattern with each tilted at 45° from horizontal to achieve the most uniform distribution of radiation (Kimball, 2005) (Figure 1.2). For this study, lasting from October 2009 to July 2010, the number of plots used varied with site, with LSA containing four pairs of heated and control plots, USA three pairs, and ALP had five pairs (Table 1.1).

Figure 1.2: An image of a heated plot from the ALP site in November. There are six heaters arranged in a hexagonal array around a 3-m diameter circle, each heater tilted downward at 45°.
<table>
<thead>
<tr>
<th>Plot</th>
<th>Sensor Height (cm)</th>
<th>Max snow depth (cm)</th>
<th>Date of snow free</th>
<th>Sensor Height (cm)</th>
<th>Max snow depth (cm)</th>
<th>Date of snow free</th>
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<td>6/1</td>
<td>235</td>
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<td>148</td>
<td>6/1</td>
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<td>220</td>
<td>33</td>
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<td>6/7</td>
</tr>
<tr>
<td>ALP 75</td>
<td>310</td>
<td>88</td>
<td>6/3</td>
<td>295</td>
<td>80</td>
<td>6/4</td>
</tr>
</tbody>
</table>

Table 1.1: For each heated and control plot, the sensor height above the ground, maximum snow depth, and date of snow disappearance is listed.

Heated plots were warmed using Mor Electric Heating Association Inc. IR ceramic heaters (Model FTE-1000). Each heater was 245 mm long x 60 mm wide, with a maximum output of 1000 W. Energy emittance was in the long-wave portion of the EM spectrum at 4.5–42 μm.
From an energy balance perspective, the energy from the heaters was an increase in incoming longwave radiation ($L_{in}$). In order to achieve a temperature target of $\sim$4-4.5 °C increase in surface soil temperatures averaged over the growing season, heaters were operated in a constant flux mode according to the protocols of Harte et al. (1995) and Harte and Shaw (1995). Because soils at Niwot Ridge are wetter than those of Harte et al. (1995), more than double the radiation is needed to double the temperature effect since more of the increased radiation goes to evaporate water. Beginning October 2009 heaters were turned on and operated at 50% power, thus each heater emitted about 500 W. It is estimated that $\sim$50% of the output is lost outside of the plot (Kimball, 2005), reducing the 500 W to 250 W. Since each heated plot was roughly 7 m$^2$ in area, if the heat was evenly distributed the energy flux would be 214 W m$^{-2}$ or 1.85 x 10$^7$ J m$^{-2}$ d$^{-1}$. Heaters were programmed to turn off in high winds ($>15$ m s$^{-1}$).

Soil moisture was measured with Decagon EC-TM sensors depths of 5–10 cm depth and 15-20 cm. Soil temperature was measured by EC-TM thermistors at 5-10 cm and 15-20 cm depths. Soil temperature and moisture sensors at 5-10 cm depth were located in the center of each quadrant of the plot ($\sim$80 cm from the corner of each quadrant) and sensors at 15-20 cm depth were located at the center of the plot. Snow depth at the edge of each plot was measured by hand about every two weeks, starting in November 2009. Located in approximately the center of each site was a 3 cup anemometer (RM Young Model 03101-L) measuring wind speed, and a Visalia HMP 45 six-wire sensor mounted inside a Gill aspiration shield measuring temperature and relative humidity. All sensors were mounted 3 meters above bare-ground. Microclimate data was logged as 15 min averages of 1-s readings.

In addition to the existing instruments, Judd Communications LLC Ultrasonic snow depth sensors were installed in each plot, which became operational in late January (ALP),
March (LSA), and early April (USA) of 2010. Each snow depth sensor was 8 x 8 x 13 cm in size with a beamwidth of 22° and an accuracy of 1 cm. Depth was recorded hourly with Campbell CR1000 and CR10X dataloggers. Each snow depth sensor was located at various heights above the above the ground so as to avoid being buried by the seasonal snowpack (Table 1.1) and located over the center of each plot on an array of pipes extending from the sides of the plots, so as to avoid possible heat conduction if a support pipe were placed in the plot itself.

1.4.2 Snow properties

Snowpits at the Saddle and C-1 sites were sampled approximately weekly for physical and chemical parameters as part of the NWT LTER project. Snowpits were excavated to sample snow properties from the snow/air interface to the snow/ground interface following the protocols of Williams et al. (1999). Density was measured in vertical increments of 10 cm using a 1-L (1000 cm$^3$) stainless steel cutter and an electronic scale (+/- 2 g). Temperature of the snowpack was measured every 10 cm with 20–cm long dial stem thermometers, calibrated using a one-point calibration at 0°C. The height of stratigraphic layers above the snow/ground interface was recorded, along with the thickness and type of layer (buried sun crust, ice lens, coarse to fine grain transition), and grain type and grain size of each layer were determined using a 10x magnifying loupe and a gridded crystal card. The working wall of the snowpit was oriented so that it remained shaded from the sunlight. The snowpit was refilled after measurements were taken. Because the snowpit results in destructive sampling, the next snowpit was excavated approximately 1 meter from the southern wall of the last snowpit to avoid edge effects. Thus snowpits did not sample exactly the same snowpack at the same location. Additional snowpits were excavated in a few selected experimental plots for more direct comparison of snow
properties; however regular snowpits were not conducted within the plots because of the concern that such sampling might induce artifacts to the experimental treatments.

Snowpit measurements of snow water equivalent (SWE) were supplemented using a Federal sampler. The sampler consists of an aluminum tube that is lowered vertically through the snowpack to obtain a core of snow. The tube is then weighed to obtain a measurement of water equivalent. Measurements at C-1 were made weekly beginning in November. Measurements in experimental plots were attempted bimonthly beginning in March, however high winds prevented consistent measurements in USA and ALP sites. Heated plots at LSA were unable to be sampled due to lack of a consistent snowpack.
1.5 Results

1.5.1 LSA/C-1 site

1.5.1.1 Climate

The mean daily air temperature at LSA dipped below freezing in November and remained between -10 and -5 °C through the end of February (Figure 1.3). Air temperature during March was somewhat warmer, with mean daily air temperature reaching the freezing point on about 1 April and remaining near freezing until about mid-May (Figure 1.3). Starting on 14 May there was a jump in daily mean air temperature to 5 °C or more. The area was sheltered, with mean daily wind speeds always below 1.5 m s\(^{-1}\) and usually below 0.5 m s\(^{-1}\) (Figure 1.3).

![Daily Mean Air Temperature and Wind Speed](chart.png)

Figure 1.3: Mean daily air temperature and wind speed for LSA, USA, and ALP sites.
Snotel measurements of SWE at C1 showed that snow began accumulating around the beginning of November, reached a seasonal maximum of 310 mm in late March, with the last snow melting on about 31 May 2010 (Figure 1.4). The seasonal maximum SWE was within 3% of the 25-year mean value of 318 mm. Early to mid-winter at the LSA for 2009 to 2010 was characterized by frequent small snow events with several larger precipitation events occurring at the end of March and late April to early May. Snow depths at the NWTLTER index pit at C1 ranged between 50 and 110 cm, reaching a maximum around 7 April, coincident with the mean daily air temperature reaching the freezing point (Figure 1.4). The presence of a crust layer at the base of the snowpack persisted throughout the winter (Figure 1.4). Above the basal layer were well-developed facets about 20-30 cm in thickness. Around 24 March, the snowpack depth increased to 100 cm due to a large storm event that occurred beginning 22 March and ending 24 March. Several thin crusts and ice layers formed as a result of warmer temperatures throughout the end of March. The snowpack evolved into 40-75% wet metamorphism grains until the end of melt. The snowpack became isothermal on 20 April, but the relatively cool spring air temperatures delayed appreciable melt until 14 May, at which point the entire snowpack melted over a few days.
Figure 1.4: Upper figure: Daily SWE from Snotel-663 at C1 is graphed as a solid black line. Bottom figure: Evolution of C1 snowpack represented by weekly snowpit profiles. Four grains types are represented according to the legend in the top right corner. Mean snow temperature (red) and mean density (blue) is located above each stratigraphic profile.
Hand measurements of snow depth in the control plots at LSA tracked those of the C1 index plot (Figure 1.5). In the heated plots, snow depth on 16 November was 10 cm, 30% the amount of snow present in the control plots (Figure 1.5). From December to the beginning of March, snow depth varied between 15 and 45 cm. Beginning on 5 March, snow was present only after new snowfalls and melted quickly after each precipitation event, never reaching more than 20-30 cm of snow depth. One exception was an event that occurred on 8 January that resulted in ~45 cm of snow in plots 6h and 14h. Manual observations after 30 March report no snow in the heated plots.

Figure 1.5: Manual snow depth measurements of LSA heated and control plots. Measurements were taken approximately bi-monthly.
Automated hourly measurements of snow depth at the LSA plots began in early March. Snow depths in the control plots followed the pattern of the C-1 index pit, reaching maximum snow depths in late March-early April (Figure 1.6). There was some spatial variation among plots, with maximum snow depth ranging from 110 to 148 cm (mean of 123 cm, n = 4) (Table 1.1). As with the C1 index pit, snow depths remained relatively constant until mid-May, when snowmelt began in earnest with most snow gone a week later. In the heated plots, the automated measurements were able to capture transient snow accumulation and melt. Because manual measurements were made at the edge of plots and automated measurements were made at the center of plots, some discrepancies are apparent between snow depth measurements.
The distribution of snow in heated plots was heterogeneous, with more snow in the center of the plot versus beneath the heaters and periphery of the plot (Figure 1.7, left). By March, a snow-free zone extended 0.5 to 1 m outside heated plots (Figure 1.7, right). At the snow surface a layer of surface hoar was evident beneath the heaters during times of snow accumulation (Figure 1.8). On top of the surface hoar layer, a melt-freeze would form, followed by another
layer of surface hoar. The melt-freeze crusts were not present in control plots and thus suggest that the heaters did in fact cause melt at the snow surface underneath the heaters.

Figure 1.7: Pattern of snow accumulation/removal in heated plot 6 at the LSA site on March 15 (left). Extent of snow-free zone beyond heated structures at the LSA plot on May 2, 2011 (right).

Figure 1.8: Photograph of melt/freeze crusts that formed below the heaters at the LSA from plot 6 (left). An example of surface hoar that formed below heaters, taken of ALP plot 75 (right). Photographs were taken April 20, 2011.
Knowing that the heaters theoretically provided $1.85 \times 10^7$ J m$^{-2}$ d$^{-1}$ of energy ($Q_m$) to each plot, the amount of liquid water production occurring during the winter from the heated plots can be calculated as:

$$\text{Melt}(\text{mm d}^{-1}) = \frac{Q_m}{\rho_{\text{liquid}} \cdot L_f},$$

where

$$\rho_{\text{liquid}} = \text{density of water} = 1000 \text{ kg m}^{-3}$$

$$L_f = \text{latent heat of fusion} = 3.34 \times 10^2 \text{ kJ kg}^{-1}$$

which equals about 14 mm d$^{-1}$. Analysis of four storms that resulted in significant snow accumulation in heated plots illustrates large differences in accumulation amount and melt rate when compared to control plots (Table 1.2). Accumulation was always greater in the control plot by 12-73%. An evaluation of energy emitted from the heaters can explain the rate of snowmelt in the heated plots. Here we ignore natural energy balance parameters and the cold content of the snowpack. Knowing that the heaters provided $1.8 \times 10^7$ J m$^{-2}$ d$^{-1}$, and assuming that 100% of the 214 W m$^{-2}$ of energy emitted from the heaters was absorbed at the snow surface, the calculated amount of time to melt the snow ranged from 0.19 to 0.70 days (Table 1.2).

<table>
<thead>
<tr>
<th>Storm Date</th>
<th>Storm SWE (mm)</th>
<th>Accumulation control (mm)</th>
<th>Accumulation Heated (mm)</th>
<th>$Q$(melt) (kJ/m2/day)</th>
<th>Melt time (days)</th>
<th>Actual Melt time (days)</th>
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</thead>
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<tr>
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<td>17.8</td>
<td>24.7</td>
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<td>5.94E+03</td>
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<td>6.08</td>
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<td>11</td>
<td>9.6</td>
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<tr>
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<td>30</td>
<td>8.0</td>
<td>8.67E+03</td>
<td>0.48</td>
<td>1.83</td>
</tr>
</tbody>
</table>

Table 1.2: Four major storm events are listed with the associated amount of SWE, the amount of accumulation in control and heated plots, the energy ($Q_{\text{melt}}$) needed to melt each event, the time in days needed to melt each event from 100% of heaters energy, and the actual amount of time melt required based on snow depth measurements.
In all cases, more time was needed to melt the snow from each storm event than calculated from the heaters energy, suggesting that less than the theoretical energy from the heaters is absorbed by the snow surface. However, actual disappearance date was generally within 20% of the calculated date, suggesting that there was some transfer of energy from the heaters to the snowpack.

1.5.1.2 Soil temperature

Soil temperature in the control plots went below freezing in mid-December at all plots. Soil temperatures then remained below but near freezing until melt out about the first of June. When the snow was gone, soil temperature in the control plots then increased to about 10 °C within a week (Figure 1.9).
In contrast to the control plots, soil temperatures in the heated plots were almost always above the freezing point and varied widely with time. Soil temperatures after snowmelt were as much as 5 °C higher in heated plots versus control plots (Figure 1.9). In contrast to the control plots, soil temperature in the heated plots varied in response to snow events. For example, in plot 3H, snowfall beginning on day 79 accumulated to 20-30 cm in depth (Figure 1.10).
temperatures decreased from about +5 °C to about +1 °C. Over the next ten days with no new snowfall, soil temperatures in the heated plot then increased from +1 °C back to +5 °C.

Figure 1.10: LSA Plot 3 snow depths for heated (red line) and control (black line) beginning 5 March (upper panel). Lower Panel: Soil temperatures (left y-axis) for heated (red, solid line) and control (black, solid line), and soil moistures (right y-axis) for heated (red, dashed line) and control (black, dashed line) from 1 November to July 19.

1.6.1.3 Soil moisture

Similar to soil temperature, there were sharp contrasts in soil moisture between control and heated plots at the LSA (Figure 1.9). As mean daily air temperatures decreased below freezing in early November, soil moisture at all plots decreased from about 30% to 22%. In the
control plots soil moisture was relatively consistent at 22 to 24% prior to DOY 90. Starting on DOY 90 (first day of the year when the mean daily air temperature reached freezing soil moisture began to increase at all the control plots. The soil moisture in the control plots continued to increase over time until reaching a maximum at 30 to 32%, about 4 to 5 days before the snow was gone.

In contrast to the control plots at the LSA, soil moisture in the heated plots increased to near maximum values during the winter (Figure 1.9). Prior to the increase in the soil moisture of control plots on about DOY 90 (1 April), soil moisture in the heated plots was about 28-30%. Thus, soil moisture at this time was about 30% greater in the heated plots compared to the control plots (Figure 1.9). As the soil moisture began to increase in the control plots after infiltration of melting snow, there was little change in soil moisture of the heated plots (p<0.01). As soil moisture reached peak values in the control plots, it was declining in the heated plots. At three of the treatments, maximum soil moisture was higher in the control plots when compared to the heated plots. Soil moisture in the heated plots responded to new snowfalls. For example, a large snowfall in early January resulted in a large increase in soil moisture at plot 3H (Figure 1.10). A following dry spell caused a large decline in soil moisture at plot 3H.

1.5.2 USA-ALP/Saddle site

1.5.2.1 Climate

The USA and ALP experienced much colder temperatures than LSA, with mean daily temperatures of -10 to -15°C from November until the end of February (Figure 1.3). March and April were slightly warmer with mean daily air temperatures around -5°C. Air temperatures experienced a spike from -5 to 1°C on 14 May and continued to steadily increase to 10°C for the
remainder of May and June. The exposed environment of the USA and ALP sites experienced wind speeds averaging 10-15 m s\(^{-1}\) from November to the end of February (Figure 1.3). Mean daily wind speeds decreased to around 7.5 m s\(^{-1}\) for March through June. Winds at the ALP site were consistently higher (by \(~2\text{-}5\text{ m s}^{-1}\) than the USA.

Snow began accumulating at the Saddle in late October, reaching a maximum SWE of 790 mm in late May, with the last snow melting on about 15 June (Figure 1.11).
Figure 1.11: Upper panel. Daily SWE at Saddle is graphed as a solid black line. Bottom panel. Evolution of Saddle snowpack represented by weekly snowpit profiles. Four grains types are represented according to the legend in the top right corner. Mean snow temperature (red) and mean density (blue) is located above each stratigraphic profile.
The seasonal maximum SWE was within 11% of the 15-year mean value of 892 mm. Early to mid-winter at the USA and ALP for 2009-2010 was characterized by small but frequent storm events, with several large storm events occurring at the end of March through the beginning of May. Snow depths at the NWTLTER index pit at the Saddle increased steadily to a maximum of 178 cm on 11 May at which point mean daily air temperatures remained consistently above the freezing point (Figure 1.11). At the base of the snowpack a well-developed layer of facets about 40-50 cm in thickness persisted throughout the winter (Figure 1.11). Several wind crusts formed throughout the winter due to high winds. Towards the end of April numerous crusts and ice layers formed as a result of warmer temperatures. At the end of May, the snowpack layers began to decompose predominately equi-temperature and wet metamorphism grains until the end of melt. The snowpack became isothermal around 19 May coincident with continuous melt (45 mm day⁻¹), resulting in the disappearance of snow by 16 June.

On 7-8 May, snowpits were excavated in ALP control plots 70 and 76 located in the southeast portion of the plot about one-third and two-thirds downslope of the upper most plot in the site, respectively. Results illustrate the presence of a large melt-freeze crust 36 cm in thickness at the base of the snow pack (Figure 1.12). On top of this crust were several ice and crust layers resulting from warm days and cold nights during the spring similar to those formed at the Saddle index pit. On the contrary, a snowpit excavated approximately 20 cm from the heated plot of ALP Plot 70 showed no basal crust layer (Figure 1.12). Instead the bottom 55 cm was composed of rounded, compact equi-temperature snow grains ranging in diameter from 0.5-0.75 mm. Above the basal layer there were three very prominent, thick ice layers 4 to 15 cm in thickness. The layers were so solid and extensive that density was unable to be taken for the depth 60-90 cm. In between these ice layers were two melt-freeze crusts that were slightly less
consolidated and not as hard as those observed in the ALP control snowpits. Observations of other USA and ALP plots also found a greater number and thickness of ice layers in heated plots relative to control plots.

Figure 1.12: Snowpit profiles of ALP Plot 70h (located 20cm from Plot 70c), 70c, and 76c.
Manual measurements of snow depth in the control and heated plots of the USA illustrate a rapid accumulation of snow reaching depths of 100-150 cm by mid-November (exception plot 22h) (Figure 1.13). Snow depths continued to increase through February, then remained relatively constant from the end of February through mid-April. Snow depths in the heated plots followed similar trends to controls but were consistently lower (25-70 cm) than the control plots until mid-April.

Unlike the USA, which lies in a snow accumulation zone, snow depths in the ALP control and heated plots followed similar patterns to the Saddle index pit (Figure 1.11). There was a spatial trend in snow depth, with snow depths decreasing from the northwest to southeast. Snow depths in ALP plots ranged from 0 to 80 cm at the beginning of November and increased steadily until maximum accumulation (Figure 1.14). Snow depths in heated and control plots followed each other closely, with heated plots in the lower portion of the site accumulating less snow (~20-25 cm) than control plots.
Figure 1.13: Manual snow depth measurements of USA plots. Measurements were approximately bi-monthly, beginning on 6 November.

Figure 1.14: Manual snow depth measurements of ALP plots. Measurements were approximately bi-monthly, beginning on 6 November.
Automated hourly measurements of snow depth at the USA began in mid-April. There was some spatial variability among plots with maximum snow depths in control plots ranging from 170 to 240 cm (mean of 213 cm, n = 3) (Table 1.1). In USA heated plots, snow depths fluctuated throughout April and May with small periods of accumulation and ablation of approximately ~20-25 cm (Figure 1.1). A large decrease in snow depth of 75-100 cm occurred around 23 April. During this time air temperatures were below freezing and winds were calm (<5 m s⁻¹). Within 5 days, winds increased up to 10 m s⁻¹ and snow depth increased to around previous levels. Maximum snow depths in heated plots ranged from 162 to 220 cm (mean of 197, n=3) (Table 1.1). Despite similar snow depths, snow in heated plots disappeared more rapidly than controls becoming snow free 5 to 18 days earlier.
Figure 1.15: Snow depth measurements for USA control and heated plots. Measurements began on 16 April 2010.

Automated hourly measurements of snow depth at the ALP plots began at the end of January. Snow depths in heated and control plots followed the pattern of the saddle index snowpit, reaching maximum snow depths in late May (Figure 1.11). Snow depths decreased from the northwest to southeast of the site with maximum snow depths in control plots ranging from 88 to 240 cm (mean of 182, n=5) and maximum snow depths in heated plots ranging from
80 to 240 cm (mean=177 cm, n=5) (Table 1.1). As with the Saddle index pit, snow depths increased steadily until mid-May when snowmelt began and snow in control plot disappeared two and half weeks to one month later (Figure 1.16). Despite similar snow depths in heated and control plots, in ALP heated plots snow disappeared 0 to 18 days earlier than control plots (Table 1.1). The difference in timing of snow disappearance between heated and control plots at the ALP decreased with decreasing snow accumulation. For example, heated plot 42 reached a maximum accumulation of 240 cm and melted out 18 days prior to control plot 42 which reached a maximum accumulation of 235 cm. Whereas, heated plot 75 reached a maximum accumulation of 80 cm and became snow free less than 24 hours prior to control plot 75 which reached a maximum accumulation of 88 cm. Because manual measurements were made at the edge of plots and automated measurements were made at the center of plots, some discrepancies are apparent between snow depth measurements.
Figure 1.16: Snow depth measurements for ALP control and heated plots. Measurements began on 31 January 2010.

Once the snow depth exceeded the height of the heaters (1.2 m) at the USA and ALP heated plots, caves began to form around the heaters. As early as November in the USA site, each cave was approximately 1 m$^3$ in volume, or together 56.6% of the plot volume when snow depth was 1.5 m. (Assuming a snow depth of 1.5 m, plot volume is 10.6 m$^3$ and cave volume is 6 m$^3$). These caves grew in size and extended beyond the heated structure, but beneath the snow
surface. This created the illusion of a solid snowpack, when in actuality the snowpack was hollow underneath a very thin snow surface layer. The cave walls were vertical. As the caves grew, the bottom 50 cm of the cave walls began to melt forming tunnels that connected caves within a plot to each other (Figure 1.17).

At the snow surface, above the heaters, holes (i.e. entrances to the caves) formed (Figure 1.18).

Figure 1.17: Cross-section of heated plots illustrating cave morphology. Caves formed vertical walls above and below heaters. Caves in some plots were connected at the bottom.

Figure 1.18: Photograph of holes that formed above the heaters. You can see the hollowness termed “caves”, through the holes.

Initial snow surface hole area for ALP and USA was approximately .1-.2 m² representing 12% of the plot surface area. Throughout the winter, intermittent surface holes covered
approximately 3 m², 16-43% of plot surface area. By the end of May, the holes covered approximately 80% of the plot’s surface. During storms or windy events (heaters turned off in wind speeds >15 m s⁻¹), holes disappeared as snow re-sintered across the holes creating a “snow cap”. Occasional field observations found the thickness of “snow caps” to vary between 5 and 35 centimeters depending on the amount of snowfall (or redistribution of snow) and how long the heaters were off. Once heater’s turned back on, holes quickly reappeared (< less than 24 hours in some situations).

Subsidence of the snow surface occurred in the heated plots due to caves below the snow surface. New snow and redistribution of snow by wind would refill sunken plots, causing a subsequent increase in snow depth. One extreme example of this process was noticed on 23 April when snow depth decreased by 75-100 cm. Within 5 days, snow depth increased to slightly higher than previous levels (Figure 1.15). Around mid-May when steady snowmelt began in all plots, subsidence of the snow surface accelerated in heated plots at the USA and at the end of May at the ALP, eventually causing the snow surface to completely collapse. Caves never grew large enough to fully penetrate the center of the plots, leaving a solid center core of snow that was last to melt (Figure 1.19). Pillars at the USA were smaller than those at the ALP. Located at the most southeast portion of the ALP site, snow depths in heated plots 70 and 75 never reached the height of the heaters, thus no caves were formed.
When snow depth was less than 120 cm, a layer of surface hoar was evident on the snow surface beneath the heaters, same in morphology and location as observed at the LSA. On top of the surface hoar layer, a melt-freeze crust would form, followed by another layer of surface hoar. In plots with snow depths greater than 120 cm, large surface hoar crystals formed inside caves along the cave walls. Occasional field observations found a melt-freeze crust on top of surface hoar where another layer of surface hoar would form. The melt-freeze crusts suggest that the heaters caused melt on the snow surface, as well as inside the caves beneath the heaters.

1.5.2.2 Soil temperature

Soil temperature in USA control plots remained around 0 °C from November until snow disappearance around mid-June (Figure 1.20). Post snowmelt soil temperatures increased to around 12 °C. In contrast to the control plots, soil temperatures in USA heated plots varied in response to snow depth (except plot 22), fluctuating between 0 and 5 °C throughout the season until the end of May. For example, in heated plot 21, snowfall on DOY 130 accumulated 50 cm
of snow. Soil temperatures increased from 0 to 4 °C. Over the next few days, about 30 cm of snow ablated and snow temperatures decreased to around 2 °C (Figure 1.21). After snow disappeared temperatures in all heated plots increased from about 3 °C to 13 °C. Soil temperatures post snowmelt remained higher by 2-3 °C in USA heated plots.

Figure 1.20: Soil temperature and moisture for all USA sites from Day of Year 300 (2009) to 200 (2010).
Soil temperatures in ALP control plots went well below freezing in mid-December (Figure 1.22). Soil temperatures then remained below 0 °C until mid-April when warmer air temperatures caused soil temperatures to remain around 0 °C until melt out. Once snow was gone, soil temperatures increased to about 13 °C to 15 °C within a week. Similar to the control plots, soil temperatures in ALP heated plots went below freezing in mid-December. Then around the end of February, soil temperatures in the three upper plots increased to 0 °C coincident with snow depths reaching 100-150 cm (Figure 1.23). The two lower heated plots followed soil temperatures in control plots very closely, reaching 0 °C in mid-April coincident
with warmer air temperatures (Figure 1.2). Soil temperatures after snowmelt were 3-5 °C higher in the three upper plots versus control plots and 1-2 °C higher in the two lower heated plots compared to control plots.

Figure 1.22: Soil temperature and moisture for all ALP sites from Day of Year 300 (2009) to 200 (2010).
Figure 1.23: ALP Plot 42 snow depths for heated (red line) and control (black line) beginning 30 January (upper panel). Lower Panel: Soil temperatures (left y-axis) for heated (red, solid line) and control (black, solid line), and soil moistures (right y-axis) for heated (red, dashed line) and control (black, dashed line) from 1 November to July 19.

1.5.2.3 Soil Moisture

There were large differences in soil moisture between heated and control plots at the USA and ALP. As mean daily air temperatures decreased below freezing in early November, soil moisture in all USA and ALP control plots decreased from 30% to 26%. Soil moisture remained relatively consistent around 22-26% prior to DOY 110 (20 April) (Figure 1.22, 1.23). Starting on DOY 110, coincident with mean daily air temperatures reaching freezing, soil
moisture began to increase. Soil moisture in control plots continued to increase over time reaching a maximum of 31% to 39%. Maximum soil moisture in USA control plots varied greatly occurring 12 days before snow disappearance in one plot, coincident with snow disappearance in another, and 8 days after snow disappearance in the third. At the ALP, maximum soil moisture in control plots occurred 0 to 2 days after snow was gone.

In contrast to control plots, soil moisture in the USA heated plots showed much less variation with time. Beginning in November soil moisture in heated plots was relatively consistent at 32-35% (except plot 22)(Figure 1.23). Plot 22h remained around 26% until 14 February when soil moisture began to increase steadily. As the soil moisture in the control plots began increasing after the infiltration of melting snow, there was little difference within the heated plots (ANOVA, p<.001). Thus, soil moisture was 6 to 9% greater in heated plots than control plots during this time. As soil moisture reached peak values in the control plots, it was declining in all of the heated plots. In two heated plots maximum soil moisture occurred the same day as snow disappearance, whereas maximum soil moisture in 21h occurred 10 days prior to snow disappearance. Maximum soil moisture in all of the heated plots was higher than control plots but post snowmelt soil moisture values were higher in control plots versus heated plots.

In contrast to the USA, soil moisture in the ALP heated plots showed trends related to the amount of snow accumulation. Soil moisture in all ALP heated plots followed similar patterns as the controls remaining around 22-24% from mid-November to the end of February. At the end of February, as soil temperatures were increasing in the three upper plots, soil moisture began to steadily increase (Figure 1.23). Similar to control plots, soil moisture in the two lower plots began increasing around 20 April. Contrary to control plots, soil moisture values at this time were higher than control plots at 30 to 35% (p<0.01). Maximum soil moisture in four of the
plots was higher than control plots at 36 to 39%. Thus, total soil moisture was greater in heated plots compared to control plots. Maximum soil moisture occurred within five days of snow disappearance. Post snowmelt soil moisture values were similar between all heated and control plots.
1.6 Discussion

1.6.1 LSA/C1 site

1.6.1.1 Climate

Measurements of snow depth and snow properties illustrate large differences between heated and control plots at LSA. Control plots resembled average snowpack conditions compared to the C1 index snowpit. In contrast, the heaters provided enough energy to prevent accumulation of snow in heated plots except for 4 to 5 major storms. Even then, the accumulation never exceeded 45 cm and quickly melted, suggesting that a substantial amount of energy was reaching the surface of snow/soil throughout the winter. A consequence of global warming is the elevational migration of the snow/rain line (IPCC, 2007). With a greater proportion of precipitation falling as rain instead of snow, snowpacks at mid-elevations may be smaller and more intermittent (Stewart, 2009). Thus, the ephemeral snowpack produced by heaters in this experiment may provide insight into the hydrological and ecological responses to a climate characterized by warmer air temperatures at low elevations experiencing a greater rain to snow precipitation ratio. With more water available throughout the winter and a dampened spring snowmelt pulse, ecosystems may experience large amounts of stress by early summer.

When accumulation of snow did occur in heated plots, there was more accumulation in the center of the plots versus under the heaters implying an uneven distribution of energy across each plot. Beneath the heaters, the presence of very large surface hoar crystals often covered the snow surface. Most likely energy from the heaters created much warmer air over the heated plots allowing the air to hold more water vapor. At night, longwave radiation is lost at the surface and air temperatures drop, causing the air above the snowpack to reach water saturation. The water vapor then condenses on the snow surface, creating large, feathery snow crystals
characteristic of surface hoar. Surface hoar formation is sensitive to wind velocity (Hachikubo and Akitaya, 1997), and growth sharply declines when wind speeds exceed 3 m s\(^{-1}\) (Fohn, 2001). Because the heavily forested site was protected from high winds, the surface hoar crystals were able to grow quite large in the heated plots. Global warming will cause absolute humidity to increase, while relative humidity is to remain constant (IPCC, 2007). Contrary to the atmosphere, the absolute humidity of the snow surface cannot increase as it is limited by a temperature of 0 °C. Thus, vapor pressure gradients between the air and snow surface may increase as a consequence of climate change, altering energy and mass fluxes. In response to the altering of energy and mass fluxes, this experiment indicates greater mass gains to the snowpack via sublimation (e.g. deposition of surface hoar crystals) relative to control plots.

Additionally, energy from heaters generated melt, causing a melt-freeze crust to form over the surface hoar crystals. Subsequent decreasing temperatures caused the process to repeat itself resulting in another layer of surface hoar crystals on the snow surface. It is likely that global warming will increase the occurrence of melt-freeze crusts as temperatures are more likely to increase above freezing during the day, then dropping below freezing again at night. In deep snowpacks, melt-freeze crusts delay the transmission of meltwater through a snowpack (Gerdel 1954; Waldner, 2004) and tend to route water laterally causing a greater delay between snowmelt and streamflow response (McGurk and Marsh, 1995). However, in shallow or intermittent snowpacks representative of the LSA site, consequences of melt-freeze crusts on delay of water transmission are likely smaller.
1.6.1.2 Soil Temperature

It is well established that soil dynamics are influenced greatly by snow cover depth, duration, and snowmelt timing and magnitude (Walker et al., 1994; Harte et al., 1995; Filipia et al., 2009, Molotch et al., 2009). Expected relationships between soil temperature and snow depth are observed in LSA control plots with soil temperatures remaining below 0°C until snow depth reaches ~120 cm and remaining at 0°C until snowmelt and infiltration begins as a result of the insulating effect of snow (Harte et al., 1995; Filipia et al., 2009). On the contrary, heated plots experienced an ephemeral snowpack and periodic snowmelt during the winter that resulted in warmer soil temperatures that varied in response to snow events. When snow accumulated, less heat was able to reach the soil resulting in decreased soil temperatures. Snow quickly melted upon reaching an above 0°C soil surface causing soil moisture to increase from the infiltration of snowmelt. Once snow disappeared, the energy that was being used for melting the snow switched to evaporation of the wet soils. In response, soil moisture began to decrease and more energy became available to warm the soils causing soil temperatures to increase. Soil temperatures continued to rise until the next snow event limited exposure to heater’s energy decreasing soil temperatures once again. This cycle is prevalent throughout the winter in heated plots resulting in warmer and wetter soils relative to control plots.

Our results contradict the “colder soils in a warmer world” predicted by Groffman et al. (2001). Global warming is expected to cause decreasing snow cover and snow accumulation to begin later in the year (IPCC, 2007). A number of studies have found that a declining snow cover results in colder soil temperatures by decreasing the insulating effects provided by snow (Groffman et al., 2001; Hardy et al., 2001; Molotch et al., 2009). For example, Molotch et al. (2009) found a snowpack in New Mexico exposed to shallower snow depths and warmer air
temperatures to have significantly colder soil temperatures than a Colorado snowpack that was deeper but exposed to colder air temperatures. Another consequence of shallower snowpacks is a greater frequency of soil freeze/thaw cycles and deeper soil frost (Hardy et al., 2001). Because our plots melted completely, both short and longwave radiation was able to reach the soil surface (which has a low albedo) and thus raise soil temperatures. This may differ from a situation where snowpack is shallower, with a relatively high albedo and little transmission of SW and LW to the soil. Therefore, while changes in snow cover may counteract the effect of global warming on the soil characteristics most relevant to winter survival (i.e. minimum and mean soil temperature) (Kreyling, 2010), our experiment produces the opposite.

Other snow manipulation studies produced warmer and drier soil conditions post snowmelt (Harte and Shaw, 1995; Harte et al., 1995; Price and Wasser, 2000; Dunne et al., 2004; Adler et al., 2007), however, few studies have observed soil dynamics during the winter. Overall, there remains some uncertainty in how minimum soil temperatures will be affected (increase or decrease) with climate change because it depends on the complex interaction between rising mean air temperatures (almost certain to increase), air temperature variability (will increase), and insulation by snow (expected to decrease) (IPCC, 2007; Kreyling, 2010). At the LSA there was no evidence of soil frost under the experimental heating and in fact, soil temperatures never dipped below $0^\circ$ C. Thus, contrary to other studies who have shown increased root mortality and loss of nutrients due to soil freezing in a warmer world (Hardy et al., 2001), our results suggest ecosystems may experience increased nutrient availability and uptake.
1.6.1.3 Soil Moisture

Interestingly, the range of soil moisture observed in this study is much less than other studies. During the winter, soil moisture values fluctuated only a few %, with the largest difference between pre-melt moisture and maximum moisture of 13%. A study performed in a subalpine meadow on Niwot Ridge illustrated soil moisture to fluctuate between 8 and 28% during the winter at soil depths of 10 cm. Filippa et al. (2009) found soil moisture values at an integrated depth of 0-30 cm to be 25% during the winter, increasing to 60% during snowmelt in the same meadow for 2006 and 2007. This could be explained by a lack of calibration, the type of sensor, the depth in which it was placed, or the rocky soil in which they were placed. Calibration of the soil moisture sensors is needed and is currently underway. Despite the lack of calibration, the relative values provide useful information on soil moisture in heated and control plots.

Significant differences are apparent in soil moisture amounts and trends between heated and control plots. Timing of snowmelt infiltration onset and peak soil moisture is largely dependent on winter season snow accumulation amounts and average winter air temperature, which controls the snowpack cold content (Molotch et al., 2009). In control plots soil moisture begins to increase once air temperatures increase above 0°C reaching maximum soil moisture a few days before snow disappearance. Thus, infiltration rate was limited by field capacity, not snowmelt rate, indicating overland flow.

As snowmelt timing shifts to earlier in the season and maximum SWE becomes smaller in magnitude with global warming, maximum soil moisture will likely be smaller in magnitude and occur earlier. On the contrary, our results indicate greater total moisture throughout the winter in heated plots compared to controls and soil moisture values that did not experience a
true maximum. Energy available to melt was much greater in heated plots than control plots, quickly melting snow within plots and from outside plots. Infiltration occurred throughout the winter maintaining high soil moisture values relative to control plots. Increases in soil moisture following periodic snowmelt events never reached values as high as maximum soil moisture in control plots indicating that field capacity was never fulfilled.

Our results contradict those of other studies that found consistently lower soil moisture during winter but higher maximum soil moisture during snowmelt in heated plots (Adler et al., 2007; Hardy et al., 2001, and Zhao and Gray, 1999). These findings have been attributed to greater extents and deeper amount of soil frost caused by shallower snowpacks (Hardy et al., 2001; Zhao and Gray, 1999). Soil frost decreases permeability, preventing infiltration of snowmelt, thus promoting overland flow, and faster stream response. Hardy et al. (2001) found approximately 12-22% less water infiltrated in treatment versus reference plots. As a consequence, greater soil frost may lead to less groundwater recharge (Stradler et al., 1996). Conversely, Campbell et al. (2010) found climate warming to have negligible effects on soil frost depth and soil freeze/thaw cycles. The most important change found was a shortening of the period of frozen ground coincident with a shorter period of snow cover. At the LSA, heaters prevented soil frost allowing infiltration to occur throughout the winter, preventing overland flow, and thus promoting greater groundwater recharge. Because snow was likely melting from areas adjacent to the heated plots, but routed laterally, heated plots may have received more total SWE relative to control plots providing ecosystems with greater water and nutrient availability and uptake during the winter, but likely leaving systems stressed later in the season. The relationship between soil dynamics, snow, and climate change is complex. Our experiment produces soil conditions more representative of what may result from intermittent or shallow
snowpacks with air temperatures that prevent significant soil frost. Additionally, because our experiment caused snowmelt to occur so quickly upon hitting above 0\(^\circ\) C soils, soil dynamics may be comparable to what may happen if the snow-to-rain ratio decreases in subalpine environments. Precipitation as rain is not stored during the winter as snow is, but becomes available immediately to the environment by either infiltration or overland flow directly to streams.

1.6.2 USA-ALP/Saddle

1.6.2.1 Site Evolution

Experimental warming resulted in significant differences in snowpack evolution between heated and control plots at the USA and ALP, as well as between the USA and ALP sites. These differences can be partially explained by the thermal regime of the heated plots where caves formed once snow depth exceeded the height of the heaters. Once snow covers the heaters, energy was efficiently delivered from the heaters to the snowpack, warming that portion of the snowpack. Because snow has a low thermal conductivity (Male and Gray, 1981), that energy cannot be efficiently transferred to colder areas of the snowpack. This high flux of energy fills the cold content of the snowpack, resulting in the formation of liquid water. Once the field capacity of the snowpack is filled, the liquid water percolates downward. Once liquid water reaches colder snow, it refreezes generating heat from the latent heat of fusion, effectively warming the snow below creating a positive feedback effect and thus, explaining the greater frequency of ice lens noticed in heated plots compared to control plots. Similarly, ice columns have been reported to form on Niwot Ridge as a result of the release of latent heat associated with the freezing of greater amounts of meltwater (Williams et al., 2000). Ice layers possess a
low permeability that slows the rate of vertical water movement causing ponding of water above
or by routing water in various other directions (Gerdel, 1949). In deep snowpacks, ice layers act
to delay the transmission of meltwater through a snowpack (Gerdel 1954; Waldner, 2004) and
tend to route water laterally (McGurk and Marsh, 1995; Waldner et al., 2004), as well as permit
greater water storage in the heated plots due to the capillary tension between the ice layer and
liquid water (Singh et al., 1999).

The low thermal conductivity of snow causes heat transfer to be most efficient in the
vertical direction, towards the ground and above the heaters thus explaining the uneven
distribution of snow in the heated plots. Kimball et al. (2008) found that ~1/6 of the heater’s
energy reaches the center of the plot. Similarly, our results illustrate less heat reaching the center
of plots by caves that represent over 50% of the plots volume (and lack of mass), as well as the
remnants of pillars during melt. Because snow exceeded the height at the USA site several
months earlier than the ALP, caves formed earlier and were larger. At the USA, bare soil, liquid
water, and flowers were apparent in USA heated plots in February, while caves were just
beginning to form at the three upper ALP plots. Pillars that formed towards the end of snowmelt
were larger at the ALP site than the USA, also an artifact of larger caves that penetrated much
further into the plot.

When surface holes above the heaters are present, less than 100% of the heater’s energy
is absorbed. Surface holes were found to cover 12-43% of the snow surface area throughout
mid-winter, and up to 80% in May, and quickly reappeared after storms (<24hrs). During
blowing snow events, surface holes acted as preferential collectors of snow that replenished
snow that had melted and filled in surface holes. The frequency of surface holes and blowing
snow events suggests that heated plots at ALP and USA may have a significantly higher total SWE than control plots.

Furthermore, several stratigraphic differences resulting from cave morphology were apparent between heated and control plots. The same layers of surface hoar and melt-freeze crusts seen in LSA heated plots were apparent on the snow surface beneath heaters at the USA and ALP. Surface hoar was also found inside caves along cave walls despite high winds. Because wind speeds decrease inside the caves, rates of sublimation (and the subsequent release of latent heat of vaporization) are apparently very high, resulting in ideal conditions for surface hoar formation. Large amounts of heat and water vapor build up inside the caves. At night or in high winds when heaters turn off, cooler temperatures cause deposition of surface hoar grains. Heaters melt the snow grains during the day or when the presence of “snow caps” permit 100% heater efficiency, and refreeze at night or when heaters are turned off creating a melt-freeze crust. While it is common to find surface hoar crystals 1 to 3mm in size, grains in this study grew larger than 10mm suggesting large vapor gradients as a result of the extra energy input from the heaters. These results suggest a greater energy and mass flux to the snowpack via latent heat (sublimation) in heated plots compared to control plots.

Early snowfall created a large layer of TG snow at the Saddle and basal crust at C1 that persisted throughout the season. Isotopic signatures support that this basal layer at the Saddle was persistent throughout the winter (i.e., no fractionation via melt or freezing altered this layer until melt began), whereas, layers above experience varying O18 values representing fractionation (Cowie, 2010). This depth hoar layer metamorphosed into a melt-freeze crust sometime between May 4th and May 11th. The crust layer was very pronounced at the Saddle index snowpit as well as a snowpit excavated in ALP control plot 70 on 7 May. However, this
basal crust was not present in the snowpit of heated plot 70. The bottom 55cm of the heated plot consisted of rounded equi-temperature grains, likely an influence of the heaters, which caused warmer ground temperatures that degraded or prevented any basal ice or crust layers.

1.6.2.2 Climate

Snow properties in control plots at the USA and ALP are comparable to average conditions recorded on Niwot Ridge. The USA experiences greater spatial variability of snow depth than the ALP due to differences in the topography, vegetation, and patterns of wind redistribution characterizing each site. Wind sheltering has a large affect on snow depth (Erickson et al., 2005). The USA site has a more rugged topography surrounded by krummholz which acts to shelter wind creating large snow drifts. Litaor et al. (2008) found that 85% of variability in snow disappearance rate was related to terrain characteristics (e.g. location in a krummholz zone).

In heated plots at the USA and ALP, differences in snow depth and stratigraphy when compared to control plots can be partially explained by the formation of caves, an artifact of the fixed heater heights. The earlier caves formed, the larger the differences between heated and control plots. In heated plots at the USA, snow depths were lower and more variable compared to control plots due to the lack of a solid snowpack created by the caves. Subsidence occurred from a lack of structural integrity, thus causing a decrease in snow depth. Subsidence also likely occurred from generation of liquid water at the snow surface, especially in the spring due to higher solar radiation. Gerdel (1954) describes dendritic patterns on snow surface due to subsidence representing the presence of underlying flow channels which is due to coarser grains formed from wet snow metamorphism (Higuchi and Tanaka, 1982). Thus, subsidence from
caves act as preferential flowpaths for the liquid water creating a positive feedback effect. Once a storm or windy event occurs, the sunken plots fill in and snow depth increases. On the contrary, at the ALP, subsidence was not as evident because the caves formed later in the season. Snow depths between heated and control plots followed each other closely.

Overall, melt occurred earlier in heated plots compared to control at both USA and ALP sites. The number of days between snow disappearance in heated and control plots appears to be related to timing and magnitude of snow accumulation and thus, cave formation. The larger the caves, the earlier heated plots melted relative to control plots. Other studies have found heating to advance melting by 4.7-14 days (Adler et al., 2007; Dunne et al., 2004; Harte et al., 1995; Price and Waser, 2000). Our results exceed the upper and lower bounds of this range due to larger range of maximum snow depths and differences in heater energy flux. Caves caused a large collapse of the snow in heated plots, exposing more surface area to solar radiation and the heater’s energy resulting in an extremely rapid melt despite snow depths similar to control plots. This explains why the heated plot with the most accumulation (ALP plot 42, 240 cm) melted before all other heated plots at the USA and ALP sites. Similarly, in ALP plots 70 and 75 there was less than one day of difference in when heated plot melted relative to the control because caves never formed and the mass between the two plots were similar. Additionally, depressions at the snow surface in these cave-free plots may have been refilled by wind redistribution of snow maintaining similar masses between heated and control pairs.

While precipitation trends are unclear in climate modeling scenarios, it is well known that climate change is expected to cause earlier snowmelt as a result of warmer spring temperatures (IPCC, 2007). Many alpine environments have not yet experienced a significant change in the quantity of winter precipitation or snow depth, but some are seeing an earlier melt-out (Mote et
The role of snow in response to changes in climate is complex (Houghton et al., 2001; Williams et al., 2003). In the western/southwestern United States, models predict declining spring snow cover extent and snow depth but greater winter snow cover extent as a result of increasing occurrence of thaw events during winter and early spring (Dyer and Mote, 2007). Williams et al. (1996) suggests that warming at lower elevations may increase precipitation above treeline through a positive feedback loop by advection of increased water vapor to higher elevations and increased orographic precipitation as snow. Late-persisting snow on the ground may provide the moisture for increased cloud cover and decreased incoming shortwave radiation which causes atmospheric cooling, lower saturation vapor pressure and increases in precipitation. Our experiment illustrates lower snow depths at treeline, little change in the alpine, but a greater overall SWE. While these results are partially an artifact of experimental design, the ecological and hydrological response may be similar to that experienced in a warmer climate experiencing a greater mid-winter snowpack, but earlier runoff. An earlier runoff will likely reduce summer and fall flows (Mote et al., 2005) leaving ecosystems water and nutrient limited.

1.6.2.3 Soil Temperature

As expected, soil temperatures in control plots at the USA and ALP were controlled by snow depth with soil temperatures remaining below 0°C until snow depths reached insulating levels. Due to this relationship, soil temperatures at the USA reached ~0°C earlier than ALP control plots. In the heated plots, snow depth and duration indirectly influenced soil temperatures due to the formation of caves. Because the USA and ALP snowpack evolved slightly differently, there were some distinct differences observed in the soil dynamics between the two sites. Additionally, results of site evolution indicate that soil temperature and moisture
values in heated plots may have been underestimated due to the uneven distribution of energy from the heaters and the location of soil temperature and moisture sensors.

In heated plots, the earlier caves formed, the greater the difference in soil temperature between heated and control plots. The hollowness associated with caves prohibited the insulation provided by a solid snowpack and instead created more interaction with the atmosphere (and heater’s energy). Thus, when surface holes above the heaters are present, less than 100% of the heater’s energy is absorbed. During these time periods, more heat is lost to the atmosphere, as opposed to being trapped inside the caves, resulting in decreased soil temperatures. On the contrary, when surface holes were covered, generally after snow events or snow redistribution by wind, little heat could escape allowing a large portion of the longwave radiation being emitted by the heaters to be absorbed by the snowpack.

In USA heated plots (except 22), soil temperatures were highly variable compared to control plots and ALP heated plots due to the intermittent existence of surface holes above the heaters. Soil temperatures never dropped below 0°C suggesting that the heater’s provided enough energy that even when holes were present and air temperatures were sub-freezing, soil temperatures were able to remain at 0°C or higher. In the three upper ALP plots and USA plot 22, the influence of caves and surface holes were not as prevalent, and soil temperatures followed control plots remaining around 0°C until air temperatures began increasing at the end of April. Molotch et al. (2009) found that the timing of soil warming is related to the snowpack cold content with longer delays in warming corresponding to deeper snowpack and colder winter air temperatures. Because heated plots had less mass and more energy, soil temperatures reached above freezing temperatures earlier than the controls.
Few alpine regions have been studied with respect to winter climate change, and those that have found strong effects of winter climate change on species ranges, species compositions, phenology, or frost injury. However, most of these studies have occurred in subalpine environments (Kreyling, 2010). Increasing temperatures have had little or no effect on snow accumulation and melt at higher elevations in climate warming research because air temperatures remain well below freezing during the winter (Stewart, 2009). Thus, it would be expected that soil dynamics in alpine environments would be little affected during the winter because energy deficits remain large. In response, the feedbacks between snow depth and soil temperature will not change much relative to current climatic conditions, as supported by soil temperature results at the ALP. Whereas, soil dynamics in environments at treeline, such as the USA, may experience a greater response to increasing air temperatures as they are located at slightly lower elevations decreasing the energy deficit of the snowpack. As a result, mid-winter melt may occur which would increase soil moisture, thus altering biogeochemical processes. However, soil temperatures would not increase above 0 °C as 40 cm of snow is enough to decouple soil and air temperatures (Strum et al., 1997). At the USA, soil temperatures that fluctuate above 0 °C soil are an artifact of the experiment and are not likely in a warmer world because heat cannot be transferred through a deep snowpack. Just as snow acts as an insulator, it would prohibit interaction with a warmer atmosphere.

Whether precipitation increases or decreases with climate change, warmer spring temperatures will likely result in earlier snowmelt which has been shown to result in warmer and drier soils (e.g. Harte and Shaw, 1995; Harte et al., 1995; Price and Wasser, 2000; Dunne et al., 2004; Adler et al., 2007). Similarly, post snowmelt soil temperatures in most USA and ALP heated plots were atleast 3 °C higher than the controls. Post snowmelt temperatures were only 1
to 2°C warmer in heated plots with snow depths consistently less than 150 cm (USA plot 22, and ALP plots 70 and 75) due to less heat flux to the soil during the winter from lack of insulation from caves and less efficient delivery of heater’s energy. Thus, our results suggest that at high elevations with deep snowpacks, the greatest influence of increasing air temperatures on soil temperatures occurs during and post snowmelt once snow depths are low enough to no longer prevent interaction with the atmosphere.

1.6.2.4 Soil Moisture

Soil moisture trends in control plots are similar to other studies of soil moisture at Niwot Ridge, Colorado (Filippa et al. 2009; Molotch et al., 2009), remaining relatively low during the winter, increasing once melt began. On the contrary, in heated plots soil moisture patterns are influenced by cave morphology and the thermal regime associated with heater’s energy. In USA heated plots soil moisture was higher throughout the season but reached maximum soil moisture of similar timing and magnitude to control plots. Three processes were likely responsible for high mid-winter soil moisture values in heated plots: 1) the addition of snow through surface holes that melted upon hitting an above 0°C soil, 2) lateral routing of melt water from outside plots, and 3) melting within the plots themselves due to heaters. Because heaters became less efficient once snow depth decreased below 150 cm, snowmelt rates became similar between heated and control plots towards the end of melt and thus, maximum soil moisture values were similar between USA heated and control plots.

At the ALP, little difference was observed between heated and control soil moisture values throughout the winter until soil temperatures rose above 0°C due to insulation from well-developed caves. Once this occurred, the same three mechanisms responsible for high soil
moisture in USA heated plots began to also increase soil moisture in the three upper ALP heated plots. The two lower plots experienced higher soil moisture from processes 2 and 3 (lateral routing of melt water from outside plots, and melting within the plots themselves due to heaters). Contrary to the LSA and USA, maximum soil moisture values were higher in all ALP heated plots (except plot 52) relative to control plots because less melt occurred mid-winter compared to the USA. While, both USA and ALP had greater total soil moisture during the snow cover season, the USA experienced higher soil moisture values during the winter permitting the growth of flowers, but lower post snowmelt values, whereas the ALP experienced higher post snowmelt soil moisture. Thus, at treeline ecological systems may become water-limited early in the growing season, whereas in the alpine water availability may sufficiently support these systems.

There was not a strong relationship between maximum soil moisture and snow disappearance date at the USA and upper two plots of the ALP. The wind driven redistribution of snow based on topographic position significantly affects soil moisture (Litaor et al., 2008) and trees have been shown to be the main cause of variability in snow redistribution at tree line as snow is scoured from the upwind side and deposited on the downwind side of trees (Daly, 1984; Liptzin and Seastedt, 2009). Thus, wind redistribution and the presence of krummholz at the USA may explain some of the variability in snow accumulation and thus timing of maximum soil moisture between heated and control plots. To explain the fact that USA heated plot 21h experienced max moisture before disappearance may be explained by the column remnant. After the majority of plot was melted and maximum soil moisture reached, the column in the center of the plot indicated the presence of snow cover when in actuality there was only a small column left.
While much uncertainty remains about how alpine environments will be affected by winter climate change, warmer air temperatures are highly probable (IPCC, 2007). If mid-winter air temperatures increase, a consequence may be more frequent winter thaw (Callaghan et al., 1998), and thus mid-winter snowmelt and infiltration into soils. Also consistent with the global warming trend is the likelihood that treeline will be more sensitive to warming due to slightly lower elevations and higher air temperatures relative to the alpine. Therefore, our experiment illustrates the potential response of soil temperature and moisture to a warmer winter climate. However, our results likely overestimate the magnitude of soil moisture increases due to caves created by our experiment design that caused large amounts of melt. Additionally, precipitation and SWE trends may vary with climate change. While our experiment illustrates larger amounts of available water throughout the winter in heated plots, post soil moisture values are either the same or lower than control plots indicating greater winter soil moisture does not mean greater soil moisture during the growing season.

1.6.3 Elevational Effects

Overall, the effect of the heaters on snow properties, soil temperature, and soil moisture decreased with increasing elevation. Sites at higher elevations experienced greater snow accumulation and greater wind speeds. Thermal radiation efficiency of the heaters decreases exponentially with increasing wind speed (Kimball et al., 2005). Heater efficiency reaches a maximum of 52% during calm conditions, compared to 4.1% efficiency at a wind speed of 10 m s\(^{-1}\) (Kimball et al., 2005). Additionally, wind speed increases with increasing distance from the snow surface (Barry and Chorley, 2003). Therefore, as distance decreases between heaters and the snow surface, more heat will reach the snow surface for a given wind speed.
The distance between heaters and the snow surface varied between sites with the LSA experiencing the largest distance. At the LSA, low snow accumulation maintained a difference of ~150cm between heaters and the snow surface. Lack of wind prevented sufficient heat losses, heater efficiency was high, and snow melted quite rapidly after precipitation events. However, at USA and ALP sites heaters were covered once snow accumulation reached 120cm (by mid-November for USA and February to March at the ALP). Thus, before snow reached the heaters early in the season, significant amounts of heat was lost due to convection by wind and little difference in snow depth was noticed between heated and control plots. Once, heaters were covered and caves were formed, caves likely protected heaters from wind resulting in greater heater efficiency.

In response to differences in site evolution, the magnitude of difference between heated and control plot’s soil temperature and soil moisture decreased with increasing elevation. At the LSA, high heater efficiency results in warmer and wetter soils. At the USA and ALP sites, when snow is beneath heaters, little difference is noticed between heated and control plots. Then, once caves form large amounts of energy reach the soil, resulting in earlier soil warming and higher soil moisture.

Thus, it appears high elevations have a threshold with heater efficiency. Heaters have low efficiency until snow gets deep, caves form, and the efficiency switches. Similarly, it is predicted that the greatest snowpack and melt responses to climate change will be for the areas that remain close to freezing throughout the winter season (IPCC, 2007) where decreasing snow to total precipitation ratios, a decreasing snowpack, earlier spring runoff, and lower summer flows have been noted (Stewart, 2009). Conversely, higher elevations remain well below freezing during the cold season and therefore have reduced sensitivity to increases in temperature...
on snow accumulation and melt (Stewart, 2009). Thus, this experiment is able to capture some processes that may result from climate change and with careful interpretation may provide insight into the elevation effects of climate change on ecological and hydrological systems.

1.6.4 Energy Balance

Various processes affect the energy balance of the snow surface. Incoming solar radiation, outgoing (reflected) shortwave radiation, incoming longwave radiation, and outgoing (emitted) longwave radiation make up the total radiative fluxes. Temperature and vapor pressure gradients determine mass losses to sensible and latent heat fluxes. Additionally, heat may be lost or gained via ground heat flux. The sum of these parameters determines the amount of energy and mass lost and gained from a snowpack throughout the winter and thus, the amount of available water (Figure 1.25).

Figure 1.24: Processes involved in the energy balance at the snow surface. Components include incoming shortwave radiation (SW$_{in}$), outgoing shortwave radiation (SW$_{out}$), incoming longwave radiation (LW$_{in}$), outgoing longwave radiation (LW$_{out}$), sensible heat (H), latent heat (LE), ground heat flux (G), and the internal energy storage ($\Delta S$).
This study illustrates the significant influence that heaters have on the snow surface energy balance by increasing $LW_{in}$. What is not clear from this study is how the partitioning of energy is altered between heated and control plots (Figure 1.26). In particular, are turbulent fluxes (i.e. latent and sensible heat) larger or smaller relative to control plots? Are there more losses to evaporation and sublimation, and thus less SWE in heated plots compared to control plots? Or are there greater amounts of melt occurring in heated versus control plots? Increased partitioning of snowmelt into atmospheric water loss may lead to reductions in groundwater recharge and surface runoff (Molotch et al., 2009). Thus, addressing these questions is essential to understanding the response of hydrological and ecological systems to a warming climate.

Figure 1.25: Snow surface energy balance of heated plots. The magnitude and direction of turbulent fluxes is unknown, as well as the portioning of energy into melt, evaporation, or sublimation.
1.7 Conclusion

Heaters altered the energy balance of heated plots relative to control plots via increased incoming radiation that was distributed unevenly resulting in an ephemeral snowcover at the LSA. In response, soil temperatures were above 0°C during periods of no snowcover. Soil moisture in heated plots was higher than control plots throughout the season reaching maximum soil moisture around the end of March and maintaining that value until after snowmelt in control plots.

At the USA and ALP sites, caves began to form once snow depth exceeded heater height due to the low thermal conductivity of snow and high input of energy from the heaters. Snow accumulated more rapidly at the USA than the ALP and as a result caves formed earlier and were larger at the USA and the upper two plots at the ALP. Because the air filled caves insulated the ground, soil temperatures never dropped below 0°C at the USA. Soil moisture in USA heated plots followed similar patterns to the LSA, illustrating higher values throughout winter but lower maximum soil moisture relative to control. At the ALP, caves formed later in the season and did not form at all in the two lower plots. As a result, there was less of a difference in snow disappearance date, soil temperature and soil moisture between heated and control plots. The magnitude of differences between ALP control and heated plots increased with larger amounts of snow accumulation and thus the influence of caves.

All three sites experienced episodic snowmelt throughout the winter, as well as lateral transport due to the influence of heaters outside of plots. At the USA and ALP, more snow (water) preferentially collected in holes and melted, lending itself to higher soil moisture in heated plots, and potentially higher SWE. Overall, the influence of heaters on snowpack properties and soil microclimate decreased with increasing elevation. In alpine environments
associated with the USA and ALP, high winds decrease the heaters efficiency and thus, the amount of energy actually reaching the snow surface.

Furthermore, this study illustrates the significant influence that heaters have on the snow surface energy balance and the important role that those effects have on snow properties, soil temperature and soil moisture.

Further research is needed to determine how an increase in energy input ($LW_{in}$) in heated plots affects the partitioning of energy into mass gains or losses relative to control plots. Understanding the relative contributions of each energy balance component would provide further insight into the relationship between climate change, hydrology, and ecological impacts from near-infrared warming experiments.

Additionally, future warming experiments should revise experiment design at higher elevations where artifacts, such as caves should be avoided if the experiment is to provide insight into the effects of climate change. One problem that remains is the lack of heater efficiency in windy environments. Even at higher heater energy fluxes, a large amount of energy would still be lost. Perhaps a more appropriate method for simulating the effects of global warming on snow properties at higher elevations is snow removal.
Modeling of Snow Surface Mass and Energy Fluxes in a Climate Manipulation Experiment on Niwot Ridge, Colorado
Abstract

Heaters altered the snow surface-atmosphere energy balance by increasing incoming longwave radiation (LW_{in}). In order to derive estimates of energy and mass balance exchange at the snow surface in heated and control plots, the one-dimensional, physically based snowmelt model SNOWPACK was used. In heated plots at the LSA, net radiation accounted for 80 to 100\% of the energy available to melt snow compared to 35\% in control plots. In heated plots at the USA/ALP, net radiation accounted for 100\% of the energy available for melt when snow depth exceeded heater height. However, when snow depth is below the heaters (1.2 m), only 5\% of the energy for snowmelt comes from net radiation. Model results illustrate greater mass losses to sublimation/evaporation (54 to 83\% of total SWE) in heated plots compared to control plots (6 to 38\% of total SWE). Thus, the increase in net radiation resulting from the heaters is compensated by greater losses of energy to turbulent fluxes and mass losses to sublimation. Additionally, grain size at the snow surface was \approx 50 \% larger in heated plots compared to control plots during snowmelt. At the USA/ALP densities between heated and control plots were similar during the winter but increased values as much as 40\% higher in the heated plots during the last week of snowmelt. Results emphasize the influence of IR heaters on the snow surface-atmosphere energy balance and the importance of understanding how energy is partitioned in these IR warming experiments for improved interpretation of feedbacks between climate, hydrology, and ecology.
2.1 Introduction

In high-latitude snow covered environments, snowmelt provides the dominant source of water for ecosystems and socioeconomic needs. Radiative and turbulent fluxes are the two primary mechanisms that drive snowmelt and control the exchange of mass and energy at the snow surface (Male and Gray, 1981). Therefore, the partitioning of energy into mass (i.e. water) loss via evaporation/sublimation or gains via melt/runoff has a direct bearing on ecology and hydrology. Determining turbulent fluxes of latent and sensible heat in mountainous terrain is complicated due to complex terrain where relevant variables vary spatially and temporally and high wind speeds are common (Oke, 1987; Pohl et al., 2006b).

The rate and magnitude of fluxes is largely determined by atmospheric conditions (air temperature, relative humidity, and wind speed) (Male and Gray, 1981; Cline, 1997a,b) and surface topography (Pohl et al., 2006 a,b). For example, Harding et al. (2002) studied energy fluxes during the winter at three high latitude sites in Fenno-Scandinavia. Results found large differences in the partitioning of energy between radiative and turbulent fluxes and resultant timing of snowmelt due to different weather and surface characteristics. Other studies have found that net radiation is the dominant source of energy for snowmelt providing 66 to 100% of the total, whereas turbulent fluxes are small in magnitude and tend to cancel (Cline, 1997b; De la Casiniere, 1974; Marks and Dozier, 1992). However, Cline (1997b) reports a 46% energy contribution from net radiation (54% contribution from turbulent fluxes) for one year, and a 75% contribution from net radiation (25% from turbulent flux) the following year due to meteorological differences between the two years with the year of greater turbulent fluxes having warmer temperatures and higher humidity. Similarly, Pohl and Marsh (2006) report contribution from turbulent fluxes to be as high as 50% on cloudy, warm, windy days. Due to the complexity
of processes involved in mass and energy exchange between the snow surface and atmosphere, the spatial variability of these fluxes across a basin and relative contribution of each are not well known (Gustaffson et al., 2001; Hood et al., 1999; Lang, 1981). Even less clear is how climate change may affect these fluxes, especially during spring snowmelt.

Energy and mass fluxes to the snowpack are reported as positive (i.e. gains) and fluxes away from the snowpack to the atmosphere are reported as negative (i.e losses). In general, during snowmelt as air temperature increases above 0°C, the temperature difference between air and snow becomes positive and sensible heat fluxes are directed to the snow cover, contributing to melt (Cline, 1997a; Marks and Dozier, 1992). Because both evaporation and condensation may occur during snowmelt, latent heat may be positive or negative (Cline, 1997a). Turbulent fluxes are particularly important in open, windy environments where sublimation and evaporative losses can be quite large due to high winds and blowing snow (Berg, 1986).

Kattelmann and Elder (1991) found sublimation losses up to 18% of total precipitation in the Sierra Nevada, whereas sublimation losses of 15% of total snow water equivalent have been reported for Niwot Ridge in the Colorado Front Range (Hood, 1999). In forested areas, snowpacks receive less incoming shortwave radiation, wind, and daily variations in temperature, whereas incoming longwave radiation and humidity are higher. Thus, turbulent fluxes are dampened more in forested areas relative to open areas (Links and Marks, 1999; Molotch et al., 2009).

A number of techniques exist to estimate turbulent fluxes, such as the Bulk Profile method (Moore, 1983), aerodynamic profile method (Cline, 1997b; Hood et al., 1999), and the eddy covariance method (Blanken et al., 2009). These methods require instrumentation and, thus cannot be applied to all situations. Physically-based modeling provides a means to calculate
turbulent fluxes in areas where instrumentation is not possible or limited. Anderson (1976) developed a point–based snow cover energy balance model to predict snowmelt that incorporated mathematical representations of the densification of snow as well as the retention and transmission of liquid-water. Since then, numerous physically-based snowmelt models have been designed, improved upon, and successfully validated in a number of locations and applications (Bartlett et al., 2002; Gustaffson et al., 2001, 2003; Jordan, 1999, Lehning et al., 2002a, b; Lundy et al., 2001; Marks and Dozier, 1992; Marks et al., 1999). SNTHERM (Jordan, 1991) is an energy and mass balance model developed to specifically address snowpack temperature. CROCUS (Brun et al., 1989, 1992) incorporates detailed descriptions of snow metamorphism for improvement in the modeling of snow cover stratigraphy for avalanche forecasting purposes. The SNOWPACK model (Bartelt et al., 2002; Lehning et al., 1999; Lehning et al., 2002a, b) was developed by the Swiss Federal Institute for Snow and Avalanche Research to predict snowpack settlement, layering, surface energy exchange and mass balance. Improvements upon previous models are the inclusion of snow grain bond size in calculations of snow microstructure which significantly influences the mechanical properties of snow (Lehning et al., 2002b). Microstructure is then used to predict bulk properties such as thermal conductivity and viscosity.

At Niwot Ridge, Colorado, Kueppers and colleagues have initiated a global warming experiment using IR heaters at three sites along an elevational gradient. The first part of this research addressed the influence on snowpack properties and soil temperature and moisture in response to the increased energy input. Another important aspect of this work involves alteration of the partitioning of energy and mass associated with the climate manipulation experiments. In order to derive estimates of energy and mass balance exchange at the snow surface in heated and
control plots, the one-dimensional, physically based snowmelt model SNOWPACK was used by performing a sensitivity analysis to see what combinations of energy inputs to the snowpack would result in melt out times similar to that observed from snow depth measurements. This research aims to address the influence that IR heaters may or may not have on radiative and turbulent fluxes. Research questions are:

- How do IR heaters alter turbulent fluxes of the snow-atmosphere interface?
- How do IR heaters effect sublimation rates?
- How do IR heaters effect snow density and grain size?
2.2 Site Description

Research was conducted during the winter and spring of 2010 on Niwot Ridge, located in the Colorado Front Range of the Rocky Mountains about 5km east of the Continental Divide (40°03’N, 105°35’W). This site is an UNESCO Biosphere Reserve and the location of the Niwot Ridge (NWT) Long-Term Ecological Research (LTER) site. Warming experiments were located at three sites located along an elevation gradient within the Niwot Ridge LTER research area, the Lower Alpine Site (LSA; 3048 m), the Upper Subalpine (USA; 3367 m), and the Alpine site (ALP; 3517 m) (Figure 2.1).

![Figure 2.1: View of Niwot Ridge and locations of LSA, USA, and ALP sites. LTER managed sites are also shown (C1, Saddle, and D1). Inset shows the layout of sensors and equipment for each site. Each heated plot contains 6 heaters placed hexagonal. Control plots are located with ~3 meters of the associated heated plot. Snow depth sensors are suspended over the center of each plot by a horizontal boom. Soil moisture and temperature sensors are also located in four equally spaced locations in the plot (5-10cm) and in the center of each plot (15-20cm).](image)

The LSA is located within a closed-canopy subalpine forest (forest height about 20 m). Less than a half km from LSA, the NWTLTER collects numerous measurements at the C-1 site, including meteorological data, the SnoTel network site NIWOT 663, and soil moisture and temperature.
This site has a mean annual temperature of 1.5 °C and receives about 800 mm of precipitation annually, with 60% as snow, and 40% as rain.

The USA site is located near treeline in an opening within a stand of krummholz consisting of Engelmann Spruce (*Picea engelmannii*) and Subalpine Fir (*Abies lasiocarpa*); maximum tree height is 3 meters. The ALP site is characterized by alpine tundra, located approximately a half km to the west of USA and is located in a moist meadow plant community. Near both the USA and ALP sites, the NWTLTER program maintains instrumentation similar to C1 at The Saddle (40° 03’17”N; 105° 35’ 21”W; 3528m), which is located in alpine tundra. The Saddle maintains a meteorological tower capable of closing the energy balance, a snow lysimeter array, an index snowpit, and a National Atmospheric Deposition Program (NAPD) precipitation collector (site CO02). The climate of the Saddle area consists of long, cool winters and a short growing season of one to three months. Mean annual temperature is -3.7 °C and annual precipitation is 1000 mm (Williams et al., 1996) with approximately 80% of the annual precipitation falling as snow (Caine, 1995). Snow depth accumulation is extremely variable due to high winds and topography. Snow cover generally lasts October to June. The alpine portion of Niwot Ridge experiences strong westerly winds that commonly redistribute snow (Erickson et al., 2005).
2.3 Methods

2.3.1 Experiment Design

Within each study site were replicated heated and control plots separated by about 3 m (Figure 2.1). Each plot was 3 m in diameter (7.07 m²). Heater arrays were designed according to the geometry in Kimball et al. (2008). Each heater was mounted 1.2 m above the ground. There were 6 heaters per heated plot arranged in a hexagonal pattern with each tilted at 45° from horizontal to achieve the most uniform distribution of radiation (Kimball, 2005) (Figure 2.2). For this study, the number of plots used varied with site, with LSA containing four pairs of heated and control plots, three pairs at the USA, and five pairs at the ALP.

Heated plots were warmed using Mor Electric Heating Association Inc. IR ceramic heaters (Model FTE-1000). Each heater is 245 mm long x 60 mm wide, with a maximum output of 1000 W and transmittance window of 4.5-42 μm. Heaters were operated in a constant flux mode with a temperature target of ~4-4.5° C increase in surface soil temperatures averaged over
the growing season, according to the protocols of Harte et al. (1995) and Harte and Shaw (1995). Because soils at Niwot Ridge are wetter than those of Harte et al. (1995), more than double the radiation is needed to double the temperature effect since more of the increased radiation gets used to evaporate water. From October 2009 – June 2010, each heater was operated at 50% power, thus each heater put out about 500 W. It is estimated that ~50% of the output is lost outside the plots reducing the 500W to 250 W. Since each heated plot was roughly 7 m$^2$ in area, if the heat was evenly distributed the energy flux would be 214 W m$^{-2}$, or $1.85 \times 10^7$ J m$^{-2}$ d$^{-1}$). Heaters were programmed to turn off in high winds (>15 m s$^{-1}$).

Located in approximately the center of each site was a 3 cup anemometer, manufactured RM Young supplied by Campbell Scientific (Model 03101-L), measuring wind speed, and a Visalia HMP 45 six-wire sensor mounted inside a Gill aspiration shield measuring temperature and relative humidity. All Sensors were mounted 3 meters above bare-ground. At C1, net radiation was measured with a Kipp and Zonen CNR-1 at 25 m above the ground. At the Saddle, incoming and outgoing shortwave radiation was measured with a Kippen and Zonen CM14 pyranometer and incoming and outgoing longwave radiation was measured with a Kipp and Zonen CG2 pyrgeometer at 6 m above the ground.

2.3.2 Snow properties

In addition to the existing instruments, Judd Communications LLC Ultrasonic snow depth sensors were installed in each plot, which became operational in late January (ALP), March (LSA), and early April (USA) of 2010. Each snow depth sensor was 8 x 8 x 13 cm in size with a beamwidth of $22^\circ$ and an accuracy of 1 cm. Depth was recorded hourly with Campbell CR1000 and CR10X dataloggers. Each snow depth sensor was located at various heights above
the above the ground so as to avoid being buried by the seasonal snowpack (Table 1.1) and located over the center of each plot on an array of pipes extending from the sides of the plots, so as to avoid possible heat conduction if a support pipe were placed in the plot itself.

Snowpits at the Saddle and C-1 sites were sampled approximately weekly for physical and chemical parameters as part of the NWT LTER project. Snowpits were excavated to sample snow properties from the snow/air interface to the snow/ground interface following the protocols of Williams et al. (1999). Density was measured in vertical increments of 10 cm using a 1-L (1000 cm³) stainless steel cutter and an electronic scale (+/- 2 g). Temperature of the snowpack was measured every 10 cm with 20–cm long dial stem thermometers, calibrated using a one-point calibration at 0°C. The height of stratigraphic layers above the snow/ground interface was recorded, along with the thickness and type of layer (buried sun crust, ice lens, coarse to fine grain transition), and grain type and grain size of each layer were determined using a 10x magnifying loupe and a gridded crystal card. The working wall of the snowpit was oriented so that it remained shaded from the sunlight. The snowpit was refilled after measurements were taken. Because the snowpit results in destructive sampling, the next snowpit was excavated approximately 1 meter from the southern wall of the last snowpit to avoid edge effects. Thus snowpits did not sample exactly the same snowpack at the same location. Additional snowpits were excavated in a few selected experimental plots for more direct comparison of snow properties; however regular snowpits were not conducted within the plots because of the concern that such sampling might induce artifacts to the experimental treatments.
2.3.3 SNOWPACK Model

SNOWPACK numerically solves the partial differential equations governing mass, energy, and momentum conservation within the snowpack using the finite element method. Snowpack behavior is described by two mass conservation equations for the vapor and water phases, one bulk temperature equation, and one momentum equation for the ice phase (Bartelt et al., 2002). Snow is modeled as a three-phase porous media consisting of volumetric fractions of ice, water, and air. Rates of snowpack settlement and snowpack density are calculated based on a microstructure-dependent viscosity which is determined by the thermal regime of the snowpack (temperature and temperature gradient). The temperature profile of the snowpack is calculated using the thermal conductivity formulation developed by Adams and Sato (1993).

The SNOWPACK model is executed using measured meteorological and snow profile data. Input variables needed include: air temperature, relative humidity, wind speed and direction, snow surface temperature, snow/soil interface temperature, incoming and outgoing shortwave radiation, incoming longwave radiation, and snow depth. Each model was run with hourly meteorological data and measured snowpack information from C1 (LSA) or Saddle (USA/ALP) index snowpits. For each snow layer, layer height, snow temperature, snow grain size, shape, dendricity, sphericity, % liquid water content, ice, and air pore space were described. SNOWPACK uses measured snow depths to determine snow precipitation rates from meteorological variables (air temperature and wind speed) and the calculated settling rates (Lehning, part III). Density of newly fallen snow is estimated from a statistical relationship derived from measurements at a field site in Davos, Switzerland and ranges between 30 to 150 kgm$^{-3}$ (Lehning et al., 2002a). Measured values in the Colorado Front Range fit within this range.
Two possible surface boundary conditions exist to govern energy exchanges at the snow surface. When the snow surface temperature is below the melting temperature (-1.3°C), the Dirichlet boundary condition is applied,

\[ T_s(z=h, t) = T_h(t) \]

where the snow temperature, \( T_s \), at a particular snow height, \( h \), and time, \( t \), is equal to the measured air temperature, \( T_h \). When the surface temperature approaches the melting temperature, the program switches to the Neumann condition where the net longwave radiation is calculated from the snow surface temperature using an estimation of the atmospheric emissivity and the Stefan–Boltzmann constant (Bartelt et al., 1999),

\[ k_s \frac{\partial T_s(z=h,t)}{\partial z} = q_{lw} + q_{sh} + q_{th} + q_{rr} \]

where \( k_s \) is the thermal conductivity of snow, \( q_{lw} \) is the net long-wave radiation energy, \( q_{sh} \) is the sensible heat exchange, \( q_{th} \) is the latent heat exchange, and \( q_{rr} \) is the heat flux from rain. Surface heat fluxes are calculated assuming a neutral atmospheric surface layer and using the Monin-Obukhov similarity theory (Monin and Obukhov, 1954). The effects of wind pumping (Kaimal and Finnigan, 1994) and blowing snow conditions (Lehning et al., 2000) are accounted for.
2.4 Results

2.4.1 LSA/C-1

2.4.1.1 Control Plots

The Model was run from 5 March and precipitation was driven with LSA plot 3 snow depths which most closely follow the C1 index pit trend. Snowpack properties were initialized with C1 snowpit results from 5 March. As a result of snow depth driven precipitation, measured and modeled snow depths follow each other very closely (Figure 2.3).

![Figure 2.3: Modeled (solid red line) and measured (solid blue line) snow depths for LSA control plot.](image)

In general, the model reproduced the bulk properties (i.e. snow depth, density, temperature) of the control plots snowpack fairly well. Figure 2.4 compares measured to modeled results of density and temperature for three dates throughout the winter: 10 March, 20 April, and 23 May.
Figure 2.4: From left to right: Measured and modeled snow profile temperature results, and measured and modeled snowpit density results for the LSA/C-1 on 10 March (upper panel), 20 April (middle panel), and 23 May (lower panel). Snow depth (cm) is located on the y-axis and temperature (°C) and density (kg m⁻³) are located on the x-axis.

Modeled temperatures were ~1 to 2 °C warmer than measured temperatures at the surface and bottom through March, with temperatures in the upper portion of the snowpack showing the largest discrepancies (Figure 2.4). Snow temperatures the rest of the season were modeled well with observed and modeled temperatures reaching isothermal conditions around 21 April (Figure 2.4). Overall, general trends in density are well simulated and within 12 to 30 % of actual values. During March, the top snow layers are well-modeled relative observed, however, the bottom 45 cm are underestimated by 53% (Figure 2.4). Densities in April and May are underestimated in the upper half of the snowpack but overestimated in the bottom 40-50 cm by 21 to 30 % (Figure 2.4). Modeled snow melt out date is within 24 hours of observed.

Satisfaction with modeled LSA control results allowed us to validate modeled results of turbulent fluxes of latent and sensible heat, as well as examine mass losses or gains of evaporation, sublimation, and runoff, in model simulations of heated plots.

In control plots at the LSA, shortwave radiation ranged between 0 and 150 W m⁻² throughout the winter increasing to values as high as 300-900 W m⁻² during snowmelt (Figure 2.5). Whereas, LW_net fluctuated between 0 and -100 W m⁻² throughout the snow cover season.
Figure 2.5: LSA control measured net shortwave (upper panel) and net longwave radiation (lower panel) from 24 March to the end of snowmelt on 4 June.
Figure 2.6: LSA control modeled latent heat exchange (upper panel) and sensible heat exchange (lower panel). Energy losses (gains) due to latent heat/sublimation are expressed as negative (positive) values.

Results illustrate latent heat fluxes that varied between -200 and 200 W m\(^{-2}\) throughout the season. Sensible heat remained positive at approximately 200 W m\(^{-2}\) (Figure 2.6). Thus, turbulent fluxes accounted for 65% of energy available for snowmelt and net radiation only accounted for 35%. Total evaporative/sublimation losses were 46 mm (6% of total mass lost). The first losses to melt/runoff occurred on 3 April (~7 mm) with consistent runoff of 10-20 mm daily beginning on 20 May and total runoff equaling 701 mm.
2.4.1.2 Heated plots

In order to simulate the effects of heaters on various energy balance parameters in warmed plots, different combinations of LW_in, soil temperatures and snow depth corresponding to different storm events were used as model input. Two storm events that resulted in snow accumulation in heated plots occurred on 24 March and 11 May resulting in 37 cm and 30 cm of snow in heated plots, respectively. Thus, one model run began 24 March and was initialized with 37 cm of low-density snow (150 kg m$^{-3}$), the same value observed in the upper layer of control plots. Another model run began on 11 May and was initialized with 30 cm of low-density snow. Soil temperatures were held constant at 5 °C (top 10 cm). This temperature was chosen because the observed average soil temperatures in heated plots during no snow accumulation periods were around 5 °C. Several model runs were carried out for each date by increasing LW_in by 50 W m$^{-2}$, 100 W m$^{-2}$, 150 W m$^{-2}$, and 250 W m$^{-2}$. Snow depth was not an input to drive precipitation, e.g. precipitation was assumed to be zero.

Initializing the model with increased LW_in of 150 W m$^{-2}$ relative to control plot resulted in snow disappearance in 7.5 days compared to 6 days observed for the 24 March snowfall event. Contrary to the modeled control, modeled LW_net in the heated plot was an initially positive during the first 2 days of snow cover period. Then LW_net values fluctuated between gains and losses of -25 W m$^{-2}$ to 25 W m$^{-2}$ until snow disappearance (Figure 2.7).
After the first two days of snow cover, latent heat flux also switched from positive to a negative flux of energy from the snowpack. Results illustrate overall greater losses to latent heat in heated plots compared to control plots (Figure 2.8). In the heated plot, sensible heat exchange varied much more with time than control. Sensible heat in control was always positive, whereas heated plots experienced energy gains and losses of -350 W m\(^{-2}\) to 350 W m\(^{-2}\) (Figure 2.9).

Contrary to control plots, net radiation accounted for 100% of energy available for melt in heated plots during March.
Figure 2.8: Model run initiated on 24 March with 37 cm of snow in heated plots. Modeled latent Heat exchange (y-axis) through time (x-axis) beginning on 24 March for LSA control plot (upper panel) and heated plot (lower panel). Energy losses (gains) due to latent heat are expressed as negative (positive) values.
Figure 2.9: Model run initiated on 24 March with 37 cm of snow in heated plots. Modeled sensible heat exchange (y-axis) through time (x-axis) beginning on 24 March for LSA control plot (upper panel) and heated plot (lower panel). Energy losses (gains) due to sensible heat are expressed as negative (positive) values.

Total mass losses to evaporative/sublimation were high at 45 mm and little was lost to snowmelt (16.6 mm) (Table 2.1). Thus, approximately 73% of the total snowcover was lost to evaporation/sublimation in heated plots compared to only 6% in control.

<table>
<thead>
<tr>
<th>Model Run</th>
<th>Evaporation/Sublimation Loss</th>
<th>Melt/Runoff Loss</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>6%</td>
<td>94%</td>
</tr>
<tr>
<td>LSA March 24 Snow Event (Heated)</td>
<td>73%</td>
<td>27%</td>
</tr>
<tr>
<td>LSA May 11 Snow Event (Heated)</td>
<td>54%</td>
<td>46%</td>
</tr>
</tbody>
</table>

Table 2.1: Mass losses to evaporation/sublimation and melt/runoff for all LSA model runs. Calculated losses are for the time period of snow cover. Values are stated in % of total mass.
Initializing the model with increased LW$_{in}$ of 150 W m$^{-2}$ relative to control plot for the 11 May snow event resulted in snow disappearance in 9.5 days, which is 1.5 days longer than observed. Initially, LW$_{net}$ provided a positive energy flux of 50 W m$^{-2}$ then after the first 6 hours of melt switched to a negative flux (Figure 2.10).

![Figure 2.10: Upper panel: Heated plot snow depth (y-axis) and snow temperature over time (x-axis). Temperatures range from 0°C (red) to -20°C (dark blue). Middle panel: Control plot LW$_{net}$ (y-axis) over time (x-axis). Lower Panel: Heated plot LW$_{net}$ (y-axis) over time (x-axis) for model run initiated on 11 May with 30 cm snow and LW$_{in}$ of 150 W m$^{-2}$.

Corresponding with positive LW$_{in}$, latent and sensible heat gains of 250 to 300 W m$^{-2}$ occurred in heated plots for the first 2 days (Figure 2.11, 2.12). Latent heat flux in control plots was around 0 W m$^{-2}$ at this time and sensible heat flux was only 50 W m$^{-2}$. After 20 cm of snow had melted, latent heat flux switched to an energy loss for the majority of the remaining snow cover
period (Figure 2.11). Overall, net radiation provided 80% of the energy available for melt and turbulent fluxes accounted for only 20%. Mass losses occurred almost equally between evaporation/sublimation (18.5 mm, 54%) and melt/runoff (16 mm, 46%) (Table 2.1).

In heated plots during March and May, densities were greater than control plots by approximately 50%. Grain size at the snow surface was also much larger in heated plots measuring 4 mm in diameter compared to 2 mm in diameter in control plots.

Figure 2.11: Model run initiated on 11 May with 30 cm of snow in heated plots. Modeled latent Heat exchange (y-axis) through time (x-axis) beginning on 11 May for LSA control plot (upper panel) and heated plot (lower panel). Energy losses (gains) due to latent heat are expressed as negative (positive) values. Note: Snow in heated plots disappears on 20 May.
Figure 2.12: Model run initiated on 11 May with 30 cm of snow in heated plots. Modeled sensible heat exchange (y-axis) through time (x-axis) from 11 May until snow disappearance for LSA control plot (upper panel) and heated plot (lower panel). Energy losses (gains) due to sensible heat are expressed as negative (positive) values.

2.4.2 USA-ALP/Saddle

2.4.2.1 Control plots

The model was run from 1 February to snow disappearance and precipitation was driven with ALP control plot 76 snow depths, which most closely follow the saddle index pit snow depth trend. As a result of snow depth driven precipitation, measured and modeled snow depths follow each other very closely (Figure 2.13).
In general, the model reproduced the bulk properties (i.e. snow depth, density, temperature) of the control plots snowpack fairly well. Figure 2.14 compares measured to modeled results of temperature and density for three dates throughout the winter, 2 February, 10 March, and 20 May. Throughout February and March, temperatures were underestimated at the surface by 1 to 2 °C and overestimated by several degrees at the bottom 20 to 30 cm (Figure 2.14). Modeled snowpack results did not reach isothermal temperatures until approximately one week after measured and May snowpack temperatures were underestimated by 1 to 2.5 °C (Figure 2.14). Overall, trends in density are simulated well. Densities were underestimated in the upper portion of the snowpack by 10 to 23 % and overestimated by 15 % for the bottom 40 cm of the snowpack throughout the entire season (Figure 2.14).
Figure 2.14: From left to right: Measured and modeled snow profile temperature results, and measured and modeled snowpit density results for the USA-ALP/Saddle on 4 February (upper panel), 10 March (middle panel), and 20 May (lower panel). Snow depth (cm) is located on the y-axis and temperature (°C) and density (kg m\(^{-3}\)) are located on the x-axis.

Modeled snow melt out date was within 24 hours of observed. This difference in melt out date between modeled and observed is likely due to spatial variability of snow properties resulting from different locations of plot 76 relative to the saddle index pit. For example, the Saddle index snowpit melted out approximately one week before plot 76, and likely reached isothermal conditions prior to plot 76. Satisfaction with modeled USA-ALP control results allowed us to validate modeled results of turbulent fluxes of latent and sensible heat, as well as mass losses or gains of evaporation, sublimation, and runoff, in model simulations of heated plots.

During the winter in control plots at the USA and ALP, shortwave radiation slowly increased from ~75 W m\(^{-2}\) in January to 450-750 W m\(^{-2}\) during snowmelt (Figure 2.15). Whereas, LW\(_{\text{net}}\) fluctuated between 0 and -100 W m\(^{-2}\) throughout the winter becoming slightly positive (~25 W m\(^{-2}\)) during snowmelt.
Modeled latent heat fluxes were negative throughout the snowcover season fluctuating between 0 and -200 W m\(^{-2}\) (Figure 2.16). Sensible heat fluxes were consistently positive, ranging between 0 and 300 W m\(^{-2}\). Thus, net radiation accounted for 100% of the energy required for snowmelt.
Figure 2.16: ALP control modeled latent heat exchange (upper panel) and sensible heat exchange (lower panel). Energy losses (gains) due to latent heat/sensible heat are expressed as negative (positive) values. The solid black line represents 0 W m$^{-2}$.

Evaporative/sublimation losses were high through the end of May and beginning of June (total of 301 mm). Runoff reached daily losses of 10-20 mm beginning around 10 June, totaling 488 mm overall. Therefore, 62% of the total mass was lost to runoff and 38% was lost to evaporation/sublimation (Table 2.2).
<table>
<thead>
<tr>
<th>Model Run</th>
<th>Evaporation/Sublimation Loss</th>
<th>Melt/Runoff Loss</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>38 %</td>
<td>62 %</td>
</tr>
<tr>
<td>Snow depth below heaters (Feb)</td>
<td>83 %</td>
<td>17 %</td>
</tr>
<tr>
<td>Snow exceeds heaters (no wind)-model 3</td>
<td>18.5 %</td>
<td>82.5 %</td>
</tr>
<tr>
<td>Full season</td>
<td>42 %</td>
<td>58 %</td>
</tr>
</tbody>
</table>

Table 2.2: Mass losses to evaporation/sublimation and melt/runoff in control and heated plot model simulations.

2.4.2.2 Heated Plots

Results from the first part of this research illustrated large differences in snowpack properties, soil temperature and soil moisture when snow depth was beneath the heaters versus when snow exceeded the heaters. When snow exceeded heater height caves formed altering the thermal regime of the snowpack relative to plots without caves. Therefore, in order to determine the influence of heaters on the partitioning of energy in heated plots, three different model runs were carried out to simulate the effect of heaters when snow was below the heaters (< 120 cm) and above the heaters (200 cm) with caves. All model runs were driven with increased LW_in values of 150 W m⁻² relative to control plots. In the first model run snow depth was not an input to drive precipitation. Instead, the model was initiated with a snow depth of 100 cm on 4 February and ran until snow disappeared to model what may be happening when snow depth is below heater height. Results show that a snowpack of 100 cm melted in 2 days, 18:15 hrs. During the snow cover period, losses to LW_net in heated plots were similar to control plots (Figure 2.17), as were fluxes in latent and sensible heat (Figure 2.18). There were slightly greater
losses to latent and sensible heat during the last 24 hours of melt in heated plots when only a few centimeters of snow remained.

Figure 2.17: Top panel: USA-ALP heated plot snow depth (y-axis) and snow temperature over time (x-axis) for model run initiated in February with 100 cm of snow, increased LW\textsubscript{in} of 150 W m\textsuperscript{-2}. Temperatures range from 0\textdegree C (red) to -20 0\textdegree C (dark blue). Bottom panel: Heated plot LW\textsubscript{net} (y-axis) over time (x-axis).
Figure 2.18: USA/ALP model run initiated on 4 February with 100 cm of snow in heated plots. Modeled latent heat exchange (y-axis) through time (x-axis) for control plot (upper panel) and heated plot (lower panel). Energy losses (gains) due to latent heat are expressed as negative (positive) values. Black line represents 0 W m$^{-2}$.

After the majority of snow had melted and only a few centimeters remained, sensible heat and latent heat exchange became negative in heated plots (Figure 2.19). Contrary to control plots, turbulent fluxes accounted for 95% of total energy for mass and only 5% from net radiation. Evaporation/sublimation accounted for 83% (7.5 mm) of total mass lost with only 17% (1.5 mm) lost to melt/runoff in heated plots (Table 2.2).
Figure 2.19: USA-ALP model run initiated on 4 February with 100 cm of snow in heated plots. Modeled sensible heat exchange (y-axis) through time (x-axis) for control plot (upper panel) and heated plots (lower panel). Energy losses (gains) due to sensible heat are expressed as negative (positive) values.

In order to simulate the effects of heaters on energy balance parameters when snow exceeded heater height (>120 cm), the model was initialized with 200 cm of snow on 4 February and wind speeds were decreased to 0 ms\(^{-1}\) to create the environment most characteristic of caves within heated plots. Results illustrate snow disappearing on 28 May, after approximately four months. Losses to LW\(_{\text{net}}\) were much smaller than control plots, ranging between 0 and -25 W m\(^{-2}\) (Figure 2.20).
Trends in latent heat fluxes were similar to control plots but much less in magnitude (~150 W m$^{-2}$) (Figure 2.21). Sensible heat fluxes were also similar in trend but smaller in magnitude in heated plots compared to control plots (Figure 2.22). Net radiation provided 100% energy available for snowmelt.
Figure 2.21: USA/ALP Model run initiated on 2 February with 200 cm of snow and 0 m s⁻¹ wind speeds. Latent heat exchange (y-axis) over time (x-axis) for control (upper panel) and heated plot (lower panel).
Figure 2.22: USA/ALP Model run initiated on 2 February with 200 cm of snow and 0 m s\(^{-1}\) wind speeds. Sensible heat exchange (y-axis) over time (x-axis) for control (upper panel) and heated plot (lower panel).

In heated plots, very small losses to evaporation/sublimation losses occurred consistently throughout the season totaling 96 mm (18.5\%) compared to a total 38\% in the control plot (Table 2.2). Most mass was lost to melt/runoff totaling 451 mm or 82.5\% compared to only 62\% in control plots.

Lastly, the model was run with inputs of snow depths from heated and control plot 76. All parameters remained the same between the heated and control plot, except model run for the heated plot increased LW\(_{\text{in}}\) by an additional 150 W m\(^{-2}\). Results indicate similar losses to LW\(_{\text{net}}\)
(Figure 2.23). Insignificant differences were noticed between control and heated model results of turbulent fluxes of latent and sensible, as well as mass losses (Figure 2.24, 2.25). Same as the control plots, net radiation accounted for 100% of the energy available for melt in the heated plots. Evaporative/sublimation loss versus melt/runoff was 42% and 58%, respectively compared to 38% and 62% in control plots (Table 2.2).

When snow was beneath the heater height, densities were similar. Results from model runs when snow exceeded heater height, as well as when snow depth from heated and control plot 76 were used as input, densities in heated plots began increasing earlier in the season. Two weeks prior to snow disappearance, snow densities were approximately 50% greater in heated relative to control plots. Grain size was also larger at the snow surface in heated plots by 1 to 3 mm in diameter.
Figure 2.23: Upper panel: Heated plot snow depth (y-axis) and snow temperature over time (x-axis) for model run initiated in February with plot 76 snow depths and increased LWin. Temperatures range from 0 °C (red) to -20 °C (dark blue). Middle panel: Control plot LW_{net} (y-axis) over time (x-axis). Lower panel: Heated plot LW_{net} (y-axis) over time (x-axis).
Figure 2.24: USA/ALP Model run initiated on 1 February with plot 76 snow depths. Latent heat exchange (y-axis) over time (x-axis) for control (upper panel) and heated plot (lower panel).
Figure 2.25: USA/ALP Model run initiated on 1 February with plot 76 snow depths. Sensible heat exchange (y-axis) over time (x-axis) for control (upper panel) and heated plot (lower panel).
2.5 Discussion

2.5.1 LSA/C-I

In control plots, turbulent fluxes provided the largest amount of energy to snowmelt at 65%. This study illustrates greater sensible heat fluxes towards the snowpack and greater latent heat fluxes away from the snowpack relative to Blanken et al. (2009) who used eddy flux measurements to estimate turbulent fluxes from a nearby subalpine site on Niwot Ridge. Sublimation losses fit well within the 4 to 7% total sublimation losses in a subalpine forest reported by Pomeroy et al. (1998). Sublimation rates are slightly higher at .64mm d⁻¹ compared to sublimation rates of .42mm d⁻¹ reported by Molotch et al. (2007) from a subalpine study site in the Rocky Mountains of Colorado. Sublimation rates may be higher in this study due to differences in atmospheric conditions (Cline, 1997b; Molotch et al., 2007) or differences in the calculations of the model versus the eddy covariance method used by Molotch et al. (2007).

In heated plots, LWₙₑᵗ was initially positive, providing energy to melt. Then, within approximately 2 days, after approximately 75% of the snow had melted, latent heat and sensible heat switched negative fluxes. While latent heat may fluctuate between positive and negative as a function of weather, it seems counterintuitive for sensible heat to be negative at the snow surface. What may be happening is the additional LWᵢₙ from the heaters raises the snow surface temperature to 0 °C causing melt to occur. During the winter, air temperatures are generally subfreezing causing a negative flux of energy from the snow surface to the air. At 0 °C the snow surface is losing the maximum amount of energy possible to Lₒᵤₜ (316 W m⁻²) causing LWₙₑᵗ to switch from positive to negative. Additionally, results illustrated larger grain sizes at the snow surface of the heated plots which acts to increase absorption of IR radiation (Wiscombe and Warren, 1980) and increase energy input to the snowpack via radiative fluxes. Overall, the
increase in net radiation provided by heaters compensated for losses to turbulent fluxes, thus accounting for 100% of the energy available for melt.

Coincident with larger losses to latent and sensible heat, mass losses to evaporation/sublimation were much greater than control plots by 48 to 67%. These mass losses support soil moisture data from the first part of this study that showed initial infiltration of snowmelt indicated by an increase in soil moisture following snow events, then a fairly rapid decrease in soil moisture as heaters energy went to the evaporation of snow melt. Some error may exist in modeled values of energy fluxes and mass losses due model limitations. First of all, the model cannot input additional LW in at 120 cm from the ground (i.e. the height of the heaters). Another consequence of the model is the inability to replicate the melt from outside of the heated plots that was likely occurring, and providing more mass losses to melt/runoff. Therefore, the amount of water availability in heated plots may be underestimated by the model.

Other studies of winter fluxes beneath forest canopies suggest smaller evaporative losses relative to open areas (Harding et al., 2002). This is because forested areas are sheltered from wind and receive less radiation resulting in a 23% decrease in sublimation rates (Montesi et al., 2004). Increasing air temperatures associated with climate change will cause an increase in absolute humidity (IPCC, 2007). While the model cannot exactly replicate the processes occurring in heated plots, the model does suggest that additional LW in from heaters may warm the snow surface enough to cause a reverse gradient in sensible heat and latent fluxes between the snow surface and atmosphere relative to control plots. Similarly, Molotch et al. (2009) performed a study on snowmelt partitioning for two different years at two different sites. Results found warmer air temperatures through mid winter and early spring increased latent heat fluxes by 28% resulting in greater water partitioning into evaporative loss and reduced spring and
summer flows. Conversely, they found that a late season rapid snowmelt in one year resulted in less evaporative loss relative to the other site with slower snowmelt. This study exceeds those latent heat losses by double the amount. In heated plots, 48% to 67% more mass was lost to evaporation/sublimation suggesting decreased water availability in a warmer climate.

2.5.2 USA-ALP/Saddle

While previous works on Niwot Ridge have found net radiation to account for 46 and 75% of energy available for snowmelt in the alpine during two consecutive years (Cline, 1997a,b), this study reports 100% in the control plots at the USA/ALP. Other studies have found net radiation was the primary source of energy for snowmelt in alpine environments (De la Casiniere, 1974; Marks and Dozier, 1992). Differences in turbulent energy exchange are likely due to different weather conditions during the two years, as seen at the LSA site as well. Cline (1997b) found that a year that experienced greater energy contributions from turbulent fluxes experienced 1.3 °C higher air temperatures and a higher mean specific humidity relative to the year that net radiation accounted for more energy available to melt.

Sublimation losses are associated with the magnitude and direction of turbulent fluxes. In alpine environments, sublimation losses to high winds are considerable (Berg, 1986; Kattelmann and Elder, 1991; Strasser et al., 2008). Model mass losses to sublimation of 33% of total SWE fits well within the reported range of 15 to 50% found on Niwot Ridge in previous years (Berg, 1986; Hood et al., 1999).

In heated plots when snow was below the heaters (100cm), model output illustrated little differences in LWnet and sensible and latent heat fluxes relative to control plots until the last 24 hours of melt. During the last stages of melt in the first model run that was initialized with 100
cm of snow, latent and sensible heat became negative and a larger portion of snowmelt was lost to evaporation/sublimation versus melt/runoff compared to control plots (83% vs. 38%). Similar to what was observed at the LSA, heater’s warmed the snow surface to 0° C causing a sensible heat flux away from the snow surface relative to the cooler atmosphere. Additionally, model results illustrate grain sizes as large as 4 mm at the surface of the heated plot. As snow grain size increases, the amount of IR radiation absorbed increases. Thus, a positive feedback loop may be occurring whereby warmer air created by heaters increases grain size which further increases the absorption of IR energy radiating from the heaters relative to control plots. Another potential mechanism responsible for greater turbulent flux loss with less than 10 cm of snow present during this time (in the model), greater soil exposure to heater’s energy caused a greater partitioning of energy into evaporation of wet soils from infiltrated snowmelt.

When the model run incorporated snow depths from the entire season, essentially no difference in the partitioning of energy between heated and control plots was observed. Net radiation accounted for 100% of energy available for melt and sublimation losses were within 16%. These findings are likely attributed to high winds which significantly increase turbulent fluxes and dissipate the amount of heat reaching the snow surface from the heaters (Kimball, 2005). Overall, results suggest an inefficient delivery of heater’s energy when snow is beneath the heaters and wind speeds are high causing little or no warming of the snowpack.

On the other hand, when snow exceeded heater height, modeled turbulent fluxes were much smaller relative to control and similarly, net radiation accounted for 100% of energy available for melt. The lack of wind coupled with warmer air temperatures resulted in stable conditions and very low turbulent fluxes (around 0 Wm⁻²). Similarly, Cline (1997a) found a dampening of turbulent fluxes during a year when air temperatures and humidity were higher.
He also found that this resulted in net energy gains at night due to sensible heat gains that compensated for LW losses causing faster snowmelt compared to the previous year with cooler temperatures and net energy losses at night. These findings support our modeled and observed results that showed heated plots experienced earlier and faster snowmelt. Furthermore, most mass was lost to melt/runoff suggesting greater amounts of available water in heated plots relative to control. These results support our field observations that illustrated liquid water during the winter and associated higher soil moisture during the spring.

As previously discussed, our model does not have the ability to exactly simulate heated plots due to the location of additional LW input from the heaters at 1.2 m above the ground. Modeling the snowpack at the USA-ALP once the snow exceeds the heaters and caves form is even more difficult. Inside caves, relative humidity should be high, air temperatures warm, and winds low or non-existent. As a result, model simulations fail to replicate surface hoar that was observed on the cave walls. Surface hoar forms from condensation due to large amounts of water vapor caused by the heaters. When heaters turn off in winds, the air temperature cools and the water vapor condenses. This process should result in large energy and mass gains due to the release of latent heat. Additionally, field observations of large amounts of melt were apparent as indicated by ice lenses and soil moisture data but were not reproduced by the model.

While the model cannot exactly replicate what is exactly going on in heated plots, results do provide insight into potential feedbacks to climate change. Current climate conditions show sensible heat gains to the snowpack during the winter. Increasing air temperatures would be expected to increase this positive gradient. However, increased incoming longwave radiation may counteract this warming affect by warming the snow surface thus maintaining the current temperature gradient that generally exists between the snow and atmosphere. Furthermore, latent
heat fluxes can be both positive and negative from the snow surface to atmosphere depending on atmospheric conditions. If air temperatures increase with climate change, latent heat exchange over the snow surface will likely increase. However, our model results show little difference in energy fluxes between the snow surface, emphasizing the role of high winds on increasing turbulent exchange. Another aspect of warming that is unclear is the effect of a warmer climate on winds/atmospheric circulation trends. Will a warmer, more stable atmosphere decrease wind speeds? If so, low winds associated with caves suggest lower turbulent fluxes and smaller water losses to evaporation/sublimation and greater water availability in alpine environments.
2.6 Overview

Model results illustrate several differences in energy exchange and snow properties between heated and control plots during various snow conditions and elevations as a result of heaters. Figure 2.25 summarizes these effects on energy and mass fluxes. At the LSA, heaters warmed snow surface causing large losses to sensible and latent heat. And at the USA-ALP, small differences were apparent in energy exchanges between heated and control plots which is largely attributed to high winds (Figure 2.25). Thus, our results suggest we are not warming plots as efficiently as we thought. On the contrary, when snow exceeds heater height and caves were present, results suggest greater heater efficiency. Alternatively, heaters may act to increase snow grain size at the surface from warmer air, which effectively increases the amount of IR radiation absorbed at the surface.
Figure 2.26: Diagram illustrating the influence of heaters on energy and mass fluxes in different conditions (e.g. low/high winds, snow depth) at the LSA (upper panel) and USA-ALP (lower panel). Length of arrows indicates relative magnitude differences between fluxes.

With the exception of caves, our study suggests that more evaporation/sublimation occurs in heated plots, thus providing less available water compared to control plots. Because the model cannot replicate the exact mechanism of infrared heaters, we are unable to state whether or not heated plots have more or less SWE relative to control plots. Soil moisture data illustrates
greater amounts of soil moisture during the middle winter. However, at the LSA these values quickly decrease after the melting of snow events suggesting greater amounts of evaporation.

Lastly, results illustrate the importance of considering the contribution of each energy balance component: Radiative, turbulent, and ground heat flux in the interpretation of warming experiments. These fluxes significantly alter the water balance. Furthermore, the relative contribution of these fluxes will likely be altered by increasing air temperatures, longwave radiation, and increased humidity associated with climate change predictions. This study emphasizes the importance of the snow surface energy balance in warming experiments for interpretation of and providing insight into how ecosystems may respond to climate change.
References


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