SUBALPINE SNOWPACK-CLIMATE MANIPULATION AND MODELING EXPERIMENT AT NIWOT RIDGE, CO AND VALLES CALDERA NATIONAL PRESERVE, NM

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ABSTRACT

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Subalpine snowpack-climate manipulation and modeling experiment at Niwot Ridge, CO and Valles Caldera National Preserve, NM

Thesis directed by Professor Noah P. Molotch

At Niwot Ridge, CO, a global warming experiment using near-infrared (IR) heaters is being conducted. Investigation of snow accumulation, snowmelt, and soil microclimate found the heaters to influence these variables at the subalpine experimental site. These changes were compared to an environmentally similar yet naturally warmer subalpine snowpack in the Valles Caldera National Preserve, NM. Over the 2010 and 2011 snow seasons, snow accumulated 42% lower on average and melted out 1-37 days earlier in the warmer plots (CO heated and NM) compared to the CO controls. Soil temperature was 2.6 °C greater on average in warmer plots compared to controls. Peak soil moisture was 0-12% lower in warmer plots versus controls. In order to estimate differences in energy and mass balance exchange at the snow surface in control versus warmer plots, the one-dimensional, physically based snowmelt model, SNOWPACK, was used. Energy and mass fluxes in control cases were compared to heated, NM, and synthetically warmer cases. Model results found that the heaters alter radiative, turbulent and mass fluxes by amounts comparable to differences between CO and NM fluxes. The sign and magnitude of energy and mass exchanges were similar between the control and synthetic models. The proportion of the energy flux associated with latent heat was 11-30% of the total energy flux in heated and NM models compared to 2-9% in control and synthetic models during snowmelt, and subsequently the associated evaporative loss to the atmosphere was much lower in control cases relative to NM cases. These results indicate that warmer conditions (i.e. increased average air

temperatures) as projected in the coming decades will change snow surface energy fluxes, timing and magnitude of snow accumulation and melt, and soil temperature and moisture. Results of this study aid in the interpretation of climate manipulation experiments and modeling as they pertain to snowpack, and contribute to a better understanding of the interactions between climate, hydrology, and ecological processes.

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1. INTRODUCTION

The complex relationships between earth surface processes and hydrologic systems and are not fully understood, especially in the context of a changing climate. Many experiments are currently being conducted in order to render a clearer understanding of what is likely to change in the coming decades and centuries and what these changes will mean for the world. Particularly, Arctic and alpine regions are among the most dramatically affected by changing climate conditions (IPCC, 2007). Changes in snowmelt under shifting climate conditions are especially significant concerns for the one-sixth of the world's population depending on snowcovered glaciers and seasonal snow cover for their water supply (Barnett et al., 2005). The geography of the mountainous western United States is such that the snowpack is particularly sensitive to temperature changes of just a few degrees since average winter temperatures in many areas are close to freezing (Bales et al., 2006). Therefore, a relatively small temperature increase of 3-5 °C—which is well within the predicted range for 2100 temperatures (IPCC, 2007)—can produce a phase change of solid (snow) to liquid (rain) precipitation resulting in major consequences for the environment and end-users of this mountain precipitation (Adam et al., 2009; Bales et al., 2006). Shifting snowpack processes are particularly difficult to quantify and plan for due to the great spatial and temporal variability inherent in alpine and subalpine snow ecosystems (Bloschl, 1999; Meromy et al., 2012).

Several mountain snowpack characteristics including snow cover persistence, snow cover extent, snow water equivalent (SWE), and the timing and magnitude of snowmelt pulses are expected to change with shifts in climate (Adam et al., 2009). One study comparing six regional climate models for the years 2071-2100 has shown that in an alpine basin of the Swiss Alps, the snow season is projected to shorten by an average of one month in the beginning of winter and

by one and a half months at the spring end. Additionally, maximum SWE values may decline by an average of 27% (Magnusson, Jonas et al. 2010). In snow-dominated ecosystems, the timing of the snow season largely controls plant growth. Based on predictions of decreasing SWE and shorter snow season length, mountain grasslands can be expected to produce more biomass in the future (Jonas et al., 2008). Vegetation in alpine regions have already been observed to respond to increasing temperatures by migrating upslope (Grabherr et al., 1994). Additionally, increasing temperatures have been linked to seasonal soil moisture stress in high-elevation, water-limited systems which greatly limits plant species' ability to establish and survive (Moyes et al., 2012). All of these changes to mountain snowpack are expected to have wide-ranging impacts on many interacting biogeochemical, ecological, atmospheric, and human systems.

Climate manipulation experiments and land surface modeling are techniques often used to better understand what may happen as aspects of snow hydrology change. Climate manipulation experiments in seasonally snow-covered regions have historically included snow fences to increase snow depth on the leeward side (Williams et al., 1998), shoveling to change snow distribution, and open top chambers (OTCs) (Wipf and Rixen, 2010) among many others. OTCs are passive, open-top devices frequently used to increase temperature in different environments. The effects of these perturbations on the surrounding ecosystem are then examined (Marion et al., 1997). Previous methods used in experimental warming are considerably different from more "active" methods including infrared (IR) heating lamps (Harte et al., 1995).

This more "active" manipulation method—the use of thermal radiation via IR heaters to warm experimental plots—is an attractive perturbation method because it simulates global warming and appears to have the fewest limitations (Harte et al., 1995; Kimball et al., 2008).

Several others have studied ecological responses to climate change using IR heaters (Adler et al., 2007; Harte et al., 1995; Wan et al., 2002), however only one other study has used IR heaters to specifically examine snowpack properties and an altered energy balance in the context of global warming (Petrzelka, 2011).

The University of California at Merced is currently conducting an experiment using IR heaters as part of the Alpine Treeline Warming Experiment (ATWE) on Niwot Ridge in the Colorado Rocky Mountains. The ATWE consists of experimentally heating plots of land to simulate warming conditions expected in the coming decades, and observing the warming effects on subalpine and alpine plant species (ATWE, https://alpine.ucmerced.edu/pub/htdocs/). The UC Merced team is addressing questions concerning effects of climate warming on subalpine and alpine species distribution (Moyes et al., 2012), plant stress levels, and ecosystem properties. Climate variables controlling snow properties and melt timing such as air temperature and precipitation have the greatest direct effects on alpine plant height and growth rates (Jonas et al., 2008), thus highlighting the importance of quantifying changing precipitation and temperature to examine resulting changes in vegetation.

This thesis will address changing snow accumulation and melt dynamics and soil microclimate in a subalpine area of the Colorado Front Range in the context of projected increasing surface air temperatures by integrating snow manipulation experiments over two snow seasons and comparing results to a naturally warmer subalpine snowpack in New Mexico. Potential changes of surface energy and mass fluxes will be examined using a snowpack model. The driving questions behind this research include:

1. How will future increases in air temperature change snow accumulation, snowmelt, and soil micro-climate?

2. How will future increases in air temperature change the partitioning of energy at the snow-atmosphere interface?

3. How do these snowpack-climate warming experiments compare with current projections of future snowpack conditions?

Section II below provides background on climate manipulation experiments, the snowpack surface energy balance, and snowpack modeling. Section III describes the two study areas. Section IV details the field and modeling methods. Field and modeling results are described in Section V. Section VI includes a discussion of results in the context of underlying assumptions and uncertainties. Conclusions are summarized in Section VII.

II. BACKGROUND

Climate Manipulation Experiments

Recent changes in global climate have been observed over the past few decades, and these changes are expected to continue on a similar trajectory over the coming decades (IPCC, 2007). These changes affect various ecosystems characteristics, and thus climate manipulation experiments have become useful tools for investigating consequences of changing climate conditions on ecosystems. A recent Intergovernmental Panel on Climate Change (IPCC) report has found that global warming may cause an increase in radiative forcing of 4-9 W m⁻² by the year 2100 corresponding to an increase in average global air temperatures of 1.5 - 6 °C. This increase in air temperature will likely have dramatic effects on mid-latitude mountain ecosystems—regions of high climatic sensitivity as average winter temperatures are already

close to freezing in many areas, and a small increase in air temperature would bring many areas above freezing (Bales et al., 2006). To address a range of unknowns regarding increased average air temperatures, gradient and experimental methods have been frequently used in ecological climate change research (Dunne et al., 2004). Specifically, the possible changes to a snowmeltdominated ecosystem have been investigated using various passive and active snowpack manipulation experiments.

Numerous passive climate manipulation methods have been employed by researchers. Since snowpack has been observed (Clow, 2010; Laternser and Schneebeli, 2003; Mote, 2003; Mote et al., 2005) and predicted (Adam et al., 2009; Gillan et al., 2010; Lapp et al., 2005) to change in many mountainous regions, snow fences have been used to simulate increased snow cover on the leeward side, and resulting ecological changes analyzed. One long-term study reporting initial short-term results showed that the deeper snowpack on the leeward side of the snow fence increased winter soil temperatures and CO₂ flux (Walker et al., 1999). Another study found the maximum N₂O flux to increase three-fold behind the snow fence in addition to greater CO_2 production due to a deeper, more insulating snowpack (Williams et al., 1998). Another long-term (> 32 years) study employing a snow fence in the Southern Alps showed changes in vegetation composition with canopy roughness higher on less-exposed leeward side with the deeper winter snowpack (Smith et al., 1995).

Other studies, which are focused on investigating possible effects of warming ecosystems, have employed other passive techniques such as open-top chambers (Aerts et al., 2006; Marion et al., 1997), nighttime warming (Llusia et al., 2006), and field greenhouses (Hobbie and Chapin, 1998). Nighttime shading curtains work by means of slowing down relative heat loss as air temperatures drop at night, or protecting the surface from boundary layer

turbulence thus retaining surface heat. Greenhouses and OTCs work by allowing in shortwave radiation while decreasing longwave radiation (infrared or IR) escape. Some limitations for these methods include complexity of design and cost for the shading curtains, and difficulties controlling the gas concentrations and air temperatures inside greenhouse structures and OTCs (Aronson and McNulty, 2009).

Given these limitations, active methods have become commonly used in climate manipulation warming experiments with varying success. Active methods use an external heat source such as heated cables or IR heat lamps to warm the environment. Heated cables are placed above, on, or buried below the surface, and temperature monitors are linked to an automated, computer-based regulation system (Grime et al., 2000). However, this method can greatly disturb surface plants and organisms due to heat conductance. This may propagate decoupling between above and belowground environments thus becoming a less realistic climate warming experiment (Aronson and McNulty, 2009).

Overhead IR heaters are another active climate manipulation warming experiment method. Using a temperature free-air controlled enhancement (T-FACE) system, IR heaters suspended above a surface allow for surface heating while blocking very little (< 2%) of incoming electromagnetic radiation (Price and Waser, 2000). The heaters can be run in two different modes: constant flux or constant temperature rise mode. The former outputs a constant amount of energy at a set power level (Harte and Shaw, 1995; Petrzelka, 2011) while the latter actively regulates the energy output from the heaters to maintain a constant temperature difference between heated and control plots (Kimball et al., 2008; Nijs et al., 1996). One limitation regarding IR heaters in constant flux mode is that with increasing wind speeds, heater efficiency decreases drastically from a theoretical maximum of 52% in still air to 4.1% in wind

speeds of 10 m s⁻¹ (Kimball, 2005). Without regulating heater output relative to surrounding temperatures as in constant temperature rise mode, the amount of energy reaching the surface from heaters will vary much more than is realistically expected in the future (Kimball, 2005).

Another limitation of IR heating is that the vapor pressure gradient (VPG) from inside leaf stomata to the air would not match with what is actually expected to happen since the air may not also be warmed (Amthor et al., 2010; Kimball, 2005). For example, an experimental increase in air temperature without a parallel increase in specific humidity will result in a lower relative humidity in the warmed environment. Subsequently, the VPG would increase and result in an unintended moisture experiment. However, Kimball (2005) proposed a first order correction: to simply water the plants by an amount enough to offset the confounding effects of drying. He derived an equation quantifying how much water an infrared-warmed plant would lose in normal air compared to how much it would have lost in air which had been warmed at constant relative humidity. That lost amount of water could be added to the surface to address this problem (Kimball, 2005). A study using IR heaters over spring wheat has utilized this adjustment successfully (Wall et al., 2011).

IR heating is currently considered by some to be the most realistic method to simulate a warmer climate as it appears to cause minimal disturbance to an ecosystem, has moderate temperature variability, and can be applied in a variety of different environmental conditions (Aronson and McNulty, 2009). However, this claim is contested by some who point out that infrared heating is not the mechanism by which environments will warm, since in reality, global warming should heat the air, not only the surface, while maintaining a constant relative humidity (Amthor et al., 2010). Though this assertion is based on an understanding of basic principles of heat transfer, one study that found that IR heaters do, in fact, cause increases in minimum and

mean air temperatures (Wan et al., 2002). Regardless of whether or not IR heaters warm the surface and not the air, Kimball (2011) illustrates that IR heaters still relate to climate warming in a quantifiable way since canopy and soil temperatures can still be made to rise as would be expected with global warming (Kimball, 2011), and thus they remain an effective method to study ecosystems in the context of global warming. Though several studies using IR heaters have been employed to examine effects of climate warming on vegetation and soil in various environments (Harte et al., 1995; Kimball et al., 2008; Nijs et al., 1997; Wall et al., 2011; Wan et al., 2002), this is the only study using IR heaters specifically for a snowpack-climate manipulation (Petrzelka, 2011). One potential issue relating specifically to IR heating of snowpack is that if the air temperature is not increased while the snowpack temperature is raised (up to a maximum of 0 °C), then the temperature gradient of the snowpack is changed relative to control plots. On the other hand, if air temperature is, in fact, increased by the heaters while the snowpack remains at 100% relative humidity, then the VPG will increase relative to control plots since warmer air can contain more moisture. These aspects of IR heaters must be considered in the context of snowpack-climate manipulations.

Snowpack Surface Energy Balance

Snowpack energy is mainly partitioned into radiative and turbulent fluxes which govern energy and mass exchange at the surface. An increase in energy to the surface (i.e. increased near-surface air temperatures) is expected based on estimates from several climate models (IPCC, 2007). With more energy in the system, the energy exchange at the snow-atmosphere interface may undergo a shift in partitioning of these surface fluxes, possibly changing when and how mass (water) is gained and lost. In this study, energy and mass fluxes are reported from the perspective of the snowpack, thus gains to the snowpack energy and mass are described as

positive, losses from the snowpack are described as negative. The snowpack energy balance is generally expressed:

$$\Delta Qs = SW_{net} + LW_{net} + Q_e + Q_h + Q_g + Q_A$$
(1)

where ΔQ_s is the change in energy storage or cold content, SW_{net} , LW_{net} , Q_e , Q_h , Q_g , and Q_A are the net shortwave radiation, net longwave radiation, latent heat flux, sensible heat flux, ground heat flux at the snow-soil interface, and energy flux due to advection, respectively. When ΔQ_s is negative, the snowpack is cooled, increasing its cold content (the energy required to warm the snowpack to 0°C). A positive energy balance warms the snowpack up to 0°C at which point melt can begin, and then ΔQ_s is energy for melt (Marks and Dozier, 1992). SW_{net} is a positive flux to the snowpack, since the albedo (measure of reflectiveness) is generally between 0.4-0.95, depending on grain size (Male and Granger, 1981). LW_{net} and Q_e are generally negative fluxes (losses) depending on atmospheric and snowpack characteristics (Marks and Dozier, 1992). However, LW_{net} can be positive during snowmelt depending on canopy thermal emission (Sicart et al., 2004). Q_h is usually positive due to the temperature gradient between the warmer air and colder snow (Marks and Dozier, 1992), though nighttime cooler air can reverse the temperature gradient so that Q_h would be negative (DeWalle and Rango, 2008). Q_g and Q_A are generally less significant relative to the other terms, and so are not examined in detail here.

Measurement and estimation of the different snowpack energy balance components can be difficult in spatially variable, remote regions with low temperatures and high windspeeds. The magnitude and rate of turbulent fluxes depend on air temperature, relative humidity, windspeeds (Male and Granger, 1981) as well as surface topography (Pohl et al., 2006). Several studies have found that radiative fluxes dominate the energy balance during snowmelt accounting for roughly 60 - 98 % of the energy for melt (Cline, 1997; Link and Marks, 1999; Marks et al., 2008). Male and Granger (1981) compared results from 15 studies showing that on average, net radiation accounts for 59% of total snowpack energy exchange. However, the magnitude of the energy balance components vary from site to site and between years with different atmospheric conditions (Molotch et al., 2009). Pohl et al. (2006) found that turbulent fluxes contribute between 30-50 % of the energy for snowmelt while Link and Marks (1999) found a turbulent flux average contribution of 11%.

Subalpine forests introduce further complexity for measuring or estimating turbulent fluxes at the snow surface since vegetation exerts a significant influence over the surface energy balance (Musselman et al., 2008; Sicart et al., 2004). Vegetation significantly alters not only radiative exchanges by absorbing and reflecting incoming solar radiation ($0.28 - 2.8 \mu m$), altering the emission of longwave radiation ($2.8 - 100 \mu m$) (Male and Granger, 1981), and changing snow surface albedo due to organic debris deposition (Melloh et al., 2001), but also lowers turbulent energy exchange efficiency at the snow or soil surface by sheltering the snow surface from wind and altering temperature and humidity gradients (Davis et al., 1997; Harding and Pomeroy, 1996). The difficulty in determining these contributions to snowmelt energy, particularly in forested subalpine ecosystems, is further compounded by the uncertainty of these energy fluxes with respect to shifting climate conditions.

One technique for turbulent flux measurement is the eddy covariance method. Though the eddy covariance method is a relatively reliable way to measure above-canopy turbulent fluxes (Turnipseed et al., 2003), it is not universally applicable as it requires delicate, expensive instrumentation.

Snowpack Modeling

Physically-based modeling offers an approach to estimate these fluxes as well as other energy balance components when direct measurements are lacking or when predicting future snowpack conditions. Numerous physically-based snow models have been developed and validated in several locations for various applications (Anderson, 1973; Boone and Etchevers, 2001; Koren et al., 1999). Several more complex snow models which explicitly treat snow grain size, different layers, and snowpack metamorphism have been recently developed including SNTHERM (Jordan, 1991), CROCUS (Brun et al., 1992), and SNOWPACK (Bartelt and Lehning, 2002; Lehning et al., 2002a; Lehning et al., 2002b). These types of models can be used to examine difficult-to-measure snowpack properties as well as simulate future snowpack conditions with "future" input parameters derived from regional climate modeling scenarios. Modeling future snowpack conditions with a greater degree of confidence can help water managers plan for future availability and allocation of resources.

III. SITE DESCRIPTIONS

Niwot Ridge, Colorado

Niwot Ridge is located in the Colorado Front Range 35 km west of Boulder, CO and ~8 km east of the Continental Divide (40° 03' N, 105° 36' W) (Figure 1).



Figure 1: Niwot Ridge, CO and Valles Caldera National Preserve, NM site locations.

This area is designated a UNESCO Biosphere Reserve and is the location of the National Science Foundation (NSF) Niwot Ridge (NWT) Long-Term Ecological Research (LTER) site. The UC Merced ATWE study sites are located within the NWT LTER site. The ATWE consists of 80 3-meter diameter plots spread between three sites along an elevational gradient; The LSA (lower subalpine) site at 3080 m was established at the lower limit of Engelmann spruce and limber pine tree distributions with 20 plots (Figure 2). Alpine Treeline Warming Experiment LSA site overview



Figure 2: Map of ATWE plots and treatments at each plot. Map adapted from Kueppers et al. (unpublished data).

At the upper limit to these species' distribution, another 20 plots were placed at the USA (upper subalpine) site at 3367 m. The remaining 40 plots were set up at the ALP (alpine) at 3517 m. For this study, I focus on subalpine snowpack dynamics, thus I incorporate only results from the LSA site into my analysis.

The LSA is located in a relatively dense, subalpine forest dominated by lodgepole pine (*Pinus contorta*) with some Engelmann spruce (*Picea engelmannii*) and subalpine fir (*Abies lasiocarpa*). Average canopy height is 11.4 m, and maximum LAI during the growing season is 4.2 m² m⁻² (Turnipseed et al., 2003). Soils at Niwot Ridge are extremely rocky (granitic parent material) consisting mainly of mineral clays covered with a ~10 cm surface layer of organic material (Turnipseed et al., 2002).

NWTLTER collects several meteorological, snowpack, and flux measurements at the nearby C-1 site, the National Resources Conservation Service SNOwpack TELemetry

(SNOTEL) network site NIWOT 663, and at the AmeriFlux tower all within 0.5 km from LSA. Long-term (1952-present) mean annual air temperature is 1.5° C. Niwot Ridge experiences an average of 930 mm annual precipitation, about 80% of which falls as snow (Caine, 1995).

Valles Caldera National Preserve, New Mexico

The 36,000-ha Valles Caldera National Preserve (VCNP) is located in the Jemez Mountains of northern New Mexico (35° 53' 18" N; 106° 31' 36" W) (Figure 1). The vegetation at this mixed-conifer study site is mainly composed of Douglas fir (*Pseudotsuga menziesii*) white fir (Abies concolor), blue spruce (*Picea pungens*), southwestern white pine (*Pinus strobiformis*), limber pine (*Pinus flexilis*), and ponderosa pine (*Pinus ponderosa*), and some aspen (*Populus*) tremuloides). Average canopy height around the flux tower is 19.6 m, and growing season LAI is $3.43 \text{ m}^2 \text{ m}^{-2}$ (McDowell et al., 2008). Soil texture at this site is a sandy loam, roughly 150 cm deep overlying volcanic parent material (Small and McConnell, 2008). Average annual precipitation is 720 mm. Supplementary precipitation data were collected at the nearby Quemazon SNOTEL site at an elevation of 2896 m. Approximately 65% of the precipitation in the Jemez mountains falls as snow between November and April and 35% falls as rain associated with the summer monsoon. Based on SNOTEL data from both CO and NM sites, precipitation and air temperature track each other fairly well reflecting generally similar regional meteorological patterns. Average temperatures at the Quemazon site were consistently 3-4 °C higher than the Niwot SNOTEL site in both 2010 and 2011 (Figures 3 and 4) suggesting these naturally warmer and artificially warmer subalpine sites are ideal to compare for evaluating the effects of a warming climate on snowpack dynamics.



Figure 3: Daily mean air temperatures at the SNOTEL sensors near the two sites track each other closely indicating similar regional-scale weather patterns.



Figure 4: Air temperatures during the snow season (roughly November – June) are highly correlated at the two sites, Pearson's $\rho = 0.92$ in both years.

IV. METHODS

Field Methods

Experiment Design

On Niwot Ridge during the 2008-2009 field season, IR heaters were installed for the ATWE over the experimental plots spaced at least 3 m apart. At each site, 25% of the plots were heated, 25% were watered, 25% were heated and watered, and the remaining 25% of plots are controls. The watering is meant to counter the probable soil drying that occurs with heating (Kimball, 2005) in order to distinguish biological effects of increased temperature from soil moisture changes (Kueppers and Harte 2008).

The IR heater plot design consists of a hexagonal steel structure with six infrared lamps spaced evenly around each of the 3 m diameter (7.07 m^2) plots 1.2 meters above bare ground (Figure 5).



(Petrzelka, 2011)

Figure 5: Heated and control plot schematic. Each plot was divided into quadrants, so soil moisture and temperature values were averaged for each plot.

The lamps are aimed at 45° angles to apply the radiation inward toward the center of the plot and as evenly as possible (Kimball, 2005; Kimball et al., 2008). Unlike the previous snow manipulation methods, this design allows for minimal wind interference, sunlight interference, and moisture redistribution thus more closely approximating potential warmed conditions.

The 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. The energy emanating from heaters was in the longwave portion of the electromagnetic spectrum at 4.5-42 μ m. From a surface energy balance perspective, the energy from the heaters was equivalent to an increase in incoming longwave radiation (L_{in}). In order to achieve a surface soil temperature increase target of ~4-4.5 °C averaged over the growing season, heaters were operated in a constant flux mode following the protocols of Harte et al. (1995) and Harte and Shaw (1995). Since soils at Niwot Ridge are wetter during the growing season than the montane meadow plots of Harte et al. (1995), more than double the radiation is needed to double the temperature effect because more of the increased radiation energy is used to evaporate water.

During the 2009-2010 season, heaters operated at 50% power so that each heater emitted 500 W (Figure 6).



Heater output percentage

Figure 6: Timing of heater output values. For the entire snow season of 2009-2010, heater output was 50% or an additional 214 W m⁻² to each heated plot. In 2010-2011, output varied between 10 %, 20 %, and 40% creating an additional radiant flux density to each heated plot of 42, 85, and 171 W m⁻², respectively.

It is estimated that ~50% of the output is lost outside of the plot (Kimball, 2005), reducing the 500 W to 250 W output. Since each heated plot was roughly 7 m² and assuming heat was evenly distributed, the radiant flux density in each plot was about 214 W m⁻² or 1.85 x 10^7 J m⁻² d⁻¹. Heaters were programmed to temporarily turn off in high winds (>15 m s⁻¹).

During the 2010-2011 winter season (mid-November through early June), heater output was lowered to 10 % - 20% power because heating efficiency decreases substantially in high winds (particularly at the higher elevations sites USA and ALP). Additionally, high heater output in winter can lead to hydrologic artifacts such as depressions in the snowpack regularly refilled by wind-redistributed snow resulting in a much greater moisture flux in the heated plots (Petrzelka, 2011). Thus during winter, each heater emitted 100 W while at 10% power giving an additional radiant flux density of 42 W m⁻² in each heated plot. In mid-March, heaters were turned up to 20 % in order to achieve an advance in the timing of snowmelt. A 20% heater output resulted in an additional radiant flux density of 85 W m⁻² at each heated plot. Heaters were turned up to 40 % output in June (Figure 6) producing a radiant flux density of 171 W m⁻². During this snow season, heaters were programmed to temporarily turn off in wind speeds of > 10 m s⁻¹. *Soil measurements*

Soil moisture or volumetric water content and soil temperature was measured with Decagon EC-TM sensors (5–10 cm depth). Sensors were located in the center of each quadrant of the plot (~80 cm from the corner of each quadrant), and averages of the four sensors in each plot were used to calculate average soil moisture per plot. Located in approximately the center of the LSA site was a meteorological instrument cluster consisting of a 3 cup anemometer (RM Young Model 03101-L) measuring wind speed, and a Campbell Scientific HMP45C-L measured temperature and relative humidity. All meteorological sensors were mounted 3 meters above bare-ground. Microclimate data were logged as 15 min averages of 1-s readings. Additionally, above-canopy radiation (25.5 m), temperature and windspeed (both 21.5 m) measurements were obtained from the Niwot Ridge AmeriFlux tower (Net Radiometer CNR-1, Kipp & Zonen, Vaisala HMP-35D, and CSAT-3 Campbell Scientific Sonic Anemometer).

Snow measurements

Snow measurements over heated and control plots at Niwot Ridge were made using Judd Communications LLC Ultrasonic snow depth sensors. These became operational in March of 2010, and for the 2010-2011 snow season, in late January 2011. Each snow depth sensor was 8 x 8 x 13 cm in size with a beamwidth of 22° and a vertical accuracy of 1 cm. Depth was recorded hourly with Campbell Scientific CR1000 and CR10X dataloggers. Each snow depth sensor was located about 2 m above bare ground so as to avoid being buried by the seasonal snowpack. In 2010, control sensors were mounted on the end of a 1-2 m boom extending outward from its paired heated plot. In 2011, control sensors were mounted on separate structures over control plots adjacent to heated plots. In both years, sensors in heated plots were 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. During both 2009-2010 and 2010-2011, snow depth in each plot was measured by hand about every two weeks, starting in November 2009 and November 2010.

Snow pits at the C-1 pit site (within 0.5 km of LSA site) were dug and sampled weekly for physical and chemical parameters as part of the NWT LTER project. Snow pits 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 as was the thickness and type of layer (e.g. buried sun crust, ice lens, coarse to fine grain transition), and the grain type and grain size of each layer was recorded. Additional snow pits were excavated next to a few selected heated plots for more direct comparison of snow properties in 2010; however regular snow pits were not dug within the plots due to concerns that such destructive sampling would interfere with the experimental treatments.

For the 2009-2010 snow season, four pairs of heated and control plots were set up at LSA. The 2010-2011 snow season had two heated and control pairings. The fewer numbers of pairs in 2010-2011 were due to field equipment and electrical malfunctions.

The Valles Caldera Mixed-Conifer instrument cluster is located on the northeastern side of Redondo Peak with snow depth and eddy flux observations at 3020 m. Nine Judd ultrasonic snow depth sensors were installed in the vicinity of the flux tower beginning data collection over the 2004-2005 snow season and continuing through the end of the 2009-2010 snow season. The sensors were positioned in a stratified sampling pattern regarding proximity to trees with three sensors in under-canopy locations, three at canopy-edge locations, and three in open areas (Molotch et al., 2009). Average snow depth data were used in this study's comparison. Snow depth and SWE in 2011 were measured at a nearby COSMOS (Cosmic Ray Soil Moisture Observing System; http://cosmos.hwr.arizona.edu/) site approximately 300 m from the flux tower (Desilets et al., 2010). Meteorological measurements were collected from instruments at the flux tower as described in Molotch et al. (2009). For 2010 soil microclimate data, soil moisture and temperature were recorded at 10 and 40 cm below ground adjacent to the flux tower footprint using a Campbell Scientific CS-615 water content reflectometer and a Campbell Scientific TCAV instrument, respectively. Soil microclimate data for 2011 were collected using Decagon 5TE Water Content EC and Temperature Sensors from the Jemez River Basin Mixed-Conifer Zero Order Basin site ~2 km SW of the Mixed-Conifer instrument cluster in an area of similar elevation and canopy cover (JRB CZO, http://www.czo.arizona.edu/JemezRiver.html).

Modeling Methods

Snowpack model

SNOWPACK, a complex, one-dimensional physical snowpack model developed at the Swiss Federal Institute for Snow and Avalanche Research (SLF), was used to model snowpack, mass, and energy fluxes from heated and control plots within UC Merced's ATWE, a synthetically warmer snowpack, and the VCNP, NM snowpack.

SNOWPACK numerically solves the partial differential equations governing mass, energy, and momentum conservation within the snowpack using a fully implicit Lagrangian Gauss-Seidel finite element method. Snow is modeled as a three-component porous structure comprising volumetric fractions of ice, water and air. Snowpack behavior is characterized by two mass conservation equations for the vapor and water phases, and one bulk temperature equation and one momentum equation for the ice phase (Bartelt and Lehning, 2002). Rates of snowpack settlement and explicit properties of multiple layers are calculated based on a microstructuredependent viscosity determined by the temperature and temperature gradient of the snowpack (Lehning et al., 2002b).

Model forcings

SNOWPACK is driven by several meteorological and physiographic variables. Required input variables include: air temperature, relative humidity, wind speed, snow-soil interface

temperature, incoming shortwave and longwave radiation, precipitation and/or snow depth,

geographic locational information, vegetation density, and soil properties. The instruments,

locations, and measurement heights for each variable are described in Table 1.

Niwot Ridge, Colorado			Ι	/alles Caldera, New Mexico
	Measurement		Measurement	
Variable	Height (m)	Instrument	Height (m)	Instrument
Temperature		Vaisala HMP-45 &		
(°C)	2 & 21.5	Vaisala HMP-35D	21.65	CSAT-3 Campbell Scientific
Relative		Vaisala HMP-45 &		
Humidity (%)	2 & 21.5	Vaisala HMP-35D	21.65	Vaisala HMP-45C
		RM Young 03101-L		
Wind Speed		& CSAT-3 Campbell		
(m/s)	2 & 21.5	Scientific	21.65	CSAT-3 Campbell Scientific
Net Radiation		CNR-1, Kipp &		4-component CNR-1, Kipp
(W/m^2)	25.5	Zonen	20	& Zonen
		Mor Electric Heating		
		Association, Inc.		
		ceramic heaters		
Heaters	1.2	FTE-1000	N/A	N/A
		Judd Comm.		
Snow Depth		Ultrasonic Snow		Judd Comm. Ultrasonic
(cm)	1.75-2	Depth Sensors	2.5	Snow Depth Sensors
Soil				
Temperature	0.4.		0.4.	TCAV Campbell Scientific &
(°C)	-0.1 to -0.05	Decagon EC-TM	-0.4 to -0.1	Decagon 51E
Soil Moisture	0.4.		0.4.	CS-615 Campbell Scientific
(% by volume)	-0.1 to -0.05	Decagon EC-TM	-0.4 to -0.1	& Decagon 5TE
		100° Transducer,		
D		Sensotec	2	1E525WS-L, 1exas
Precipitation	Sub-canopy	(@SNOTEL)	Z	Electronics

Table 1: Instrument descriptions and measurement heights on Niwot Ridge, CO and in the Valles Caldera, NM.

It should be noted that precipitation data from the Niwot SNOTEL site was scaled by a factor of 0.9 in 2010 and 0.8 in 2011 to correct for slight precipitation gage over-catch during the snow season. This was based on a ratio of precipitation/SWE calculated from liquid depth of precipitation from the gage compared to SNOTEL snow pillow SWE measurements (under the assumption that the snow pillow has greater accuracy).

Model Scenarios

Control Case

Each model was run with hourly data from the AmeriFlux tower at the subalpine forest site on Niwot Ridge and the Niwot SNOTEL for the LSA as well as using data from UC Merced's in-situ instrumentation described above.

Heated Case

Heated model runs at LSA were produced by increasing the amount of incoming longwave radiation (LW_{in}) by the theoretical amount added to each plot from the heaters (Figure 6). Maximum, still-air heater efficiency to the surface is estimated to be ~50% (Kimball, 2005). For the entire snow season of 2009-2010, heater output was 50% (500 W) or an additional 429 W m⁻² to each heated plot. Taking into account 50% heater efficiency in calm conditions (Kimball, 2005), the additional radiant flux density to each heated plot was 214 W m⁻². In 2010-2011, output varied between 10 %, 20 %, and 40% (Figure 6) producing an additional radiant flux density to each heated plot of 42, 85, and 171 W m⁻², respectively. For example, in the heated plot model run for 2011, the LW_{in} column in the model input file was increased by 42 W m⁻² at each timestep from mid-November 2010 – early March 2011.

Synthetic Warming Case

To address the issue that these IR heaters do not warm the environment by the same mechanisms expected to occur with climate change (i.e. warming of the air, not only the surface) (Amthor et al., 2010), synthetic model runs at LSA were also conducted. Because changes in precipitation are predicted with far less certainty than projected temperature changes (AGCI, 2006), no precipitation changes were accounted for in this study. Since the only difference between heated and control models are increases in LW_{in}, synthetic models were run with

increased air temperature also. Synthetic models were produced by using most of the same CO control plot forcings, yet systematically increasing air temperature and incoming longwave radiation values. Air temperature was increased by taking the average monthly temperature difference (derived from hourly data) between the CO and NM sites, and increasing the CO temperature by this difference in each month (Table 2).

Average air temperature differences,		
NM - CO (°C)	2009-2010	2010-2011
October	3.79	1.95
November	2.26	2.51
December	3.75	2.64
January	1.87	5.08
February	1.74	5.69
March	1.15	6.82
April	3.02	7.98
May	4.14	6.74
June	2.36	6.81
July		4.42

Table 2: Average air temperature differences between CO and NM by month.

 LW_{in} was systematically increased with the following process: Measured values of LW_{in} and air temperature were used to calculate atmospheric emissivity (ϵ_a):

$$\varepsilon_{a} = \frac{M}{\sigma T^{4}} \tag{2}$$

where M is the exitance of the atmosphere (or LW_{in}), σ is the Stefan-Boltzmann constant (5.67 x 10^{-8} W m⁻² K ⁻⁴), and T is air temperature in Kelvin. With this calculated ε_a value, a new LW_{in} was calculated by adding the average difference in temperature between NM and CO to the measured CO air temperature (T_{new}), and using the Stefan-Boltzmann Law:

$$LW_{new} = \varepsilon_a \sigma T_{new}^{4}$$
(3)

Synthetic model runs were completed with these increased temperature and LW_{new} values.

Natural Warming NM Case

The Valles Caldera, NM model was driven with data from the instrument cluster at the flux tower site and at the COSMOS site described above.

SNOWPACK model runs for each of the four cases (control, heated, synthetic, NM) in 2010 and 2011 were produced for the entire snow season, October – summer (June or July) of the following year.

V. RESULTS

Field Results

Meteorology

Mean daily air temperature (above-canopy) at LSA in 2009-2010 dropped consistently below freezing in late November and mostly remained between about -20 and 0 °C through mid-March (Figure 7a).



Figure 7: Daily average air temperature and windspeeds at each site (above canopy measurements).

Air temperature beginning mid-March warmed to $\sim 5^{\circ}$ C or more on average. This site is sheltered with low daily wind speeds below the canopy, but greater windspeeds at the above-canopy instrument height (21.5 m) averaging 4.6 m s⁻¹.

The 2010-2011 mean above-canopy air temperatures at the LSA dropped consistently below freezing in early November and averaged -5 °C through mid-March (Figure 7b). Air temperature began to peak above freezing in mid-March, and increased to \sim 5° C or more on average after late April. Mean daily wind speeds were 5.4 m s⁻¹.

In the Valles Caldera during the 2009-2010 snow season, mean daily air temperature never dropped consistently below freezing for more than two weeks in a row. Daily mean temperature fluctuated between -15° C and $+5^{\circ}$ C until early May after which temperatures remained above freezing at an average of 6° C (Figure 7c). Daily average windspeeds were about 3 m s⁻¹ and always below 6 m s⁻¹.

In the Valles Caldera during the 2010-2011 snow season, mean daily air temperature never dropped consistently below freezing for more than three weeks in a row. Daily mean temperature after snow began to accumulate fluctuated between -25° C and $+5^{\circ}$ C until mid-March after which temperatures still dipped below freezing on occasion yet averaged 7.5° C (Figure 7d). Average daily windspeeds measured 1.8 m s⁻¹ and were never above 6 m s⁻¹.

Snow accumulation and melt

During the 2009-2010 snow season, SWE measurements at the Niwot SNOTEL site show snow accumulation beginning in early November 2009, reaching a seasonal maximum of 310 mm in early April 2010, and melting out completely by May 31 (Figure 8a).



Figure 8: Average observed snow depth and SWE at a) Niwot Ridge, CO, LSA site 2010, b) Niwot Ridge, CO, LSA site 2011, c) VCNP, NM 2010, and d) VCNP, NM 2011.

Seasonal maximum SWE was within 3% of the 25 year mean value of 318 mm. Snow accumulation at the LSA progressed via small, frequent precipitation events with some larger events in late March and late April/early May.

Automated hourly snow depth measurements in heated and control plots began in early March (averages of control and heated depths are plotted). Snow depths in the control plots followed the pattern of the recorded SNOTEL snow depth reaching maximum depths in late March and early April (Figure 8a). Control plots displayed some spatial variability with maximum snow depths ranging from 112 to 145 cm (standard deviation of 15 cm, coefficient of variation (CV) of 0.12). In the heated plots, snow was present only directly after snow events, and melted quickly, never reaching depths greater than 30 cm. Manual observations after late March indicate there was no snow in heated plots (Figure 8a). The ultrasonic snow depth sensors were able to capture this transient snow accumulation and melt in the heated plots.

It should be noted that snow observations were derived from averages of several ultrasonic depth sensors for each control, heated, and NM "observation," and that in the case of the CO snow depth sensors, each sensor is located in the middle of the experimental plot. In addition to small-scale spatial variability in snowmelt caused by varying canopy structure in all plots (Musselman et al., 2012a), the snow does not melt uniformly in the heated plots with much greater melt occurring under the heaters and snow persisting longer in the center (directly under the sensors) (Figure 9a).



Figure 9: Panel a) heated plot during melt on 5/31/11, b) heated plot in mid-winter, 4/15/11.

The 2010-2011 SWE measurements at the Niwot SNOTEL site show snow accumulation beginning in late October, reaching a seasonal maximum of 430 mm in mid-late May, and
melting out completely by June 10 (Figure 8b). Seasonal maximum SWE was 35% greater than the 25 year mean value of 318 mm.

Snow accumulation for 2010-2011 progressed via several smaller precipitation events in the early part of the season, followed by one large storm in early February and subsequent storms in late March, mid and late April, and mid-May. Average peak snow depth was just 6% greater than in 2010. Automated hourly snow depth measurements in heated and control plots began in late January. Snow depths in the control plots followed the pattern of the recorded SNOTEL snow depth reaching maximum depths in late April with another peak following the mid-May event. Control plots showed 10% greater peak accumulation values on average compared to heated plots (Figure 8b). Control plots displayed greater spatial variability than the previous season with maximum snow depths ranging from 100 to 160 cm from manual measurements (standard deviation of 24 cm, CV = 0.20 compared to 2010 control plot standard deviation of 15 cm, CV of 0.12). In the heated plots, snow depth at maximum accumulation ranged from 110 to 140 cm (standard deviation of 13 cm, CV = 0.10). Overall and without taking plot location relative to the overhead canopy cover into consideration, heated plots measured just 4 cm less snow depth than control plots on average. Snowmelt onset and snow disappearance at control and heated plots occurred within 1-2 days, and melt rates were slightly higher in heated plots at $1.8 \text{ cm d}^{-1} \text{ v}$. 1.4 cm d^{-1} in control plots (rates calculated from modeled SWE results).

Again, snow distribution within heated plots was much less uniform than in control plots. The snow levels reached—and in some cases slightly exceeded—the height of the heaters. When snow levels approached or reached or heater height, small melt depressions developed beneath each heater, and melt pockets radiated from the heaters outward toward the center of the plots and downward several centimeters (Figure 9b). Similar to the previous season's melt pattern in heated plots, snow first melted below the heaters with snow persisting in the center of each heated plot several days after areas under the heaters became snow-free. Several layers of meltfreeze crusts and surface hoar were apparent in the snowpack closest to the heaters, while these layers were absent from control plots indicating that the heaters caused snowmelt at the snow surface nearest to them.

Snow accumulation in the Valles Caldera 2009-2010 season began in in late November. Accumulation proceeded with two large precipitation events, one in early December, the next in mid-January. Several smaller precipitation events occurred following these storms up through peak accumulation in mid-March (Figure 8c). Peak accumulation occurred on 15 March, two weeks later than the seven-year average (2005-2011) of 1 March. Peak depth was recorded as 98 cm, about 46 % greater than the seven-year average of 67 cm. Snow disappearance date of 6 May was 2 days later compared to the seven-year average of 4 May.

Snow accumulation in the Valles Caldera 2010-2011 season began in mid-December. Accumulation proceeded with two large precipitation events, one in late December, the next in early January. Several smaller precipitation events occurred following these storms up through peak accumulation on 7 March (Figure 8d), one week later than the seven-year average (2005-2011) of 1 March. Peak depth was recorded as 58 cm; 13% below the seven-year average of 67 cm. Melt-out date by 2 May was very close to the seven-year average of 4 May.

Overall, control plots experienced greater peak snow accumulation compared to heated plots (390% greater in 2010, 10% greater in 2011) and NM snow depth (22% greater in 2010, 220% greater in 2011). NM snowpack was 3 times greater than heated snow in 2010, yet only half the heated plot snow depth in 2011. Snowmelt onset was 9 weeks earlier in NM compared to CO on average, and snow melted 31 days earlier on average in NM compared CO (Figure 10).



Figure 10: Snow accumulation and melt dynamics. Melt onset and snow disappearance was not plotted for LSA heated in 2010 due to the intermittent snowpack.

Comparisons of melt onset and snow disappearance were not possible in 2010 heated plots due to the intermittent nature of the snowpack. Total number of days with snowcover was lowest at CO heated plots in 2010 (90 days) compared to 2010 control and 2011 heated and control (244-251 days). On average, NM experienced 32 % fewer days of snowcover than in CO (except 2010 heated plots) with 158 days in 2010 and 137 days in 2011.

Examining average soil temperature in control plots at LSA in 2009-2010, temperature remained just below freezing for most of the snow season with a mean temperature -0.4 °C. Average soil temperatures began to climb around the time of snowmelt, with the first significant spike in average soil temperature coinciding with the date of snow disappearance on 2 June (Figure 11).



Figure 11: Soil microclimate results. Panel a) shows soil temperature at CO control and heated plots and at NM in 2010, b) soil temperatures in 2011, c) soil moisture as volumetric water content (VWC) in 2010, d) soil moisture in 2011.

After snow disappeared, soil temperatures fluctuated above freezing with a mean temperature of 8 °C. In contrast, soil temperatures in heated plots show a completely different regime that reflects the ephemeral snowpack observed; temperatures fluctuated at or above 0 °C until early March, after which they remained above freezing. Once snow melted in control plots, average soil temperatures in heated plots still remained consistently greater than control soil temperature (average difference = 5.9 °C).

Soil temperature in control plots at LSA in 2010-2011 averaged less than 1° C above freezing for most of the snow season. Temperatures began to climb around the time of snowmelt, the first significant spike in average soil temperatures coinciding closely with the date of snow disappearance on 10 June (Figures 10 and 11). After snow disappeared, soil temperatures remained above freezing, averaging 5.8 °C. Soil temperatures in heated plots closely tracked soil temperatures in control plots, but remained steadily warmer and spiked earlier coinciding with the slightly earlier melt-out date of 7 June. Once snow disappeared, average soil temperatures in heated plots.

Average soil temperature in NM 2010 remained mostly just below freezing before the melt period began with the exception of a small dip to nearly -2° C during February. Soil temperatures rose above 0 °C coincident with snowmelt in early May, and remain above freezing thereafter averaging 9° C (Figure 11).

Average soil temperature in NM 2011 remained around 0 °C before the melt period began. Soil temperatures rose above freezing approximately coincident with snowmelt in midlate March, and remained above 0 °C thereafter averaging 9° C (Figure 11).

Soil Moisture

Average measured soil moisture in 2009-2010 CO control plots show low values (average VWC = \sim 0.12) through the end of February. Beginning in March, soil moisture began to increase steadily reaching an initial peak coincident with the main snowmelt pulse in early June (Figure 11). Average, mid-winter soil moisture in heated plots (VWC = 0.25) was twice as great as control plots reflecting the frequent mid-winter melt events observed well before the main snowmelt pulse in control plots. These soil moisture spikes coincided with snow events that show accumulation immediately followed by rapid and total melt-out of snow (see Figure 8 near days of year 67, 85, 130). However, beginning right before snowmelt (\sim DOY 130), soil moisture in control plots outstripped that in heated plots, and was 14% greater on average through the end of the measurement period (Figure 11).

Average measured soil moisture in CO 2010-2011 control plots showed VWC < 0.2through the beginning of March. Soil moisture then began to increase reaching a peak coincident with the main snowmelt pulse in early June (Figure 11). Average soil moisture in heated plots closely tracked that in control plots, and was on average 5% lower than in control plots. After melt water began infiltrating soil in earnest, control plots averaged 8% greater VWC than heated plots. The spike in soil moisture near DOY 170 was due to a rain event.

Average measured VWC in NM 2010 was ~ 0.2 until mid-March when it began to rise. Peak VWC of 0.34 was reached on 8 May, three days after snow disappeared. Soil moisture receded following the end of snow disappearance (Figure 11).

Average measured VWC in NM 2011 fluctuated between 0.15-0.18 until early March when it began to rise in response to the main snowmelt pulse. Two peaks in VWC of 0.30 and 0.29 were reached on 23 March, about 2 weeks after melt onset and on 19 April, 43 days after

melt onset. Soil moisture receded following the second peak through the end of snow disappearance (Figure 11).

Overall patterns for soil temperature showed heated plots were 2-5 times warmer than control plots in 2010, and 1.5 times warmer in 2011 during snow accumulation and melt. NM soil temperatures varied by year compared to CO soil temperatures, and no overall relationships relative to CO were observed. Soil moisture was generally greater in heated plots before snow melted, but greater in control plots after melt onset. Soil moisture in NM varied, but was 23% greater on average over the entire snow season than CO in 2010, and 21% lower than CO in 2011. These differences in soil moisture values may partially be explained by the differences in soil properties of the two sites; the sandy loam soils of the NM site versus the clays at the CO site. It should also be noted that some soil moisture sensors were not calibrated (varied by site and year), and thus the main utility in examining these soil microclimate data lie in the timing and overall shape of the data, not the specific magnitude of the values.

Modeling Results

Model Validation

Model results show that despite some discrepancies in magnitude of snow accumulation, SNOWPACK model runs estimate the overall patterns of snow accumulation and melt relatively well (Figure 12).



Figure 12: Snow depth is plotted here; observed snow depth and modeled snow depth track each other relatively well except in the heated model for 2011.

Model RMSE ranged from 3 to 12 for all models except the heated case in 2011 (RMSE = 30). The beginning of the increase in model residual starts on 7 March 2011, the same day the heaters were turned up from 10 to 20% output. The RMSE for the 2011 heated case before 7 March was 3, and after was 45. This residual increase can potentially be attributed to the model limitations regarding extra LW_{in} added at the top of the canopy versus the reality of the extra LW_{in} below the canopy (further discussed in Section VI). The increasing model residual in NM 2011 starting

around DOY 40 can be attributed to the different location of the snow depth sensor (at the COSMOS site), and the location of the meteorological and radiation measurements. These locations are 300 meters apart, and the COSMOS site is more open, and north-facing relative to the flux tower. Thus an increase in snow depth in the more open location paired with a north-facing opening resulting in longer snow persistence could explain why the model melts earlier than observations (Musselman et al., 2008).

One canopy parameter was altered from default settings in order to produce a more realistic snowpack. The Beer's Law extinction coefficient default value is set at 0.7 in SNOWPACK, and this value was lowered to 0.4 which "opened up" the canopy since a lowered negative exponent will decrease absorption and increase overall canopy transmissivity:

$$\sigma_{absorb} = 1 - \exp\left(-\alpha LAI'\right) \tag{4}$$

where σ_{absorb} is a dimensionless absorption factor, α is the Beer's Law extinction coefficient which is a function of needle orientation and stand structure, and is typically between 0.3 to 0.8 (length⁻¹), and LAI' (m² m⁻²) is effective LAI (Link and Marks, 1999; Musselman et al., 2012b; Nijssen and Lettenmaier, 1999).

Comparison of bulk snowpack properties such as temperature and density show relatively good agreement during the mid-winter period (mid-January 2010). Modeled temperature falls within ~1 °C of observations, though density is underestimated by 9-30% (Figure 13).



Figure 13: Plots of modeled and observed snowpack temperature and density profiles. Observations are taken from the nearby C-1 snowpits. Plot a) temperature on 1/13/10, b) density on 1/13/10, c) temperature on 5/18/11, d) density on 5/18/11.

During the melt season example, both observed and modeled temperatures are 0 °C in the isothermal snowpack, however the model more frequently underestimates density by 4-25% and by up to 60% in the top 40 cm of the snowpack (mid-May 2011). Overall agreement of modeled and observed snowpack dynamics and physical characteristics allows for a comparison of the surface energy fluxes and mass gained or lost.

Snowpack Accumulation and Melt

The CO control snowpack accumulates 18-77% deeper compared to the other three cases,

and melts the latest (between 16 and 81 days after the warmer cases). The synthetic and NM

snowpacks resemble each other closely in 2010 regarding timing and magnitude of accumulation and melt. In 2011, the synthetic snowpack was similar to CO heated snowpack in terms of accumulation and melt magnitude, while NM experienced reduced snow accumulation and earlier melt. The heated snowpack in 2010 does not resemble any other cases, reflecting the ephemeral snowpack observed that year (Figure 14).



Figure 14: Modeled snow depth for the four modeling cases.

Quantitative comparisons of selected snowpack metrics reveal heated snowpack accumulated 77 % lower snow depth than the control in 2010, and 21% lower and 10 % earlier in 2011 (Figure 15).



Figure 15: Model results comparing timing and magnitude of snow accumulation and melt of all 4 cases in both study years. The * refers to the ephemeral snowpack observed and modeled in heated plots during 2010. There was not one, specific date of melt onset or snow disappearance since melt was occurring continually throughout the entire season.

Synthetic snow depth peaked 15% lower and melted 14% earlier than the control in 2010, and peaked 38% lower and melted 31% earlier than the control in 2011. NM snow depth was 19% lower at peak accumulation and melted 21% earlier than the control in 2010, and was 58% lower at peak accumulation and melted 51% earlier than the control in 2011. Model runs estimate greater differences between 2011 control and heated snowpacks than observations indicate, likely a result of the spatial variability masked when using average overall snow depth data to drive each model. However, similar to observations, snowmelt onset was approximately two months later in control cases compared to the artificially (synthetic) and naturally (NM) warmer snowpacks, and snow disappeared about 1-2 months later in CO compared to synthetic and NM

snowpacks in both 2010 and 2011 (Figures 14 and 15). Based on these metrics, it is evident that the heated case generally does not fall within the bounds of these other two warmer scenarios; timing and magnitude of accumulation and melt either higher or lower than the other two warm cases relative to the control.

The accumulation season was defined to be the time step at which snow > 0 cm until the time step immediately preceding snowmelt onset. Melt was defined as beginning at the first melt onset time step through snow disappearance. The exception to these definitions is the heated case in 2010 due to its ephemeral nature, and new definitions for this case were based on the largest individual accumulation and melt-out event (11 May – 19 May). Snowpack accumulation and melt timing and magnitude as well as average energy and mass fluxes for the "accumulation" period were calculated from 11 May to 14 May 2010 (accumulation event). Melt calculations were for average fluxes and mass losses or gains from 14 May through 19 May 2010.

Mass Fluxes

Modeled, mid-winter sublimation rates varied widely between the warmed model cases compared to the control. In 2010, the heated and NM models showed 85% greater mass losses to sublimation vs. control while synthetic model run had a similarly low rate to the control (Table 3).

Sublimation rates	(mm d-1)
(mid-winter)	
Control 10	-0.08
Heated 10	-0.42
Synthetic 10	-0.05
NM 10	-0.44
Control 11	-0.08
Heated 11	-0.14
Synthetic 11	-0.06
NM 11	-0.34

Table 3: Modeled mid-winter sublimation rates in 2010 ("10") and 2011 ("11").

In 2011, the mid-winter sublimation rate was 76% higher than the control case in NM, the heated model 43% greater than the control, and synthetic case again similar to the control.

Modeled vapor losses from the snowpack to the atmosphere during snowmelt

(sublimation + evaporation while excluding transpiration) also varied among the warm cases

(Table 4).

Table 4: Modeled v	apor loss rates vi	a sublimation + evaporation in	2010 and 2011.
Curle Error motor	(mana d-1)	-	

Sub+Evap rates	$(mm d^{-1})$
(snowmelt)	
Control 10	-0.08
Heated 10	0.87
Synthetic 10	-0.10
NM 10	-1.21
Control 11	-0.15
Heated 11	-0.24
Synthetic 11	-0.21
NM 11	-0.96

The NM vapor loss rate in 2010 was greatest at 1.21 mm d⁻¹, and the control and synthetic modeled vapor losses showed similarly low rates. However, there was net condensation (mass gain) to the snowpack in the 2010 heated model run. In 2011, the NM model produced the greatest vapor loss rate (control 84% less than NM) while heated and synthetic cases had greater vapor loss rates compared to the control by 38% and 29%, respectively.

The percentage of mass lost via melt/runoff during snowmelt is generally between 78-100 % compared to 0-21% for sublimation plus evaporation (Table 5).

	% Sub +	% Melt+Runoff
Model	Evap Loss	Loss
Control 10	0.56	99.44
Heated 10	0.00	100.00
Synthetic 10	1.31	98.69
NM 10	21.41	78.59
Control 11	1.51	98.49
Heated 11	2.46	97.54
Synthetic 11	2.37	97.63
NM 11	14.98	85.02

Table 5: Modeled mass loss proportions (vapor loss to atmosphere v. melt/runoff).

Heated and synthetic cases showed similar proportions of mass loss proportions to control cases in both years. NM models had a much greater proportion of mass loss via sublimation and evaporation compared to all other cases.

Overall, heated and NM cases displayed greater mass loss rates than control cases while synthetic cases had similar mass loss rates to controls.

Energy Fluxes: Accumulation

The partitioning of energy differed between the four modeled cases during the snow accumulation season. Net energy was close to zero during accumulation, though notably positive for the CO heated model in 2011 (Figure 16a).



Figure 16: In a) hourly average net surface fluxes during the accumulation period in both years, and b) individual components of the turbulent and radiative fluxes. Q_A , energy of advection, is not shown since average Q_A was never greater than 0.3 W m⁻².

Net radiative fluxes were negative (a snowpack energy loss) during accumulation for all control and synthetic model cases as well as the 2010 NM case. Net radiation was positive for both heated cases and NM 2011. The magnitude of the net radiative flux was greatest in the 2010 heated case; more than 4.5 times greater in magnitude than the control. Net radiation was of the same magnitude for heated and control cases in 2011, yet opposite in sign (an energy source to the snowpack for the heated case).

The accumulation season partitioning of individual components of the net radiative flux, shortwave radiation (SW) and longwave radiation (LW), differed from the control case to the warmer cases (Figure 16b). SW radiation was always an energy source to the snowpack while LW radiation was always an energy loss from the snowpack. In 2010 net SW in the warmer

cases was 1-38% lower than net SW in the control, and 83-100% greater than the control in 2011. Net LW was of similar magnitude in control v. warmer cases for 2010, except the heated case which was just 14% of the net LW in the control case. Net LW in 2011 was 20% and 27% less than control net LW for the heated and NM cases, respectively. The 2011 synthetic case had 65% greater net LW than the control.

The net turbulent flux was positive (snowpack energy gain) and of similar magnitude for all control and synthetic cases, while it was negative (snowpack energy loss) for all heated and NM cases. The magnitude of the net turbulent flux was greatest in the 2010 heated case; more than eight times greater in magnitude than the control and in 2011 17% greater than the control.

The accumulation season partitioning of individual components of the net turbulent flux, sensible heat (Q_h) and latent heat (Q_e) , differs from the control case to the warmer cases—both in magnitude of the fluxes and the proportion of the fluxes (Figure 16b, Table 6).

Table 6: The latent and sensible heat fluxes for each model as a percentage of total energy exchange during the snow accumulation season.

Model,		
Accumulation	% Qe	$% Q_h$
Control 10	7.5	11.4
Heated 10	36.9	6.0
Synthetic 10	5.4	13.4
NM 10	24.5	23.0
Control 11	11.5	21.1
Heated 11	18.1	8.5
Synthetic 11	6.1	14.7
NM 11	32.2	14.1

The proportion of Q_h in heated cases was 47-62% of Q_h compared to control cases while synthetic and NM cases showed between 18-100% greater proportions of Q_h compared to controls in 2010, and ~30 % less Q_h in 2011. Q_h is an energy source to the snowpack during accumulation in all cases except the 2010 heated (Figure 16b), a possible artifact of the heaters which will be addressed in Section VI.

More of a pattern appeared when examining proportions of Q_e for the different models (Table 6). The synthetic case showed a similar proportion of Q_e partitioning in 2010 compared to the control while Q_e was five times and three times greater than the control for the heated and NM cases, respectively. In 2011, the synthetic case showed a 47% lower Q_e flux compared to the control while heated and NM cases had 57% and 280% greater proportions of Q_e compared to the control case, respectively.

The ground heat flux (Q_g) was an energy source to the snowpack for all models during accumulation (Figure 16a). The Q_g in NM 2010 was the largest in magnitude at more than three times greater than all other models. The Q_g in NM 2011 was not as different in magnitude, just 15% greater than the control Q_g .

The higher proportions of Qe in heated and NM cases are illustrated by examining the Bowen (β) ratio: the proportion of Q_h / Q_e. Calculated β values indicate the greater amount of energy being dissipated as Q_e rather than Q_h in heated and NM cases compared to control and synthetic cases during both 2010 and 2011 (Table 7).

Table 7: Bowen (β) ratio (Q_h/Q_e) and evaporative fraction, EF ($1/(1+\beta)$) for average turbulent fluxes during accumulation.

8		
Model,		
Accumulation	β	EF
Control 10	1.52	0.40
Heated 10	0.16	0.86
Synthetic 10	2.49	0.29
NM 10	0.94	0.51
Control 11	1.83	0.35
Heated 11	0.47	0.68
Synthetic 11	2.40	0.29
NM 11	0.44	0.70

Additionally, similar evaporative fractions (EF, defined as $(1/(1 + \beta))$) between control and synthetic models and between heated and NM models indicate similar partitioning of latent and sensible heat.

By and large, the partitioning of energy during the accumulation season in CO control models appear to match that in synthetic models while the heated and NM models are more similar.

Energy Fluxes: Snowmelt

Energy partitioning during snowmelt differed among the four modeled cases. Net energy was positive providing energy for snowmelt (Figure 17a).



Figure 17: Average surface fluxes, showing the breakdown of net turbulent and radiative flux components.

In contrast to the accumulation season, all models showed positive net radiative fluxes. The warmer cases had smaller magnitudes of net radiative fluxes (20-70% less) compared to the control in 2010. However, the net radiative flux in the heated case was greatest in magnitude in 2011 (26-250% greater than the other three cases).

The partitioning of individual components of the net radiative flux, net shortwave radiation (SW) and net longwave radiation (LW), differed between the control case and the warmer cases (Figure 17b). Net SW radiation was always an energy source to the snowpack while net LW radiation was usually an energy loss from the snowpack or zero (except the heated 2010 case). In 2010 net SW in the warmer cases was 29-65% lower than net SW in the control, and 61-100% greater than the control in 2011. Magnitude of net LW varied greatly in control v. warmer cases in both study years. The 2010 control case showed 20% greater magnitude of net LW than the heated case, yet opposite signs. Control net LW in 2010 was 26% and 37% less than synthetic and NM models, respectively. Net LW in 2011 was 82% and 66% less than control net LW for the heated and NM cases, respectively. The 2011 synthetic case showed net LW six times greater in magnitude than the control.

The net turbulent flux was positive or close to zero for almost all cases during snowmelt, and, as in the accumulation season, greatest for the 2010 heated case. The magnitude of the 2010 heated model was 13 times greater than the 2010 control. The snowmelt season partitioning of individual components of the net turbulent flux differed between the control case and the warmer cases—both in magnitude of the fluxes and the proportion of the fluxes (Figure 17b, Table 8).

Model, Melt	% Qe	$% Q_h$
Control 10	2.6	12.9
Heated 10	17.4	60.4
Synthetic 10	3.9	10.4
NM 10	25.1	31.9
Control 11	9.2	32.6
Heated 11	11.5	8.6
Synthetic 11	7.2	13.7
NM 11	30.9	29.8

Table 8: The latent and sensible heat fluxes for each model run as a percentage of the sum of turbulent and radiative fluxes during the snowmelt season.

 Q_h was greatest in magnitude for the heated 2010 model, and twice as great as the model with the second greatest Q_h flux (NM 2010). The proportion of Q_h in the 2010 heated case was nearly five times greater than Q_h in the control. The synthetic case Q_h was close in magnitude and proportion to the control while the proportion of Q_h in the NM case was 2.5 times greater than the control in 2010. In 2011, the control had the greatest proportion of energy exchange associated with Q_h . The three warmer cases had proportions of Q_h ranging from 9% to 74 % less than the control.

The Q_e flux was negative during snowmelt in all cases except the 2010 heated model. NM models showed the greatest proportion and magnitude of Q_e-- 94% greater than the 2010 control and 86% greater than the 2011 control (Figure 17b). The synthetic case Q_e flux was of similar magnitude to the control in both 2010 and 2011, and showed a similar proportion of Q_e to control cases. The relatively higher proportions of Q_e in NM cases and the 2011 heated case are also illustrated by examining β ratios. Calculated β values support the greater amount of energy exchange associated with Q_e in NM cases and the 2011 heated case compared control and synthetic cases (Table 9).

Model, Melt	β	EF
Control 10	4.95	0.17
Heated 10	3.47	0.22
Synthetic 10	2.65	0.27
NM 10	1.27	0.44
Control 11	3.53	0.22
Heated 11	0.75	0.57
Synthetic 11	1.90	0.34
NM 11	0.96	0.51

Table 9: Bowen ratio (Q_h/Q_e) and evaporative fraction, EF $(1/(1+\beta))$ for average turbulent fluxes during snowmelt.

No discernible EF similarities as found during the accumulation season appear during the melt season for 2010. Yet similar EFs between control and synthetic models and between heated and NM models in 2011 again indicate similar partitioning of latent and sensible heat for these two sets of cases.

The ground heat flux was negative or zero in all cases, with the largest flux in the 2010 heated model (Figure 17a).

Despite greater variability than during the accumulation season, partitioning of energy during the snowmelt season in CO control models appear to match that in synthetic models while the heated and NM models are more similar in terms of magnitude and proportion of individual fluxes.

VI. DISCUSSION

Observations

Snowpack dynamics changed greatly between CO and NM sites, and between heated and control plots. The NM site, though in a similar environment and 3 °C warmer on average than CO, experienced a shorter snow season with 55% lower snow depth, 24% less SWE, 9 weeks earlier snowmelt onset, and 4 weeks earlier snow disappearance (Figures 8, 10). Differences

between heated and control plots varied greatly depending on heater output and different snow conditions over the two seasons examined, and overall heated plots had an estimated 59% less SWE than controls. If we consider the differences in snowpack and soil dynamics of the CO and NM sites from this study to describe typical ranges of subalpine snowpack characteristics across a latitudinal gradient, we can make some inferences about the future of the CO snowpack as well as compare how the heating experiment fits within this range. Additionally, considering the NM snowpack as a rough "reality check" for future changes to CO snowpack, the heaters appear to have been too high at 50% maximum output in 2010 to produce a "realistic" future snowpack, and too low at 10% maximum output in 2011.

One main difference between the two sites was the markedly decreased snowpack in NM compared to CO. Snowpack throughout the western U.S. and in several other mid-latitude mountain ranges has been observed to be declining over the past several decades (Hamlet et al., 2005; Mote et al., 2005). Mote et al., (2005) found SWE decreases of 2-30% throughout the western US from reconstructed SWE records spanning years 1915-2003. This decrease corresponds relatively well with the 24% less SWE in NM compared to CO found in this study, indicating a continuing trajectory of declining SWE if CO snowpack begins to resemble NM snowpack in the future. Several other studies predict between 35-60% SWE reductions in the coming decades (Adam et al., 2009; Beniston et al., 2003; Lapp et al., 2005; Lopez-Moreno et al., 2008), and mirror results from this study showing a 59% SWE decrease in heated plots vs. controls. Thus, a lower snow accumulation as we see in NM compared to CO corresponds to observed, historical snowpack decreases, while decreased SWE in heated vs. control plots corresponds to what may occur in future snowpack.

Additionally, the timing of snowmelt was observed to be much earlier in NM with an overall shorter snow season compared to CO. Earlier snowmelt timing by several weeks has been observed (Clow, 2010; Hamlet et al., 2007) and is estimated to continue toward earlier melt by 1-2 months in the future (Adam et al., 2009; Rauscher et al., 2008; Stewart et al., 2004). The trajectory of earlier snowmelt is comparable to the difference between CO and NM snowmelt timing (9 weeks earlier melt onset, 4 weeks earlier snow disappearance).

Though some of these hydrologic changes in snowmelt-dominated regions can be partially attributed to large, synoptic-scale weather patterns (Beniston, 1997), several other studies have shown that changes in snowpack dynamics are linked to increasing air temperatures (Akyurek et al., 2011; Hamlet et al., 2005; Kapnick and Hall, 2012), distinguishable from natural climate variations such as the Pacific Decadal Oscillation (PDO) (Stewart et al., 2005).

The differences between control, heated, and NM plots regarding snowmelt dynamics highlight these regions' sensitivity to changes in climate, particularly areas with average snow season temperatures close to freezing (Maurer et al., 2007). Bales et al., (2006) highlight areas of the western U.S. particularly prone to transitions of winter precipitation from snow to rain by highlighting the areas where average temperatures range from -3 to 0° C in winter. Indeed, average winter temperatures at both study sites in both study years are greater than -4° C, thus snowmelt dynamics here are likely to exhibit greater sensitivity to changing climate conditions.

Soil microclimate dynamics in snowmelt-dominated ecosystems are largely governed by snow depth, persistence, and snowmelt timing (Filippa et al., 2009; Harte et al., 1995; Walker et al., 1999). Considering the soil temperature differences in the CO control plots compared to the naturally (NM) and experimentally (CO heated) warmer plots, the prediction, "colder soils in a warmer world" by Groffman et al. (2001) is contradicted by our results from the heated plots and

those of Petrzelka (2011). Climate warming is expected to result in shallower snowpack, later snow accumulation, and earlier snow melt (IPCC, 2007; Magnusson et al., 2010). Several studies have found colder soils in areas of declining snowpack due to decreased effects of soil insulation by snow (Groffman et al., 2001; Molotch et al., 2009). However, our heated plots demonstrate much warmer wintertime soil temperatures during the year with transient snowpack, and slightly, but consistently warmer soil temperatures in the year with a consistent snowpack in both heated and control plots. Yet at the same time, soil temperature in NM before snowmelt onset is lower than both CO heated and control soil temperatures. Therefore, we must ask the question: how realistic is this heating experiment with respect to soil dynamics during winter? Though our experimental results contradict Groffman et al.'s "colder soils warmer planet" theory, we find the NM site to be a non-manipulated, warmer snowpack environment which supports more than detracts from this theory during the accumulation season. Further fine-tuning of the heating experiment and further study is necessary to determine what changes to soil temperatures can be expected regarding ecosystem warming.

Peak soil moisture values were greatest in CO control plots, but greater in NM compared to CO heated plots. Timing of peak soil moisture occurred at the same time or earlier in heated compared to control plots, and earlier than all CO plots in NM (coincident with earlier snow disappearance). It has been established that trends in soil moisture, evapotranspiration (ET), and stream runoff in snowmelt-dominated basins of the western US are highly correlated to shifts in snowmelt timing (Hamlet et al., 2007). As previously discussed, snowmelt timing in the heated plots compared to CO control plots occurred 1 day to several weeks earlier, and snow melted in NM plots 1-5 weeks earlier than CO plots. These differences are reflected in timing of peak soil moisture which occurred 2-5 weeks earlier in NM compared to CO plots (Figure 11). Other

studies have predicted (Vicuna et al., 2011) or observed (Molotch et al., 2009) enhanced water losses to ET in snowmelt-dominated mountain basins due to a warmer climate. Barnett et al. (2005) discuss possible ET shifts in snowmelt-dominated catchments highlighting ET changes initiated by earlier snowmelt runoff. Earlier snowmelt increases soil moisture earlier in the year, as seen in the ephemeral snowpack of 2010 CO heated plots and the warmer NM environment. Yet evaporation potential at this time of year is lower (lower net energy relative to late spring and summer). Thus later in the year, there is lower soil moisture due to earlier peaks coupled with a higher potential for evaporation (more available energy) create a stronger vapor gradient and attenuate the effects of what may be a global decrease in ET (Barnett et al., 2005; Ohmura and Wild, 2002). Further exploration and discussion on changing surface turbulent fluxes with changing surface energy balance characteristics can be found in the subsequent discussion of model results.

Certain problems in the experiment design of these IR heated plots with respect to snow studies include the heater output level, and the physical set-up of the plots. Firstly, it is evident from the two study years that heater output levels at 50% of the maximum power produce a highly ephemeral snowpack which, according to the literature, is unlikely to actually occur in the near future (timescale of decades to the year 2100) (Bavay et al., 2009; Lopez-Moreno et al., 2008). Additionally, the heater output level at 10% of maximum power renders a very small change between heated and control snowpacks, which is generally less than predicted changes to snowpack in the coming decades. Thus, a tuning of winter heater output to between 20-40% of maximum power is suggested to produce a snowpack regime likely to occur in the following decades. Secondly, the observations of snow patches persisting in the center of these heated plots (Figure 9a) days to weeks longer than the snow directly underneath the heaters indicates that a

smaller plot diameter may be more suitable (produce more even snowmelt inside each plot) for IR-snow manipulation experiments.

Modeling

Snowpack dynamics varied greatly between the four modeling cases, often reflecting the lower accumulation and earlier snowmelt predicted to occur in mountain snowpack as climate shifts (Adam et al., 2009; IPCC, 2007; Lopez-Moreno et al., 2008). Synthetic and NM cases from this study show that a warmer (artificial or natural) snowpack melt between 23-81 days earlier than the Colorado control snowpack. Consistent with results from this study, others have found that peak SWE may decline by 35% -78% at similar elevations, and that melt may occur one to two months earlier in the season (Bavay et al., 2009; Beniston et al., 2003; Lopez-Moreno et al., 2009; Lopez-Moreno et al., 2008). The other "warmer" case, the heated snowpack, does not match this earlier melt pattern in either 2010 due to the high heater output creating an ephemeral snow regime, or in 2011 due to the low level of heater output (modeled snow melted two weeks earlier on average than the control).

Heater settings could be adjusted in the future to somewhere between the 2010 and 2011 settings during winter to simulate a more realistic future snowpack. Here, I will suggest that the slightly decreased snowpack created by the 2011 heater level output can be likened to the snow conditions of the near future (2020s-2050s) based on a study which found no significant decrease in maximum SWE, yet did find a shift toward earlier snowmelt (MacDonald et al., 2012). The snowpack differences observed between CO control and NM snowpack generally fit within the range predicted for the time period 2071-2100 (Rauscher et al., 2008; Stewart et al., 2004) Additionally, I suggest that since the intermittent snowpack created by the 2010 heater levels is well beyond the range of projected decreases in SWE and earlier snowmelt timing (Bavay et al.,

2009; Beniston et al., 2003; Lopez-Moreno et al., 2008), this situation could characterize snowpack conditions beyond the year 2100.

Though the direct consequences of reduced winter snowpack and earlier melt timing are not fully understood, likely effects include changes to water availability and thus growing season photosynthesis and carbon uptake (Pataki et al., 2000; Sacks et al., 2007), subsequent increases in water stress later in the growing season (Bales et al., 2006; Moyes et al., 2012), and adjustments to water management practices and policy (Viviroli et al., 2011).

Mass losses also differed among the four modeling cases. Though sublimation loss rates from this study were generally lower $(0.05 - 1.21 \text{ mm d}^{-1})$ than those from similar studies using different measurement methods $(0.41 - 3.7 \text{ mm d}^{-1})$ (Molotch et al., 2007; Schmidt et al., 1998) or different model settings (0.64 mm d^{-1}) (Petrzelka, 2011), several revealing patterns emerged. In both study years, heated and NM models show the greatest sublimation rates during midwinter $(0.14-0.44 \text{ mm d}^{-1})$, while control and synthetic models had similarly low sublimation rates $(0.05-0.08 \text{ mm d}^{-1})$. However, mass loss rates via sublimation and evaporation during the melt period look different between heated and NM models with the NM run showing 5-10 times greater loss rates than the other three cases, and net condensation (mass gain) in the heated plot for 2010.

If the heated and NM models are considered more realistic representations of future snowpack conditions than the synthetically warmer models, then greater mass loss to the atmosphere followed by subsequent reductions in groundwater recharge and surface runoff could be expected (Molotch et al., 2009). If the synthetic model run represents the more realistic future snowpack scenario, mass losses via latent heat flux may not change greatly with warmer conditions since this case was more similar to control cases in terms of mass loss rates.

However, in creating the synthetic models, the likely increase in vapor pressure was not accounted for. Future modeling scenarios which take changing atmospheric conditions into account will likely produce more reliable results, and thus output from heated and NM cases versus controls are seen as more realistic than synthetic output for characterizing future CO snowpack.

Main differences in energy partitioning reveal that heated and NM model runs have consistently greater proportions of latent heat flux than the control and synthetic models during accumulation and melt seasons in both study years. This increased latent heat flux in the artificially and naturally warmer cases is consistent with results from Molotch et al. (2009) showing that a warmer year with earlier snowmelt onset increased the latent heat flux by 28%.

Artifacts resulting from IR heaters in the context of a snow manipulation experiment are described in Petrzelka (2011), and several are observed in this study. To examine heater artifacts in the context of this study, consider the following two theoretical scenarios (Figure 18):





Figure 18: Theoretical diagram illustrating the greater loss via sensible heat (increasing the gradient away from the snowpack) due to heaters warming the snowpack, and sensible heat gain during melt (increasing the gradient toward the snowpack during the melt season)—an artifact of the IR heaters. Upward arrows represent energy leaving the snow going into the atmosphere; downward arrows represent energy entering the snowpack.

Scenario 1 in mid-winter shows the heaters raising the temperature of the snow to 0°C while ambient air temperature is unaffected, thus increasing the snow-air temperature gradient relative to the control plot, and subsequently increasing energy loss via sensible heat (Figure 18). This scenario is observed in Figure 16b where the 2010 heated case is the only model with a negative sensible heat flux. Scenario 1 is also evident in the modeling results of Petrzelka (2011). Scenario 2 shows the snowpack in the heated and control plot to both be at the freezing point during the melt period. The sensible heat flux during this time is greater in the heated plot, suggesting that the air above the plot must be warmed to produce a larger temperature gradient to

generate a greater sensible heat flux. Though it contradicts the assumption that air temperatures are unaffected by heaters (Kimball, 2005), Scenario 2 may be an explanation for the greater sensible heat flux in the heated cases compared to the controls.

The increased latent heat flux in the heated plots (Figures 16, 17) as well as the large model error in the heated case for 2011 (Figure 12) might be explained by the way in which SNOWPACK represents atmospheric conditions. Specifically, the model may calculate greater specific humidity values than those that occur in reality for the heated cases, and thus the air with more water vapor will have a higher emissivity and emit more energy to melt more snow. This issue must be investigated in future modeling studies.

Other sources of uncertainty include model limitations. Firstly, SNOWPACK cannot replicate the exact logistics of the infrared heater set-up (adding extra LW_{in} sub-canopy independent of canopy LW enhancement). Secondly, model error may be related to the greater error tendency during snowmelt (Figure 12); several validation studies have shown that SNOWPACK model error increases as a snow cover becomes isothermal at 0°C (Lundy et al., 2001; Rasmus et al., 2004; Yamaguchi et al., 2004). Thirdly, this model is one-dimensional, and real snowpack conditions may not necessarily be adequately represented in 1-D.

Several model sensitivity tests were conducted to see the relative effect of adjusting certain model parameters and inputs such as the percentage of direct canopy throughfall, LAI, and shortwave radiation. The results indicate that all of these variables exert varying influences, but particularly great were differences among % canopy openness, or throughfall (Figure 19a-c).



Figure 19: SNOWPACK model sensitivity testing adjusting input values of a) percentage of canopy openness or direct throughfall percent, b) LAI values, and c) extra shortwave radiation.

VII. CONCLUSIONS

The two main differences in snowpack dynamics observed between CO and NM sites and between heated and control plots—declining winter snowpack (45-81 % less accumulation in NM versus CO) and earlier timing of the main snowmelt pulse (melt onset 2 months earlier in NM versus CO)—may result in decreased water availability during the growing season and thus greater water stress. Soil temperatures were greater in heated plots during snow-free periods, but lower than both heated and control CO plots in NM while snow covered the ground. Soil moisture patterns showed NM peaks earlier than both heated and control plots. High heater settings in 2010 produced an ephemeral snowpack while low heater settings in 2011 did not produce a large difference in terms of timing and magnitude of snow accumulation and melt indicating heater levels should be adjusted to greater than 10 and less than 50% of maximum output to mimic naturally warmer snowpack dynamics such as those in NM.

Reductions in peak SWE and earlier snowmelt are also seen in the three warmer model cases. Both the heated and NM cases indicate changes to energy partitioning will accompany warmer snowpacks, particularly greater latent heat fluxes and mass losses during both accumulation and melt. This scenario would result in less water availability during the growing season for vegetation, reduce groundwater recharge, and require water resource management adjustments. Though energy partitioning in the synthetic cases resembled that of the control cases, the lack of agreement with the other two "warm" models (heated and NM) along with the lack of comprehensive meteorological adjustments (VPG consideration) indicates the synthetic model may not produce a realistic future snowpack while the heated and NM models do represent possible future CO snowpack.

These observed and modeled differences in snowpack, soil dynamics, and energy exchange can have dramatic consequences if the trends we see in this study continue to progress in a similar trajectory. Snowmelt-dominated watersheds in geographically similar but lessdeveloped regions of the world (such as the Himalaya and Andes mountain ranges) without enough man-made water storage infrastructure are currently unable to replace natural water storage in snowpack if observed and estimated changes occur, and are thus particularly vulnerable to dry-season water stress (Barnett et al., 2005; Nogues-Bravo et al., 2007). Further research is necessary to better determine the potential changes to subalpine snowpack and the subsequent ecohydrologic consequences.

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