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Ecohydrological controls on snowmelt partitioning in mixed-conifer sub-alpine forests

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ABSTRACT

We used co-located observations of snow depth, soil temperature, and moisture and energy fluxes to monitor variability in snowmelt infiltration and vegetation water use at mixed-conifer sub-alpine forest sites in the Valles Caldera, New Mexico (3020 m) and on Niwot Ridge, Colorado (3050 m). At both sites, vegetation structure largely controlled the distribution of snow accumulation with 29% greater accumulation in open versus under-canopy locations. Snow ablation rates were diminished by 39% in under-canopy locations, indicating increases in vegetation density act to extend the duration of the snowmelt season. Similarly, differences in climate altered snow-season duration, snowmelt infiltration and evapotranspiration. Commencement of the growing season was coincident with melt-water input to the soil and lagged behind springtime increases in air temperature by 12 days on average, ranging from 2 to 33 days under warmer and colder conditions, respectively. Similarly, the timing of peak soil moisture was highly variable, lagging behind springtime increases in air temperature by 42 and 31 days on average at the Colorado and New Mexico sites, respectively. Latent heat flux and associated evaporative loss to the atmosphere was 28% greater for the year with earlier onset of snowmelt infiltration. Given the large and variable fraction of precipitation that was partitioned into water vapour loss, the combined effects of changes in vegetation structure, climate and associated changes to the timing and magnitude of snowmelt may have large effects on the partitioning of snowmelt into evapotranspiration, surface runoff and ground water recharge.

KEY WORDS snowmelt infiltration; evapotranspiration; snow-vegetation interactions; vegetation change

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INTRODUCTION

In the higher elevations of the western US, seasonal snow accumulation provides the primary source of water input to the terrestrial ecosystem. Understanding the mechanisms that control the accumulation, melt and partitioning of melt water into the various hydrologic pathways has been limited by a lack of integrated measurements of governing fluxes and states. Developing this integrated measurement strategy is particularly important as recent evidence suggests that the mountain snowpack is declining in response to regional increase in spring air temperature (Mote *et al*., 2005). The impact of these changes on sub-alpine forests remains unknown although a reasonable hypothesis is that earlier snowmelt will lead to intensified and prolonged periods of water stress (Bales *et al*., 2006). These effects will likely vary across gradients in elevation, aspect, and physiographic and climatic factors, which control energy exchange between the land surface and the atmosphere during and after the snow cover period.

In areas where winter snowfall dominates over summer rainfall, snowmelt controls the timing and magnitude of both runoff events and soil moisture, which can sustain photosynthesis and carbon uptake late into the summer season (Sacks *et al*., 2007). Complex spatial and temporal heterogeneity in local energy climates, vegetation, topography and associated variability in snow accumulation and melt processes complicate attempts to quantify and model snow distribution and to estimate the timing and magnitude of snowmelt (Molotch and Bales, 2005), the distribution of soil moisture (Zehe and Bloschl, 2004) and rates of evapotranspiration (Wigmosta *et al*., 1994).

Coniferous forests have profound effects on snow accumulation and snowmelt (Faria *et al*., 2000). Large-scale changes in forest distribution associated with beetle infestation, fire, disease and changes in climate have bearing on downstream water quality due to increased erosion and sediment transport and may impact local soil moisture availability and future forest spatial patterns

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and biodiversity (Carignan *et al*., 2000). Additionally, changes in forest ecology from fire suppression have increased forest density (Johnson, 1994), which is known to decrease snow water yield (Golding and Swanson, 1986). We currently lack a mechanistic understanding of the effects of vegetation on snow distribution necessary to address the impacts of these annual, decadal and long-term forest dynamics on water resources.

With regard to these mechanisms, studies at the catchment scale have revealed the general response of hydrological processes to reductions of forest cover. For example, Hibbert (1969) showed that reduction of forest cover decreases water yield, and Kattelmann *et al*. (1983) and Stednick (1996) showed that the timing and duration of snowmelt-induced runoff is highly sensitive to forest cover properties. Controls of forest canopy on snowpack-atmosphere radiative and turbulent transfer have been well documented (Price and Dunne, 1976; Hardy *et al*., 1997; Link and Marks, 1999; Woo and Giesbrecht, 2000; Gelfan *et al*., 2004). Similarly, the effects of vegetation on snow accumulation have been extensively evaluated (Golding and Swanson, 1978, 1986; Davis *et al*., 1997; Faria *et al*., 2000).

To date, the aforementioned works related to snow–vegetation interactions have not been extended to applications related to soil moisture or vegetation response to water availability. Here it is important to note that several works have documented the importance of snowmelt on water availability and therefore photosynthesis and carbon uptake during the growing season (Pataki *et al*., 2000). Direct measurements of these processes are lacking and therefore our understanding of governing dynamics has been limited. Such understanding is critical for predicting vegetation response to shifts in climate and for understanding how vegetation change impacts the basin-scale water balance. In this regard, the strongest signal of changes in climate and vegetation distribution may be observed in snowpack processes; e.g. increased interception and prolonged shading of the snow-surface associated with increased vegetation density (Lopez-Moreno and Latron, 2008); earlier snowmelt associated with increased temperature (Stewart *et al*., 2004) and increased soil freezing due to reduced snow accumulation (Brooks *et al*., 1997; Monson *et al*., 2002). Therefore, ecohydrological responses to shifts in climate and vegetation change may largely depend on snowpack processes and the complex interactions between vegetation distribution, snow redistribution, variability in solar irradiance, snowmelt, soil moisture and soil temperature. These states and fluxes need to be observed directly and continuously, from the onset of snow accumulation to the end of the snowmelt infiltration period—a focus of this article.

Our objective here is to use direct observations to improve understanding of the spatial and temporal relationships between snow accumulation and melt distribution, the distribution of soil moisture and temperature, and vegetation structure. Snow, soil moisture and eddy covariance instrument clusters were used to monitor water fluxes and states within two mixed-conifer subalpine forests at the Valles Caldera, New Mexico (three-year observation period) and Niwot Ridge, Colorado (two-year observation period). Using data from these instrument clusters, we evaluate the following questions:

(i) How does vegetation structure influence the magnitude and timing of snow accumulation and snowmelt?
(ii) How does variability in snow accumulation, snowmelt and snow cover persistence influence the temporal variability in soil temperature and moisture?
(iii) How does the timing and magnitude of snowmelt affect vegetation water use and the partitioning of water into different pathways?

STUDY SITES

Valles Caldera National Preserve, New Mexico

The Valles Caldera Mixed-Conifer instrument cluster (35·888447 N, 106·532114 W) is located in the ~1200 km^2 Jemez River basin in north-central New Mexico at the southern margin of the Rocky Mountain ecoregion (Figure 1a) (Brooks and Vivoni, 2008). The instrument cluster is distributed across the northeast flank of Redondo Peak with snow depth and eddy flux observations at an elevation of 3020 m and soil moisture observations just west of the flux footprint at the Redondito Saddle at 3240 m. Ancillary precipitation and temperature data were collected at the Vacas Locas SNOwpack TELemetry (SNOTEL) site (2844 m), located 28 km to the northwest. Precipitation in the region is bimodal, where ~65% of the annual precipitation falls primarily as snow between October and April and ~35% falls as rain during the monsoon months between July and September. The primary forest type of the study site is a mixed-conifer forest, consisting 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*) along with scattered aspens (*Populus tremuloides*) and very little understory. The mean canopy height around the flux tower (Figure 1c) is 19·6 m and the Leaf Area Index (LAI) during the growing season is 3·43 m^2 m^{-2} (McDowell *et al*., 2008). The region provides a unique setting for study of snow–vegetation interactions as such studies in mid-latitude ecosystems are under-represented within the literature. In this regard, the impact of solar radiation on the snowpack energy balance is greater relative to higher latitudes. As a result, micro-scale gradients in energy fluxes associated with canopy structure may be more significant as compared to higher latitude systems.

Niwot Ridge, Colorado, ameriflux site

The Niwot Ridge, Colorado Ameriflux site (40° 1′ 58″ N; 105° 32′ 47″ W) is located at an elevation of 3050 m approximately 8 km east of the Continental Divide

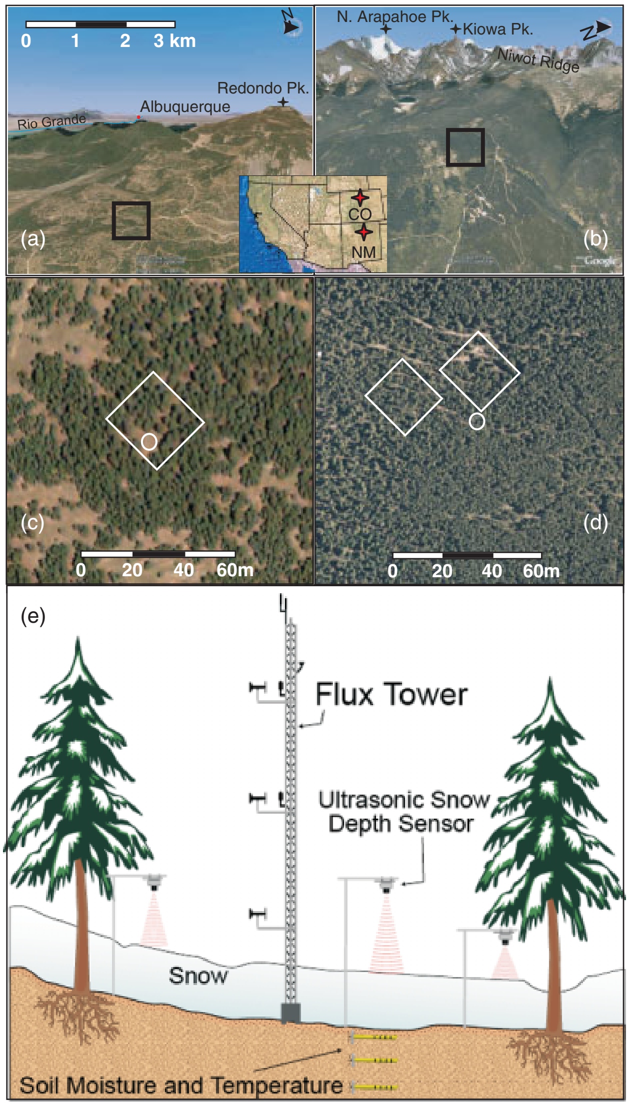

Figure 1. Oblique aerial views of the study sites in the Jemez Mountains of New Mexico (a) and Niwot Ridge Colorado (b). Panels (c) and (d) show vegetation structure surrounding the flux towers (white circles) and the instrument clusters (white diamonds). Panel (e) shows a diagrammatic representation of the instrumentation at each study site.

(Figure 1b). Engelmann spruce (*Picea engelmannii*) and lodgepole pine (*Pinus contorta*) are the dominant species within the area 1 km^2 east of the flux tower. The area 1 km^2 to the west of the tower rises at a slope of about 6–7° and is dominated by sub-alpine fir (*Abies lasiocarpa*), Engelman spruce and lodgepole pine. Maximum LAI during the growing season is approximately 4·2 m^2 m^{-2}. The average gap fraction is 17% and the average canopy height is 11·4 m. The area is in a condition of aggradation, recovering from timber harvesting in the early twentieth century. Annual water input to the area is dominated by moderate snowpacks, which constitute approximately 80% of precipitation (Caine, 1995). Winds predominantly derive from the west, particularly in the winter when periods of high wind velocity and neutral atmospheric stability conditions are common (Turnipseed *et al*., 2002). Site characteristics are described in detail by Turnipseed *et al*. (2002). Relative to the New Mexico site the vegetation structure is more organized with greater vegetation density and smaller gap sizes (Figure 1d).

METHODS

Snow measurements

Nine ultrasonic snow depth sensors (Judd Communications) were installed at each of the two study sites. The nine sensors were positioned in a stratified sampling pattern with respect to proximity to trees, with three sensors in each of three classes: under-canopy, canopy-edge and open areas (Figure 1e). Open and under-canopy conditions corresponded to areas with zero and full canopy coverage, respectively. Canopy edge corresponded to areas with partial canopy coverage and sky view; considered to be an area that receives additional snow input through mechanical removal (unloading) of snow on the canopy (i.e. throughfall). The study duration reported here was dictated by the record length of these distributed snow depth clusters, with a period of record of 2004–2007 at the New Mexico site and 2005–2007 at the Colorado site.

Soil moisture and soil temperature measurements

Observations of soil moisture and soil temperature were used to develop relationships between snowpack accumulation and melt and soil states. In this regard, we investigated relationships between soil temperature and snow depth as snow accumulation insulates the soil from cold winter air temperatures. Similarly, we investigated the timing of snowmelt infiltration onset and the timing of maximum soil moisture with respect to snowpack ablation; emphasis is placed on the timing of soil moisture changes as opposed to magnitude given inherent measurement errors, particularly for frozen soils. Two profiles of water content reflectometers (Campbell Scientific model CS-615) were used to monitor soil moisture conditions surrounding the tower at the Colorado site and a single profile was used at the New Mexico site (Figure 1e; Table I); these differences arise from pre-existing differences in site instrumentation design.

Flux measurements

The eddy covariance technique (Webb *et al.*, 1980; Baldocchi *et al.*, 1988) was used to calculate the turbulent exchange of sensible heat (H), water vapour (expressed as latent heat flux, λE) and carbon dioxide (F_c). In simplified form (i.e. without including details of the so-called Webb–Pearman–Leuning (WPL) 'correction'), these fluxes are calculated as

$$H = (c_{pa}\overline{\rho_a} + c_{pv}\overline{\rho_v})\overline{\omega' T'} \quad (1)$$

$$\lambda E = L_v \overline{\rho'_v \omega'} \quad (2)$$

$$F_c = \overline{\omega' \rho'_c}, \quad (3)$$

where ρ_a is the mean dry air density, w is the vertical velocity, c_{pa} is the dry air specific heat at constant pressure, c_{pv} is the specific heat capacity for water vapour,

Table I. Sensor description and sensor heights for the instrument clusters on Niwot Ridge, Colorado and in the Valles Caldera, New Mexico.

Observation	Niwot Ridge, Colorado		Valles Caldera, New Mexico	
	Measurement height, meters	Instrument	Measurement height, meters	Instrument
Relative humidity (%)	21·5	HMP-35D, Vaisala, Inc.	21·65	HMP-45C, Vaisala, Inc.
Air temperature (°C)	21·5	HMP-35D, Vaisala, Inc./CSAT-3 Campbell Scientific	21·65	CSAT-3, Campbell Scientific
Pressure (kPa)	12	PT101B, Vaisala, Inc.	2	PT101B, Vaisala, Inc.
Net radiation, W m^{-2}	25·5	CNR-1, Kipp & Zonen	20	4-component CNR-1, Kipp & Zonen
H_2O flux (mg m^{-2} s^{-1})	21·5	LI-6262, LI-COR Inc./Krypton Hydgrometer/CSAT-3, Campbell Scientific	21·65	LI-7500, LI-COR Inc./CSAT-3, Campbell Scientific
CO_2 flux (mg m^{-2} s^{-1})	21·5	LI-6262, LI-COR Inc./CSAT-3, Campbell Scientific	21·65	LI-6262, LI-COR Inc./CSAT-3, Campbell Scientific
Wind speed (m s^{-1})	21·5	CSAT-3, Campbell Scientific	21·65	CSAT-3, Campbell Scientific
Wind direction (degrees)	21·5	CSAT-3, Campbell Scientific	21·65	CSAT-3, Campbell Scientific
Precipitation (mm)	10·5	385-L, Met One	2	TE525WS-L, Texas Electronics
Soil heat flux (W m^{-2})	−0·1	HFT-1, REBS	−0·08	HFT-1, REBS
Soil moisture (% by volume)	−0·05, −0·15	CS-615 and CS-616, Campbell Scientific	−0·10, −0·4	CS-615, Campbell Scientific
Soil temperature (°C)	−0·05, −0·15, −0·35	STP-1, REBS	−0·01, −0·10, −0·40	TCAV, Campbell Scientific

T is the temperature, ρ_v is the water vapour density, L_v is the latent heat of vaporization of water (or latent heat of sublimation when no liquid water is present), ρ_c is the partial density of CO_2. An overbar indicates a 30-min mean value and a prime indicates fluctuations around the mean (flux parameters were sampled at a rate of 10 Hz). The flux-measuring instrumentation at the Colorado and New Mexico sites were located at 21·5 and 21·65 m above ground, respectively (Table I). Additional details on the Colorado flux measurements are provided by Monson *et al.* (2002). The covariance of the sonic anemometer temperature fluctuations T' with w' are corrected for the effect of water vapour and velocity fluctuations following Schotanus *et al.* (1983). The primary instrument measuring ρ_v at the Colorado site is the open-path krypton hygrometer, where a correction for both oxygen and temperature-induced density fluctuations (i.e. WPL 'correction') are included in the λE calculation. The closed-path Licor-6262 is a redundant measurement of ρ_v at the Colorado site. A short section of copper tubing removes temperature fluctuations from the air sample before it enters the LI-6262 sample cell so that λE does not require the WPL correction; λE measured with the Licor-6262 has been shown to be 3–7% smaller than λE measured with the krypton hygrometer though larger differences can occur during the few days immediately following snowfall (Turnipseed *et al.*, 2002). From April 2006 to February 2007 the krypton hygrometer was unavailable at the Colorado site (due to a shortage of krypton source tubes) so λE during this period is calculated exclusively with the Licor-6262. The primary instrument at the New Mexico site providing ρ_v observations is an open-path Licor-7500 (Table I).

Measurements of above-canopy F_c were used to determine when latent heat fluxes were associated with ecosystem-scale photosynthesis by the forest canopy (i.e. negative F_c values indicate stomatal uptake of carbon and release of water vapour). Prior to this period λE can be assumed to be associated with snow sublimation (Molotch *et al.*, 2007). Turbulent flux measurements at the Colorado site have been evaluated by comparing the available energy (i.e. net radiation minus soil heat flux) to the sum of H and λE. During the daytime the sum of the turbulent fluxes account for 80–90% of the radiative energy input into the system (Turnipseed *et al.*, 2002). At night, under moderate turbulent conditions, the energy balance is comparable to the daytime; however, when the night-time conditions are either calm or extremely turbulent, H and λE only account for 20–60% of the net longwave radiative flux; Turnipseed *et al.* (2002) explored several possible reasons for this night-time discrepancy (e.g. instrumental error, footprint mis-match, horizontal advection), but could not explain the reason for the night-time imbalance.

Ancillary meteorological measurements

Annual differences in climatology were recorded at the SNOTEL sites located adjacent to the study areas (i.e. the University Camp, Colorado and Vacas Locas, New Mexico SNOTEL sites). SNOTEL data were used for this comparison as the instrumentation is consistent at the two sites, facilitating direct comparisons of basic climate data. In this regard, the temperature and precipitation comparisons were performed solely to obtain a general understanding of how different the climatological conditions were at these two sites during the study period. These comparisons are not the focus of the analyses but provide useful background information.

RESULTS

Three seasons of data from New Mexico and two seasons from Colorado represent a wide range of variability in both the amount and timing of snow accumulation, winter temperature, soil moisture and onset of the growing season. Each of these components of the terrestrial water balance is described below. Sections on Temperature and Precipitation describe the general differences between the climatic conditions at these two sites. Variability in snow–vegetation interactions (see Section on Snow–vegetation interactions) and associated observations of soil temperature (see Section on Soil temperature) and soil moisture (see Section on Soil moisture) are described as are observed variability in water vapour fluxes to the atmosphere and energy fluxes (see Section on Sublimation and Evapotranspiration).

Temperature

A direct comparison of temperature measurements from the two local SNOTEL sites indicates a strong linear correlation ($R^2 = 0{\cdot}89$) (Figure 2). The average air temperature recorded at the Vacas Locas, New Mexico SNOTEL site was 0·6, 3·2 and 2·1 °C during the 2004–2005, 2005–2006 and 2006–2007 snow seasons, respectively. In 2004–2005, the 10-day average air temperature dropped below zero in November and remained below zero until mid-April (Figure 3). Conversely, in 2005–2006, the 10-day average air temperature oscillated around the 0° mark, indicating mid-winter periods with considerable warming and potential surface snowmelt. In 2006–2007, the 10-day mean fell below 0 °C in early December and remained below zero until mid-March. Spring onset, defined here as the date when the 10-day running mean temperature reached a threshold of 0 °C, began on 12 April 2005, 31 March 2006 and 11 March 2007 (Figure 3). Relative to 2004–2005, spring onset occurred 13 and 29 days earlier in the 2005–2006 and 2006–2007 snow seasons, respectively.

The mean snow-season air temperature at the Colorado site was 3 °C colder than the New Mexico site during our study period and was much less variable, averaging −1·4 in 2005–2006 and −1·07 in 2006–2007. Unlike the New Mexico site, the 10-day running average air temperature remained below 0 °C throughout the winter period (Figure 3). The 10-day average temperature dropped below zero in late November for both years and rose above zero on 6 April 2006 and 12 March 2007 for the two years, respectively. Hence, spring onset commenced 25 days earlier in 2006–2007.

Precipitation

Measured precipitation at the Vacas Locas SNOTEL site was equal to 47, 14 and 37 cm in the 2004–2005, 2005–2006 and 2006–2007 water years, respectively (Figure 3); equivalent to 134%, 40% and 106% of the six-year average. During the 2004–2005 water year, precipitation was spread throughout the winter. Conversely, only one notable precipitation event occurred during the winter of the 2005–2006 water year (i.e. on 10 March). The 2006–2007 water year was marked by a large early season snow storm around 20 December, a notable dry

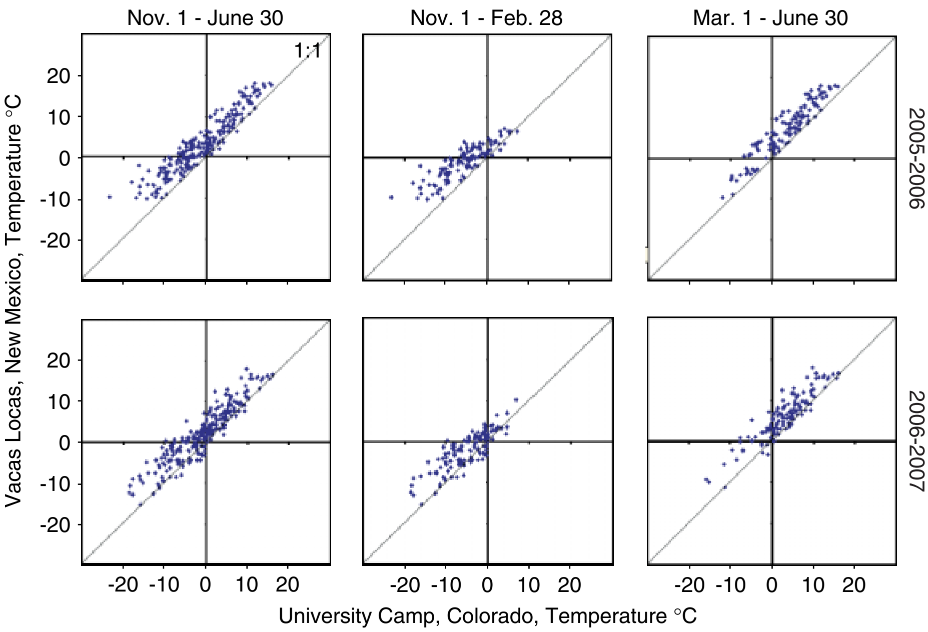

Figure 2. Relationship between snow-season air temperatures at the Vaca Locas, New Mexico (y-axis) and University Camp, Colorado (x-axis) SNOTEL sites; a similar seasonal distribution of temperatures, but approximately 3 °C warmer temperatures in New Mexico suggest that these sites are good comparisons for evaluating the effects of a warming climate on snow–vegetation interactions.

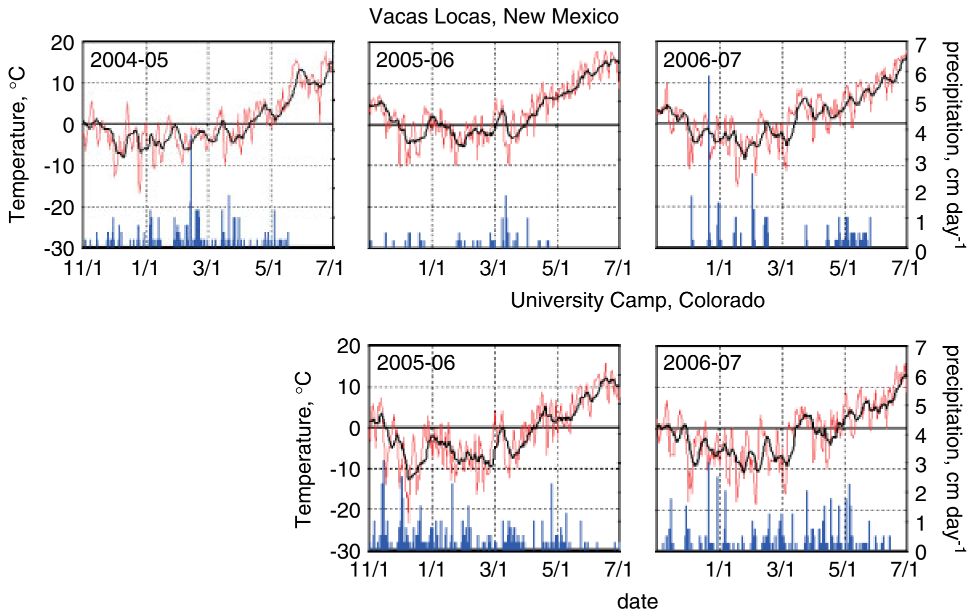

Figure 3. Time series of temperature (red) and precipitation (blue) during the three study seasons at Vacas Locas, New Mexico (top row) and two seasons at University Camp, Colorado (bottom row).

period from mid-February to mid-April, and a series of late season precipitation events from mid-April to mid-May.

Average precipitation during the two-year study period for the Colorado site was 66 and 62 cm for 2005–2006 and 2006–2007 water years, respectively. These totals represent 88% and 82% of the 29-year average, respectively, and represent far less inter-annual variability than observed at the New Mexico site over the same time period (Figure 3). In both years, precipitation was evenly distributed throughout the season relative to the high temporal variability observed at the New Mexico site.

Snow–vegetation interactions

Using the ultrasonic snow depth sensors, we observed four primary interactions between vegetation and snow, including snowfall recorded during individual snowfall events; total winter snow accumulation; decreases in depth before spring melt due to either settling or sublimation and decreases in snow depth during the melt season associated with snow settling and melt. Here it should be noted that with the ultrasonic snow depth sensors, we cannot attribute decreases in snow depth to snow settling and melt independently and therefore in the section on snow settling and ablation we describe the combined processes of snow settling and melt using the term *snow ablation*.

Snow depth. Maximum snow depth at the New Mexico site was six times greater in the wet winter of 2004–2005 than the dry 2005–2006 winter and 54% greater than in 2006–2007 (Figure 4a–c; Table II). The influence of vegetation on snow depth variability was much greater in 2004–2005 and 2006–2007 (Coefficient of Variation (CV) = 0·45 and 0·5, respectively) relative to the dry year of 2005–2006 (CV = 0·18) (Table II). In 2004–2005 maximum snow accumulation was 49% greater in open versus under-canopy areas (Figure 4d). Conversely, during the low snowfall year of 2005–2006, maximum snow accumulation was actually 8% greater in under-canopy locations (Figure 4e); likely a result of greater mid-winter ablation in open relative to under-canopy areas. In 2006–2007 maximum snow accumulation patterns were more similar to 2004–2005 with open areas 15% greater than under-canopy locations (Figure 4f; Table II). The date of peak snow depth was not related to the amount of snow fall, occurring on 26 March 2005, 23 March 2006 and 2 February 2007. Note that the peak accumulation in 2006–2007 was followed by additional accumulation and significant ablation did not begin until 3rd March which is 23 days earlier than the 2004–2005 date of peak accumulation.

In contrast to the high inter-annual snow depth variability at the New Mexico site, maximum snow depth at the Colorado site differed by less than 3% for the two study years (Figure 5a–d). Spatial variability in maximum snow accumulation was significantly greater in 2006–2007 relative to 2005–2006 (Figure 5a and b); the CV was 0·14 and 0·21, respectively (Table II). Vegetation played a strong role in controlling this variability as maximum snow depth was 27% and 63% greater in open versus under-canopy locations in 2005–2006 and 2006–2007, respectively (Figure 5c and d; Table II). Additionally, there was a 36-day difference in the timing of maximum snow depth, occurring on 20 March 2006 and 25 April 2007.

Snowfall. During the 2004–2005, 2005–2006 and 2006–2007 snow seasons, the number of notable snowfall events (i.e. snowfall greater than 10 cm) measured at the New Mexico site was 4, 3 and 8, respectively (Figure 4a–f); note that in 2004–2005 the observation period began in mid-February and thus all snowfall events

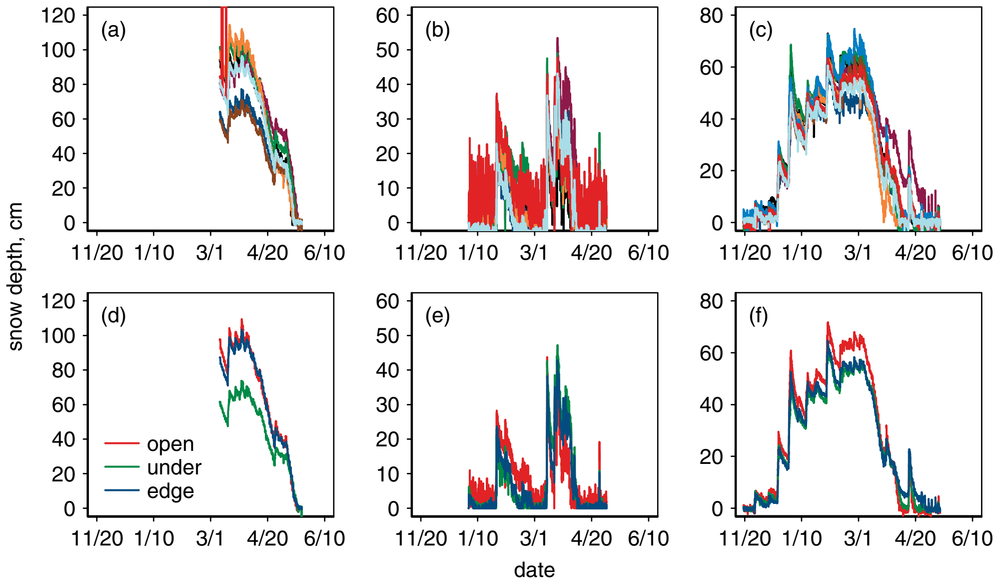

Figure 4. Time series of snow depth in 2004–2005 (a), 2005–2006 (b) and 2006–2007 (c) observed using nine ultrasonic snow depth sensors at the New Mexico site. Average snow depth observed at under-canopy (green), canopy-edge (blue) and open (red) areas are shown for the three years in panels (d)–(f).

Table II. Depth and timing of snowpack accumulation and ablation during the course of the study; under, open and edge refer to sensor position with respect to tree canopy.

	Valles Caldera, NM			CU-Ameriflux, CO	
	2004	2005	2006	2005	2006
Snow onset	11/22[a]	1/25	11/28	11/12	10/18
Date max. snow	3/26	3/23	2/02	3/20	4/25
Snow dissap.					
Under	5/14	4/10	4/19	5/16	5/27
Open	5/14	4/08	4/17	5/25	6/11
Edge	5/16	4/10	4/27	5/25	6/11
Max. snow (cm)					
All	124·4	17·6	80·8	115·8	112·8
Under	73·8	18·6	61·1	98·8	89·6
Open	109·5	17·2	70·6	125·7	145·7
Edge	103·4	15·6	64·4	118·2	117·9
Max. snow, CV	0·45	0·18	0·5	0·14	0·21

[a] Date inferred from observation at Vacas Locas SNOTEL site.

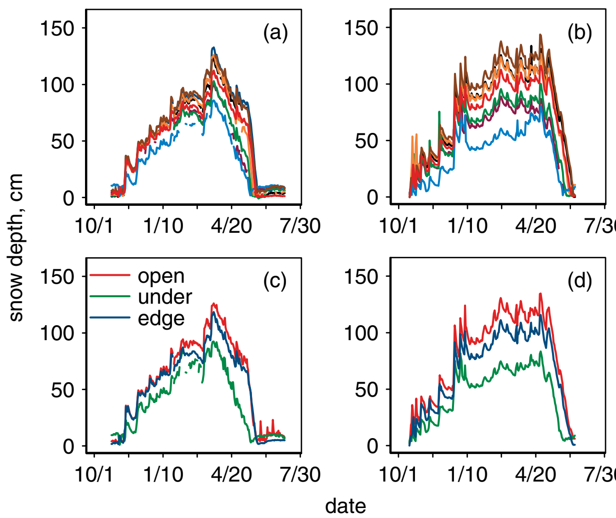

Figure 5. Time series of snow depth in 2005–2006 (a) and 2006–2007 (b) at the Colorado site. Snow depth averages for under-canopy (under), canopy-edge (edge) and open (open) areas are also shown (c) and (d).

were not recorded. Total snowfall during these events was 20% and 5% greater in open and canopy-edge locations relative to under-canopy locations, respectively. Total snow accumulation during 2004–2005, 2005–2006 and 2006–2007 was 38%, 5% and 19% greater in open versus under-canopy locations, respectively; note that differences were lowest during the low snow year. Snowfall recorded at canopy-edge locations was 25% and 4% greater than under-canopy locations in 2004–2005 and 2006–2007, respectively. Conversely, in the low snow year of 2005–2006, measured snowfall at canopy-edge locations was actually 10% lower than under-canopy locations for the three notable events. These differences indicate significant spatial variability in snow accumulation patterns associated with vegetation structure and that these patterns exhibit considerable inter-annual variability. Furthermore, these observations suggest that snow accumulation (and water availability) under the canopy may be less sensitive to inter-annual variability and mid-winter melt relative to open areas.

At the Colorado site, 7 notable snowfall events were observed in 2005–2006 and 11 notable events were observed in 2006–2007 (Figure 5a–d). It is important to note that SNOTEL observations were used to identify snowfall events during a 75-day data gap from 5 January to 22 March 2007; this gap has been filled using a linear regression between SNOTEL snow depth measurements and observed snow depth at each ultrasonic snow depth sensor (R^2 values were 0·95 for open and canopy-edge locations and 0·88 for under-canopy locations on average; $p < 0{\cdot}005$). The timing and magnitude of snowfall was quite variable for the two years, with the 2005–2006 season largely composed of a series of smaller snowfall events and the 2006–2007 season consisting of relatively large magnitude events early in the winter (Figure 5a and b); on average notable snowfall events were 17% greater in 2006–2007 relative to 2005–2006. Averaged for all snowfall events in both years, total accumulation was, respectively, 47% and 31% greater in open and canopy-edge locations relative to under-canopy locations (Figure 5c and d). Interestingly, snowfall in open areas was only 14% greater than under-canopy areas in 2005–2006 but was 74% greater in 2006–2007. As with the New Mexico site, these observations indicate that considerable inter-annual variability exists in the relationships between vegetation structure and snow distribution. In general, greater accumulation was observed in open areas, particularly for years with greater total snowfall.

Snow settling and ablation. At the New Mexico site, onset of the snow ablation season (as defined by the date of local maximum snow accumulation) preceded the spring onset by 14, 6 and 8 days in 2004–2005, 2005–2006 and 2006–2007, respectively (Table II). While snow ablation associated with snowmelt is unlikely if air temperatures are significantly below 0 °C, our sensors also measure snow settling that begins immediately after snowfall. After local maxima in snow depth, snow ablation rates at the New Mexico site were twice as rapid in open and canopy-edge locations versus under-canopy areas (Figure 4d–f). In 2004–2005 and 2006–2007, snow cover duration in under-canopy locations was equivalent to open areas despite the lower amount of maximum snow accumulation (Figure 4d–f).

The onset of the snow ablation season in Colorado occurred 36 days later in 2006–2007 relative to 2005–2006 (Table II). Ablation began 17 and 5 days prior to the onset of spring in 2005–2006 and 2006–2007, respectively. During the 2005–2006 ablation period, snow settling and ablation rates were 32% and 19% greater in open and canopy-edge locations relative to under-canopy areas, respectively. Relationships were consistent in 2006–2007 with 28% and 16% greater ablation rates in open and canopy-edge locations relative to under-canopy. Despite the more rapid snow ablation in open areas, snow cover duration was 9 and 15 days longer in open areas relative to under-canopy areas in 2005–2006 and 2006–2007, respectively (Figure 5c and d); a result of the significantly greater maximum accumulation in open areas. These snow disappearance dates lagged behind the onset of spring by 46 days in 2005–2006 and by 26 days in 2006–2007. While these results indicate substantial variability from year to year and from site to site, the overall signal is clear in that snow settling and ablation rates are greater in open areas relative to under-canopy areas.

Soil temperature

Timing of soil insulation. Soil temperatures were highly sensitive to snow accumulation as indicated by the significant differences in winter soil temperatures during the three years of observations at the New Mexico site (Figure 6a–c). The well developed snowpack at the New Mexico site during the winters of 2004–2005 and 2006–2007 insulated the ground from cold winter air temperatures whereas in the shallow snow year of 2005–2006 soil temperatures remained sensitive to diurnal variability in air temperature throughout much of the winter season (Figure 6a–c). Average subnivean soil temperatures were colder beneath shallow snow (i.e. 2005–2006) and warmer for deeper snow (i.e. 2004–2005) (Table III).

Similar relationships between snow accumulation and soil temperature were observed at the Colorado site. Despite the colder air temperatures, the deeper snowpack at the Colorado site kept subnivean soil temperatures well above those observed at the New Mexico site (Table III). Average soil temperatures during the snow

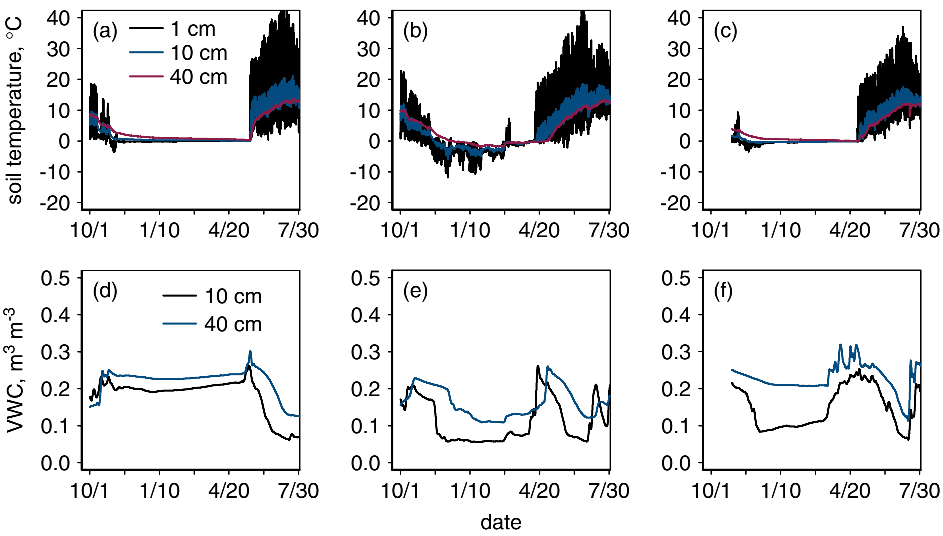

Figure 6. Time series of soil temperatures (a)–(c) and volumetric soil water content (d)–(f) at the New Mexico site for the 2004–2005, 2005–2006 and 2006–2007 snow seasons (left to right).

Table III. Date of soil insulation, soil thaw, infiltration onset, and maximum soil moisture, soil temperature and soil moisture values.

	Valles Caldera, NM			CU-Ameriflux, CO	
	2004	2005	2006	2005	2006
Soil insulation	11/08	3/7	12/19	11/13	10/16
Soil thaw	5/18	4/11	4/20	5/18	6/14
Avg. soil temp.					
1 cm	−0·1	−1·8	−0·4	0·12	0·45
10 cm	0·2	−1·5	−0·3	0·18	0·53
40 cm	0·9	−1·0	0·45	0·56	0·91
Infiltration onset	5/12	4/2	3/13	4/24	3/18
Date of max. moisture	5/18	4/15	4/17	5/21	5/28
Max. VWC					
Surface	0·291	0·27	0·27	0·2	0·43
Depth	0·385	0·25	0·41	0·45	0·51

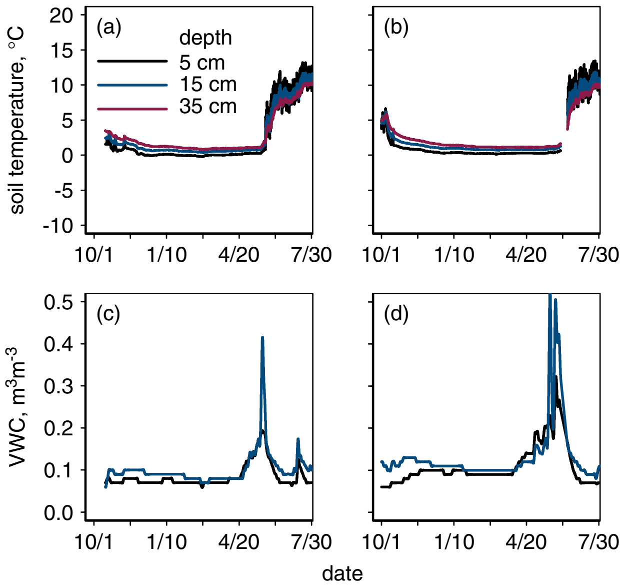

Figure 7. Time series of soil temperatures (a) and (b) and volumetric soil water content (c) and (d) at the Colorado site during the two years of the study.

cover period were 35% warmer in 2006–2007 relative to 2005–2006; air temperatures were warmer in 2006–2007 and snowpack development began earlier. Relative to the New Mexico site, early-season snow accumulation was greater and therefore soil temperatures were warmer throughout the snow season despite considerably colder air temperatures.

Timing of spring soil warming. At the New Mexico site, increase in soil temperature associated with snow disappearance occurred 37 and 28 days later in 2004–2005 relative to 2005–2006 and 2006–2007, respectively (Figure 6a–c). The timing of this spring soil warming was strongly dependent on the presence of snow cover; soil warming lagged snow disappearance by 3, 2 and 0 days during the three years, respectively. Spring soil warming was delayed by 36, 12 and 41 days, respectively, relative to the increase in spring air temperatures, indicating that temporal dynamics in snow cover plays a larger role than air temperature in controlling spring soil temperatures. The warming of soils at the Colorado site commenced 27 days earlier in 2005–2006 relative to 2006–2007 (Figure 7a and b) and also was dependent on the disappearance of seasonal snow. Soil warming lagged behind the onset of spring by 42 and 55 days for the two years, further indicating the weak dependence of soil temperatures on air temperature and strong dependence on the magnitude of snow accumulation.

Soil moisture

Onset of snowmelt infiltration. At the New Mexico site, the timing of initial snowmelt infiltration was 40 and 60 days later in 2004–2005 relative to 2005–2006 and 2006–2007, respectively (Table III). This follows the trends observed with soil temperature, whereby snow cover depletion and increases in soil temperature occurred much later in 2004–2005 relative to 2006–2007 despite small differences in total accumulation. With regard to the timing of spring onset (i.e. the date when the 10-day mean air temperature exceeded 0 °C), initiation of snowmelt infiltration was delayed by 33, 4 and 2 days for the three years, respectively. The differences between 2004–2005 and 2006–2007 are intuitive as the deeper snowpack and colder winter air temperatures resulted in significantly greater cold content in the 2004–2005 snowpack and therefore more energy was needed to ripen the snowpack and release melt water.

At the Colorado site snowmelt infiltration occurred 24 days earlier in the warmer year of 2006–2007. The temporal lag between infiltration onset and spring onset was 18 days in 2005–2006 and 6 days in 2006–2007. Thus, temperature and snow accumulation largely controlled the timing of infiltration onset with increase (decrease) in snow accumulation and decrease (increase) in winter air temperature resulting in longer (shorter) delays between spring onset and snowmelt infiltration.

Peak soil moisture and water availability. At the New Mexico site, peak soil moisture at 40 cm depth was 35% and 38% lower in 2005–2006 relative to the 2004–2005 and 2006–2007 seasons, respectively (Table III). The timing of peak soil moisture was 33 and 17 days earlier in the low snow year of 2005–2006 relative to 2004–2005 and 2006–2007, respectively. The timing of peak soil moisture at the surface lagged spring onset by 39, 17 and 37 days for the three years, respectively. Hence, a trend is apparent whereby peak soil moisture and water availability is delayed for deeper snowpacks; cold content is greater for deeper snow, requiring more energy for significant melt-water production as previously noted. Interestingly, maximum soil moisture and water availability occurred nearly coincident with the date of snow disappearance; peak soil moisture occurred within four days of the snow disappearance date on average (Tables II and III).

At the Colorado site, peak soil moisture near the surface was 115% greater in 2006–2007 versus 2005–2006 (Table III); at a depth of 15 cm, peak soil moisture was 13% greater in 2006–2007. The overall water availability was greater in 2006–2007 as indicated by the greater area beneath the snowmelt pulse curve shown in Figure 7c and d. Relative to the spring onset, peak soil moisture was delayed by 45 and 38 days in the two respective years. Given that total snowfall was consistent for these two years, the differences in the timing of peak soil moisture largely result from differences in winter and spring air temperature and associated impacts to snowpack cold content and surface energy balance. As with the New Mexico site, peak soil moisture occurred nearly coincident with the date of snow disappearance (Table III). Although both sites exhibited considerable variability in both the timing and magnitude of peak soil moisture, the general patterns show that increase (decrease) in snow accumulation and decrease (increase) in air temperature result in longer (shorter) lags between spring onset and peak soil moisture.

Sublimation and Evapotranspiration

Variability in net radiation and sensible and latent heat fluxes at the New Mexico site are shown in Figure 8a–c. Net radiation averaged 58 W m^{-2} during the 2006–2007 snow cover period. During the period prior to the onset of snowmelt (i.e. 28 November to 12 March) net radiation averaged 18·2 W m^{-2} and after the initiation of snowmelt (i.e. 13 March) net radiation averaged 99·8 W m^{-2}. Available energy was largely partitioned into sensible heat fluxes with a median ratio of sensible heat flux to net radiation of 0·67 relative to a latent heat flux to net radiation ratio of 0·11. Mid-winter latent heat fluxes were sensitive to the timing and magnitude of snowfall and subsequent sublimation (Figure 8c). Based on these observations, total snow sublimation averaged 0·66 mm d^{-1}. Latent heat fluxes responded rapidly to soil water availability during the snowmelt period as illustrated by the average evapotranspiration during this period of 2·1 mm d^{-1}; note that this represents the combined vapour flux associated with transpiration, evaporation of liquid water at the snow–atmosphere interface and snow sublimation. Integrated for the entire 2006-07 season, the total water vapour flux to the atmosphere was 386 mm. As flux observations were only available for the single year at the New Mexico site, inter-annual comparisons are focused on the Colorado site in the text below.

Average net radiation at the Colorado site during the 2005–2006 and 2006–2007 snow seasons was 91·9 and 86 W m^{-2}, respectively (Figure 9a and d). Prior to the initiation of snowmelt infiltration (i.e. the pre-snowmelt period) average net radiation was 9% greater in 2006–2007. Conversely, after the initiation of snowmelt infiltration average net radiation was 19% lower in 2006–2007. This difference largely results from the persistence of storms and cloud cover in the spring of 2006–2007 (Figure 3d and e) and the earlier onset of

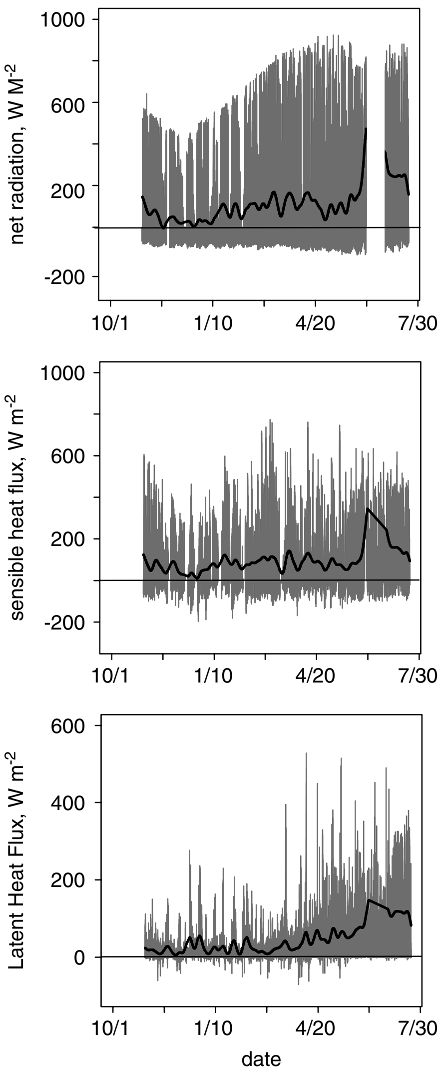

Figure 8. Net radiation (a), sensible heat flux (b) and latent heat flux (c) at the New Mexico sites during the winter of 2006–2007. The black line indicates the 10-day running mean.

snowmelt when solar elevation is relatively low. Despite the relatively low available energy during the 2006–2007 snowmelt period, latent heat fluxes were 18% greater than 2005–2006 (Figure 9c and f) due to the earlier onset of snowmelt and greater overall water availability associated with spring snowfall events (Figure 5a–d). As a result, the proportion of available energy partitioned into sensible heat flux was lower in 2006–2007 (Figure 9b and e); the median sensible heat flux to net radiation ratio was 0·66 and 0·61 for 2005–2006 and 2006–2007, respectively.

Given the greater latent heat flux in 2006–2007, snow sublimation during the pre-snowmelt period was higher in 2006–2007 at 0·9 versus 0·78 mm d^{-1} for 2005–2006. After the onset of snowmelt infiltration water vapour flux averaged 1·84 mm d^{-1} in 2005–2006 versus 1·92 mm d^{-1} in 2006–2007. Integrated over the entire snowmelt infiltration period, the total water loss to the atmosphere was 114·3 and 163 mm, respectively, for the two years. Integrated for the entire 2005–2006 and 2006–2007 seasons, the total water vapour flux to the atmosphere

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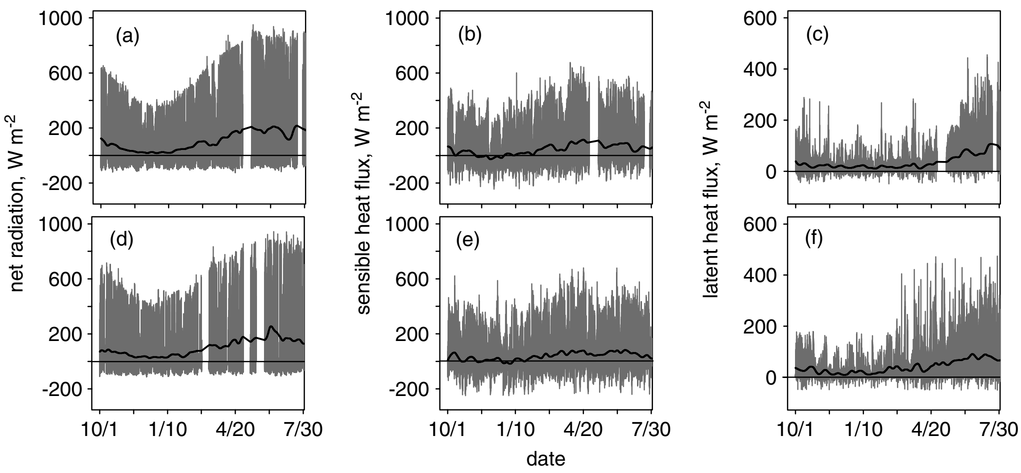

Figure 9. Net radiation (a), sensible heat flux (b) and latent energy flux (c) at the Colorado sites during the winter of 2005–2006 (top row) and 2006–2007 (bottom row). The black line indicates the 10-day running mean.

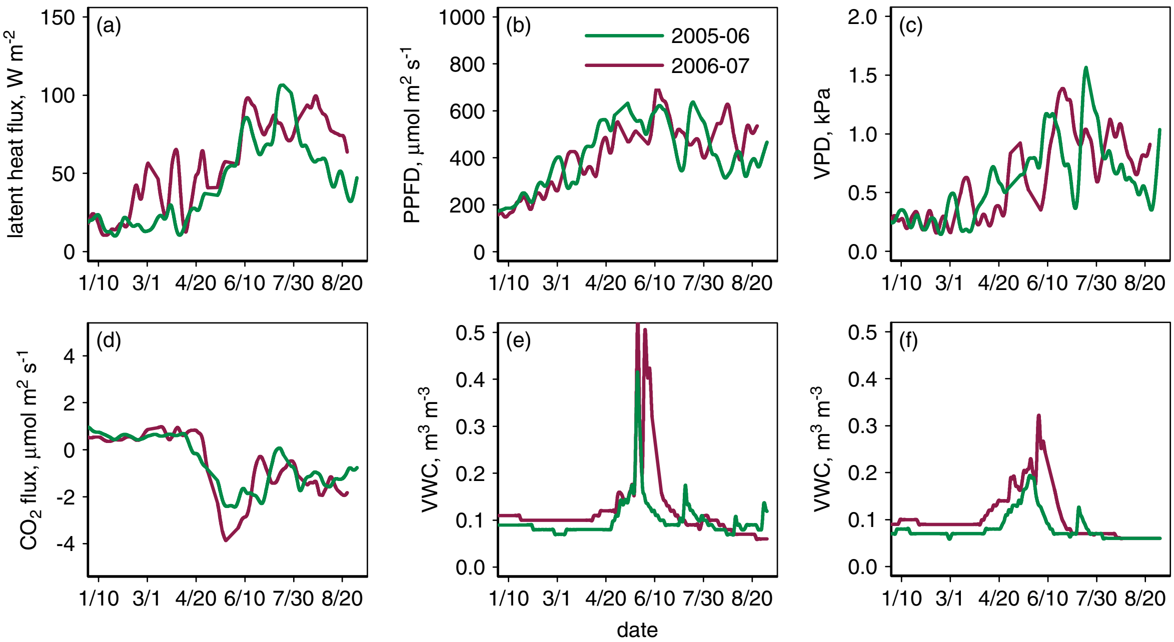

Figure 10. Latent heat flux (a), photon flux density (b), vapour deficit (c), net ecosystem exchange (d), and volumetric water content for the surface (e) and at depth (f) for the 2005–2006 (green) and 2006–2007 (red) snowmelt infiltration period.

was 274 and 351 mm, respectively. This vapour flux represents 41% and 56% of total snow-season precipitation for the two respective years with 2006–2007 having a 28% greater water loss to the atmosphere relative to 2005–2006. As precipitation amounts were nearly identical in these two years, these differences in water vapour fluxes result largely from the warmer temperatures in 2006–2007, the earlier onset of snowmelt, and associated longer snowmelt infiltration period.

The observed differences in water partitioning, with greater water availability and longer duration snowmelt season in 2006–2007, significantly affected net ecosystem carbon exchange (Figure 10a–f). A series of early melt season pulses in latent heat flux were evident in 2006–2007 (Figure 10a). Variables controlling vegetation water use such as photon flux density (Figure 10b) and vapour pressure deficit (Figure 10c) do not show distinct differences in 2005–2006 and 2006–2007 but CO_2 flux (Figure 10d) appeared to respond strongly to greater water availability at 15 cm depth (Figure 10e) and 5 cm depth (Figure 10f) in 2006–2007. Here it is evident that net ecosystem CO_2 exchange shifted from strong carbon uptake to weak carbon uptake during the same period that the soil moisture pulse in the upper 15 cm of soil was decreasing. This suggests that drying soil after the snowmelt pulse caused greater reductions in ecosystem photosynthesis versus ecosystem respiration.

DISCUSSION

The strong correlation between temperatures at Vacas Locas and University Camp SNOTEL sites (Figure 2) suggests that these two regions experience similar annual temperature regimes. The 3 °C difference between the sites is within the range of predicted warming for the region (Intergovernmental Panel on Climate Change, 2007). These strong correlations suggest that these two sites are ideal for comparing differences in the timing of snowmelt onset and associated vegetation response to water availability.

Snow accumulation and ablation

In boreal regions, inverse relationships between accumulation and ablation rates have been observed as snow interception reduces accumulation near trees while enhanced thermal emission from the canopy increases snowmelt rates near the canopy (Faria *et al*., 2000). However, these relationships are dependent upon canopy density and latitude, both of which dictate the effect of vegetation on net radiation and rates of snowmelt (Sicart *et al*., 2004). The observations here indicate that this inverse relationship may not exist at mid-latitudes as we found that both snow accumulation and ablation rates were greater in open areas. The greater accumulation of snow in under-canopy locations at the New Mexico site in 2005–2006 suggests that snow accumulation (and water availability) under the canopy may be less sensitive to mid-winter melt or sublimation and therefore these areas may be less sensitive to shifts in climate.

The role of location (canopy, edge, open) in controlling snow ablation rates was consistent between years. Snow cover persistence was greater in under-canopy locations relative to open areas at the New Mexico site with the opposite true at the Colorado site. Although increases in forest density are known to reduce snowmelt rates at these latitudes (Sicart *et al*., 2004), it is interesting that these two sites behave differently. It is possible that a combination of lower solar radiation associated with the more northern latitude and a more uniform canopy structure at the Colorado site results in greater shading in open areas relative to the open areas at the New Mexico site. The inter-relationship of these two factors suggest that as canopy gap size decreases and as latitude increases overall sub-canopy net radiation increases relative to open areas (Link *et al*., 2004).

Soil moisture, temperature and vegetation response

Soil moisture observations from both sites highlight two critical transitions in the ecohydrology of these seasonally snow-covered forests. First, the timing of snowmelt infiltration onset and peak soil moisture was largely dependent on winter-season snow accumulation amounts and the average winter air temperature—both of which control the cold content of the snowpack. Second, maximum soil moisture roughly coincided with snowpack disappearance. This suggests that soil moisture throughout snowmelt at both study sites was limited by snowmelt rate, not infiltration rate or field capacity. This is consistent with the coincidence of maximum soil moisture at surface and depth. Since snow cover detection and the date of snow disappearance is routinely measured via satellite remote sensing (Dozier and Warren, 1982; Hall *et al*., 1995; Rosenthal and Dozier, 1996; Salomonson and Appel, 2004), future efforts to estimate the timing of peak soil moisture via remote sensing may be fruitful. Given that vegetation response to water availability was highly sensitive to maximum water availability (Figure 10a–f), these future efforts could also provide information on the timing of peak evapotranspiration. Such studies could build on previous efforts to show the impact of earlier snowmelt on ecosystem photosynthesis during the growing season, the overall terrestrial carbon cycle (Sacks *et al*., 2007) and the impact of snowmelt timing on forest fire frequency (Westerling, 2006). In this regard, studies conducted across a variety of biomes and across grassland-forest ecotones (Vivoni *et al*., 2008) will be needed to fully realize these ecohydrological feedbacks.

Similar to other studies, this work highlights the importance of snow cover on soil temperature (Marsh and Woo, 1984; Brooks *et al*., 1996, 1997, 1999; Groffman *et al*., 1999; Fassnacht and Soulis, 2002; Monson *et al*., 2006a,b). The coldest soils observed in our study were at the warmer New Mexico site during the warmest winter (2005–2006) and provide a striking example of 'colder soils in a warmer world' predicted by Groffman *et al*. (1999). The interactions between snowpack development, soil temperature and soil moisture is critical for the balance between mineralization and immobilization of nutrients (Brooks *et al*., 1996, 1997, 1998), nutrient and dissolved organic matter export during melt (Brooks *et al*., 1999), and trace gas production (Brooks *et al*., 1996, 1997; Monson *et al*., 2006a,b). Although the primary focus of this article is on interactions between water from seasonal snow packs and vegetation, feedbacks involving nutrient availability mediated by soil microbial activity before and during snowmelt may affect longer term ecosystem response. The pulse of water into the upper 15 cm of the soil that accompanies snow melt had largely diminished by early summer at the Colorado site (Figure 10f). In fact, by July 15, midway through the summer growth period, soil moisture had returned to pre-snowmelt values. Past studies at this site have shown that most of the fine root biomass is in the upper 15 cm of the soil profile, presumably due to high nutrient availability in the layers with the highest amount of decomposing organic matter, and most of the soil respired CO_2 is produced in these upper layers (Scott-Denton *et al*., 2003). The fact that net ecosystem CO_2 exchange shifted from strong carbon uptake to weak carbon uptake during the same period that the soil moisture pulse in the upper 15 cm of soil was decreasing (Figure 10d and f) suggests that drying soil in the mid-summer causes greater reductions in ecosystem photosynthesis than ecosystem respiration. Furthermore, our observations suggest that much of the snowmelt water disappears as a resource to trees, at least in the shallower soil layers where the bulk of organic decomposition and mineralization occurs, prior to the regular occurrence of mid-summer drought and the late-summer onset of convective monsoon rains

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(Monson *et al.*, 2002). The pulse of snowmelt at these high-elevation sites may, therefore, be critical in determining the degree to which trees are stressed during the mid-summer portion of the growing season before late-summer precipitation events arrive to relieve water stress. The degree to which they are stressed will, of course, be partially offset if they have access to deeper moisture reserves that are recharged during the snowmelt period and remain through the summer. The full rooting profile for the trees at these sites is not known, and is thus a critical piece of information for understanding the importance of snowmelt water on ecosystem function. Similarly, the dynamics of sub-surface flow in these systems are poorly known, motivating studies aimed at estimating flow-paths in mountainous systems (Liu *et al.*, 2004, 2008; Lyon *et al.*, 2008; Molotch *et al.*, 2008).

The relatively short duration study here adequately captured intuitive differences between the sites and between study years. At both sites, significant variability in the duration of the snowmelt season and the timing of precipitation resulted in large differences in water and carbon fluxes during the growing season. These observations suggest that increase in vegetation density may prolong the persistence of snow cover and buffer forest sensitivity to increase in regional air temperature, which act at cross-purpose to diminish snow cover persistence. In this regard, snowmelt partitioning may become increasingly dominated by vegetation water use as increases in forest density prolong the length of the snowmelt season and enhance forest water uptake. This increased partitioning of snowmelt into atmospheric water loss may also lead to reductions in groundwater recharge and surface runoff. For example, flux observations at both sites in 2006–2007 showed that warmer atmospheric conditions during the winter and early spring periods contributed to greater partitioning into latent heat fluxes. Conversely, in 2005–2006 we observed a late season rapid snowmelt pulse at the Colorado site with an overall lower amount of snowmelt partitioned into vapour flux to the atmosphere. Hence, years with warmer spring temperatures may result in more water partitioning into evaporative loss and reduced spring and summer streamflow.

CONCLUSIONS

Although the average snow-season air temperature at the New Mexico site was 3 °C warmer than the Colorado site, the time series of daily average air temperatures at the two sites were highly correlated ($R^2 = 0{\cdot}89$). Comparisons of snow–vegetation interactions at the two study sites indicated that total snow accumulation was greater in open areas versus under-canopy locations. Snow settling and ablation rates were lower in under-canopy locations, effectively prolonging the snowmelt season in some cases. The deeper snowpack at the Colorado site resulted in warmer soil temperatures throughout the winter period despite the colder air temperatures relative to the New Mexico site. Clear relationships between snow accumulation, temperature and soil moisture were observed in which increase (decrease) in snow accumulation and decrease (increase) in air temperature act to increase (decrease) the lag between onset of spring and snowmelt infiltration. Associated temporal dynamics in water availability and partitioning of available energy into sensible and latent heat fluxes were evident, with warmer temperatures and earlier snowmelt onset associated with greater water vapour flux to the atmosphere. Implicit impacts to the terrestrial carbon budget were evident as decreases in soil moisture after peak snowmelt caused greater reductions in ecosystem photosynthesis relative to ecosystem respiration.

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REFERENCES

Baldocchi DD, Hicks BB, Meyers TP. 1988. Measuring Biosphere-atmosphere exchanges of biologically related gases with micrometeorological methods. *Ecology* **69**: 1331–1340.

Bales R, Molotch N, Painter T, Dettinger M, Rice R, Dozier J. 2006. Mountain hydrology of the western United States. *Water Resources Research* **42**: W08432, DOI:08410·01029/02005WR004387.

Brooks PD, McKnight DM, Bencala KE. 1999. The relationship between soil heterotrophic activity, soil dissolved organic carbon (DOC) leachate, and catchment-scale DOC export in headwater catchments. *Water Resources Research* **35**: 1895–1902.

Brooks PD, Schmidt SK, Williams MW. 1997. Winter production of CO2 and N2O from Alpine tundra: Environmental controls and relationship to inter-system C and N fluxes. *Oecologia* **110**: 403–413.

Brooks PD, Vivoni ER. 2008. Mountain ecohydrology: quantifying the role of vegetation in the water balance of montane catchments. *Ecohydrology* **1**: 187–192.

Brooks PD, Williams MW, Schmidt SK. 1996. Microbial activity under alpine snowpacks, Niwot Ridge, Colorado. *Biogeochemistry* **32**: 93–113.

Brooks PD, Williams MW, Schmidt SK. 1998. Inorganic nitrogen and microbial biomass dynamics before and during spring snowmelt. *Biogeochemistry* **43**: 1–15.

Caine N. 1995. Temporal trends in the quality of streamwater in an Alpine environment: Green Lakes valley, Colorado Front Rand, USA. *Geografiska Annaler Series A-Physical Geography* **77A**: 207–220.

Carignan R, D'Arcy P, Lamontagne S. 2000. Comparative impacts of fire and forest harvesting on water quality in Boreal Shield lakes. *Canadian Journal of Fisheries And Aquatic Sciences* **57**: 105–117.

Davis RE, Hardy JP, Ni W, Woodcock CE, Mckenzie JC, Jordan R, Li X. 1997. Variation of snow cover ablation in the boreal forest: A sensitivity study on the effects of conifer canopy. *Journal of Geophysical Research* **102**: 29389–29395.

Dozier J, Warren SG. 1982. Effect of viewing angle on the infrared brightness temperature of snow. *Water Resources Research* **18**: 1424–1434.

Faria DA, Pomeroy JW, Essery RE. 2000. Effect of covariance between ablation and snow water equivalent on depletion of snow-covered area in a forest. *Hydrological Processes* **14**: 2683–2695.

Fassnacht SR, Soulis ED. 2002. Implications during transitional periods of improvements to the snow processes in the land surface scheme—Hydrological model WATCLASS. *Atmosphere-Ocean* **40**: 389–403.

Gelfan AN, Pomeroy JW, Kuchment LS. 2004. Modeling forest cover influences on snow accumulation, sublimation, and melt. *Journal of Hydrometeorology* **5**: 785–803.

Golding DL, Swanson RH. 1978. Snow accumulation and melt in small forest openings in alberta. *Canadian Journal of Forest Research-Revue Canadienne De Recherche Forestiere* **8**: 380–388.

Golding DL, Swanson RH. 1986. Snow distribution patterns in clearings and adjacent forest. *Water Resources Research* **22**: 1931–1940.

Groffman PM, Hardy JP, Nolan S, Fitzhugh RD, Driscoll CT, Fahey TJ. 1999. Snow depth, soil frost and nutrient loss in a northern hardwood forest. *Hydrological Processes* **13**: 2275–2286.

Hall DK, Riggs GA, Salomonson VV. 1995. Development of methods for mapping global snow cover using Moderate Resolution Imaging Spectroradiometer data. *Remote Sensing of Environment* **54**: 127–140.

Hardy JP, Davis RE, Jordan R, Li X, Woodcock C, Ni W, McKenzie JC. 1997. Snow ablation modeling at the stand scale in a boreal jack pine forest. *Journal of Geophysical Research-Atmospheres* **102**: 29397–29405.

Hibbert AR. 1969. Water yield changes after converting a forested catchment to grass. *Water Resources Research* **5**: 634–640.

IPCC. 2007. Climate Change 2007. The Physical Science Basis. Contribution of WGI to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. Solomon S, Qin D, Manning M, Chen Z, Marquis MC, Averyt KB, Tignor M, Miller HL (eds.). Cambridge University Press: Cambridge; 996.

Johnson M. 1994. Changes in Southwestern forests—stewardship implications. *Journal of Forestry* **92**: 16–19.

Kattelmann RC, Berg NH, Rector J. 1983. The potential for increasing streamflow from sierra-nevada watersheds. *Water Resources Bulletin* **19**: 395–402.

Link TE, Flerchinger GN, Unsworth M, Marks D. 2004. Simulation of water and energy fluxes in an old-growth seasonal temperate rain forest using the simultaneous heat and water (SHAW) model. *Journal of Hydrometeorology* **5**: 443–457.

Link TE, Marks D. 1999. Distributed simulation of snowcover mass- and energy-balance in the boreal forest. *Hydrological Processes* **13**: 2439–2452.

Liu FJ, Parmenter R, Brooks PD, Conklin MH, Bales RC. 2008. Seasonal and interannual variation of streamflow pathways and biogeochemical implications in semi-arid, forested catchments in Valles Caldera, New Mexico. *Ecohydrology* **1**: 239–252.

Liu FJ, Williams MW, Caine N. 2004. Source waters and flow paths in an alpine catchment, Colorado Front Range, United States. *Water Resources Research* **40**: W09401 DOI:10.1029/2004WR003076.

Lopez-Moreno JI, Latron J. 2008. Influence of canopy density on snow distribution in a temperate mountain range. *Hydrological Processes* **22**: 117–126.

Lyon SW, Troch PA, Broxton PD, Molotch NP, Brooks PD. 2008. Monitoring the timing of snowmelt and the initiation of streamflow using a distributed network of temperature/light sensors. *Ecohydrology* **1**: 215–224.

Marsh P, Woo MK. 1984. Wetting front advance and freezing of meltwater within a snow Cover.1. Observations In The Canadian Arctic. *Water Resources Research* **20**: 1853–1864.

McDowell NG, White S, Pockman WT. 2008. Transpiration and stomatal conductance across a steep climate gradient in the southern Rocky Mountains. *Ecohydrology* **1**: 193–204.

Molotch N, Bales R. 2005. Scaling snow observations from the point to the grid-element: implications for observaiton network design. *Water Resources Research* **41**: W11421 DOI: 10·1029/2005WR004229.

Molotch NP, Blanken PD, Williams MW, Turnipseed AA, Monson RK, Margulis SA. 2007. Estimating sublimation of intercepted and sub-canopy snow using eddy covariance systems. *Hydrological Processes* **21**: 1567–1575.

Molotch NP, Meixner T, Williams MW. 2008. Estimating stream chemistry during the snowmelt pulse using a spatially distributed, coupled snowmelt and hydrochemical modeling approach. *Water Resources Research* **44**: W11429 DOI: 10.1029/2007WR006587.

Monson RK, Burns SP, Williams MW, Delany AC, Weintraub M, Lipson DA. 2006a. The contribution of beneath-snow soil respiration to total ecosystem respiration in a high-elevation, subalpine forest. *Global Biogeochemical Cycles* **20**: GB3030 DOI: 10.1029/2005GB002684.

Monson RK, Lipson DL, Burns SP, Turnipseed AA, Delany AC, Williams MW, Schmidt SK. 2006b. Winter forest soil respiration controlled by climate and microbial community composition. *Nature* **439**: 711–714.

Monson RK, Turnipseed AA, Sparks JP, Harley PC, Scott-Denton LE, Sparks K, Huxman TE. 2002. Carbon sequestration in a high-elevation, subalpine forest. *Global Change Biology* **8**: 459–478.

Mote PW, Hamlet AF, Clark MP, Lettenmaier DP. 2005. Declining mountain snowpack in western north America. *Bulletin of the American Meteorological Society* **86**: 39–49.

Pataki DE, Oren R, Smith WK. 2000. Sap flux of co-occurring species in a western subalpine forest during seasonal soil drought. *Ecology* **81**: 2557–2566.

Price AG, Dunne T. 1976. Energy-balance computations of snowmelt in a Subarctic Area. *Water Resources Research* **12**: 686–694.

Rosenthal W, Dozier J. 1996. Automated mapping of montane snow cover at subpixel resolution from the Landsat Thematic Mapper. *Water Resources Research* **32**: 115–130.

Sacks WJ, Schimel DS, Monson RK. 2007. Coupling between carbon cycling and climate in a high-elevation, subalpine forest: a model-data fusion analysis. *Oecologia* **151**: 54–68.

Salomonson VV, Appel I. 2004. Estimating fractional snow cover from MODIS using the normalized difference snow index. *Remote Sensing of Environment* **89**: 351–360.

Schotanus P, Nieuwstadt FTM, Debruin HAR. 1983. Temperature-measurement with a sonic anemometer and its application to heat and moisture fluxes. *Boundary-Layer Meteorology* **26**: 81–93.

Scott-Denton LE, Sparks KL, Monson RK. 2003. Spatial and temporal controls of soil respiration rate in a high-elevation, subalpine forest. *Soil Biology & Biochemistry* **35**: 525–534.

Sicart JE, Pomeroy JW, Essery RE, Hardy J, Link T, Marks D. 2004. A sensitivity study of daytime net radiation during snowmelt to forest canopy and atmospheric conditions. *Journal of Hydrometeorology* **5**: 774–784.

Stednick JD. 1996. Monitoring the effects of timber harvest on annual water yield. *Journal of Hydrology* **176**: 79–95.

Stewart IT, Cayan DR, Dettinger MD. 2004. Changes in snowmelt runoff timing in western North America under a 'business as usual' climate change scenario. *Climatic Change* **62**: 217–232.

Turnipseed AA, Blanken PD, Anderson DE, Monson RK. 2002. Energy budget above a high-elevation subalpine forest in complex topography. *Agricultural and Forest Meteorology* **110**: 177–201.

Vivoni ER, Rinehart AJ, Mendez-Barroso LA, Aragon CA, Bisht G, Cardenas MB, Engle E, Forman BA, Frisbee MD, Gutierrez-Jurado HA, Hong SH, Mahmood TH, Tai K, Wyckoff RL. 2008. Vegetation controls on soil moisture distribution in the Valles Caldera, New Mexico, during the North American monsoon. *Ecohydrology* **1**: 225–238.

Webb EK, Pearman GI, Leuning R. 1980. Correction of flux measurements for density effects due to heat and water-vapor transfer. *Quarterly Journal of the Royal Meteorological Society* **106**: 85–100.

Westerling AL, Hidalgo HG, Cayan DR, Swetnam TW. 2006. Warming and Earlier Spring Increases Western U.S. Forest Wildfire Activity. *Science* **313**: 940–943.

Wigmosta MS, Vail LW, Lettenmaier DP. 1994. A distributed hydrology-vegetation model for complex terrain. *Water Resources Research* **30**: 1665–1679.

Woo MK, Giesbrecht MA. 2000. Simulation of snowmelt in a subarctic spruce woodland: Scale considerations. *Nordic Hydrology* **31**: 301–316.

Zehe E, Bloschl G. 2004. Predictability of hydrologic response at the plot and catchment scales: Role of initial conditions. *Water Resources Research* **40**: W10202 DOI: 10.1029/2003WR002869.