Estimating Sublimation of Intercepted and Sub-Canopy Snow Using Eddy Covariance Systems

NOAH MOLOTCH[1], PETER BLANKEN[2], MARK WILLIAMS[2,3], ANDREW TURNIPSEED[4], RUSSELL MONSON[5], AND STEVEN MARGULIS[1]

ABSTRACT:

Direct measurements of winter water loss due to sublimation were made in a sub-alpine forest in the Rocky Mountains of Colorado. Sub-canopy and over-story eddy covariance systems indicated substantial losses of winter-season snow accumulation in the form of snowpack (0.41 mm d^{-1}) and intercepted snow (0.71 mm d^{-1}) sublimation. The partitioning between these over and under story components of water loss was highly dependent on atmospheric conditions and near-surface conditions at and below the snow / atmosphere interface. High over-story sensible heat fluxes lead to strong temperature gradients between vegetation and the snow-surface, driving substantial specific humidity gradients at the snow surface and high sublimation rates. Intercepted snowfall resulted in rapid response of over-story latent heat fluxes, high within-canopy sublimation rates, and diminished sub-canopy snowpack sublimation. These results indicate that sublimation losses from the under-story snowpack are strongly dependent on the partitioning of sensible and latent heat fluxes in the canopy. This compels comprehensive studies of snow sublimation in forested regions that integrate sub-canopy and over-story processes.

Keywords: vegetation canopy; snow interception; sublimation; Rocky Mountains; eddy covariance

INTRODUCTION

Sublimation of intercepted snow constitutes a significant component of the overall water balance in many seasonally snow-covered coniferous forests [*Essery, et al.*, 2003; *Lundberg and Halldin*, 1994; *Pomeroy and Gray*, 1995; *Schmidt and Troendle*, 1992]; sublimation losses are capable of exceeding 30% of total winter snowfall [*Montesi, et al.*, 2004]. For a given canopy structure and snowfall history the distribution of radiant and turbulent fluxes dictates sublimation rates and therefore strongly influences the magnitude of spring snowmelt and subsequent growing-season water availability. Interactions between these fluxes and the sublimation of intercepted snow and the sub-canopy snowpack are poorly understood in forested mountainous regions [*Bales, et al.*, 2006]. This knowledge gap and the complexity of interactions between the snowpack and vegetation have motivated detailed analyses of mass and energy fluxes between the snowpack, vegetation, and the atmosphere [*Davis, et al.*, 1997; *Sicart, et al.*, 2004].

[1]Department of Civil and Environmental Engineering, University of California, Los Angeles, California, 90095.
[2]Department of Geography, University of Colorado, Boulder, Colorado, 80309.
[3]Institute for Arctic and Alpine Research, University of Colorado, Boulder, Colorado 80309.
[4]National Center for Atmospheric Research, Boulder, Colorado, 80305.
[5]Department of Ecology and Evolutionary Biology; Cooperative Institute for Research in Environmental Sciences, University of Colorado, Boulder, Colorado 80309

Various techniques have been used to estimate sublimation rates from intercepted snow. Measurement of the components of snow sublimation is particularly challenging in forested terrain as winter-time above-canopy water vapor flux measurements integrate mass loss from intercepted snow and from the sub-canopy snowpack. In this regard, numerous studies have focused on estimating sublimation losses from snowpacks in unforested areas. Similarly, much work has been devoted toward estimating sublimation losses from intercepted snow [*Montesi, et al.*, 2004; *Pomeroy and Schmidt*, 1993; *Schmidt and Troendle*, 1992]. Lacking is a thorough analysis of the proportion of these two different components of snow sublimation at an individual site.

Measurement of sublimation from intercepted snow has primarily focused on tree-weighting techniques [*Montesi, et al.*, 2004; *Nakai, et al.*, 1994; *Schmidt*, 1991; *Schmidt, et al.*, 1988]. Several factors lead to uncertainty in this approach and toward limiting applicability at the stand scale. First, a somewhat subjective analysis must be used to separate unloading from sublimation. Second, sublimation of unloaded snow is not considered and thus sublimation losses may be underestimated [*Montesi, et al.*, 2004]. Third, tree-instability can cause false readings. Finally, intermittent snowfall events and small trace events can introduce uncertainty, effectively countering sublimation losses and leading to underestimates in sublimation losses if not considered. In terms of scaling from individual trees to the stand scale, challenges are encountered with regard to the lack of detailed canopy information. This lack of detailed canopy information also complicates the use of models for estimating sublimation losses [*Pomeroy, et al.*, 1998; *Pomeroy and Schmidt*, 1993]. All of these limitations could be accounted for in techniques that integrate all of these processes by measuring above and below canopy water vapor flux.

Advances in process-level knowledge have been limited as sublimation can occur either from snow intercepted by the canopy, and/or from the snow that reaches the ground. Coniferous forests can intercept large quantities of snow, much of which sublimates from the canopy and does not reach the ground. Sublimation from the below-canopy snowpack is thought to be insignificant due to the low exposed surface area of the snowpack and low below-canopy wind speeds. However, there are potentially large longwave radiation fluxes if the canopy above is warm and snow-free, thus promoting sublimation and/or melting [*Woo and Giesbrecht*, 2000]. Understanding the balance between sublimation from the canopy and snowpack is crucial to assist water and forest managers, especially in regions where forest thinning treatments are being considered to increase water yield.

Direct measurements of winter water loss by sublimation of snow from a subalpine forest in the Rocky Mountains of Colorado are presented here. Eddy covariance instruments were placed both above and beneath the canopy during March and early April 2002; the time before melting begins when winter sublimation is thought to be large due to the heavy late-winter snows. The above and below-canopy measurements allowed sublimation of intercepted snow to be separated from that of the snowpack, and estimates obtained over a much larger sample area than individual trees. Simultaneous measurements of the physical properties of the snow pack, soil moisture, as well as carbon dioxide flux measurements ensured that sublimation and not evaporation of melting snow or transpiration were being measured. The specific objectives of this research were to: a) determine snow sublimation rates in a sub-alpine forest; b) partition snow sublimation into above and below canopy components; and c) explore relationships between atmospheric and snowpack conditions, and snow sublimation rates.

STUDY SITE

This work was conducted at the Niwot Ridge, Colorado Ameriflux site (40º 1’ 58”N; 105º 32’ 47” W), located at an elevation of 3050 m approximately 8 km east of the Continental Divide (Figure 1). The area 1 km^2 east of the tower is dominated by Engelmann spruce (7 trees ha^{-1}) and lodgepole pine (27 trees ha^{-1}). Rising at a slope of about 6 – 7º, the 1 km^2 area west of the tower contains subalpine fir (16 trees ha^{-1}), Engelman Spruce (10 trees ha^{-1}) and lodgepole pine (9 trees ha^{-1}). Maximum leaf area index during the growing season is approximately 4.2 m^2 m^{-2}. Canopy height averaged 11.4 m with an average gap fraction of 17%. The site is in a state of aggradation, recovering from logging activities in the early part of the 20th century. The hydrology of the site is dominated by moderate snowpacks that account for approximately 80% of total annual water input to the system [*Caine*, 1995]. The prevailing wind direction is from the west, particularly in the winter when periods of high wind speeds and neutral atmospheric stability conditions are frequent [*Turnipseed, et al.*, 2002]. A detailed description of the site characteristics can be found in Turnipseed et al. [2002].

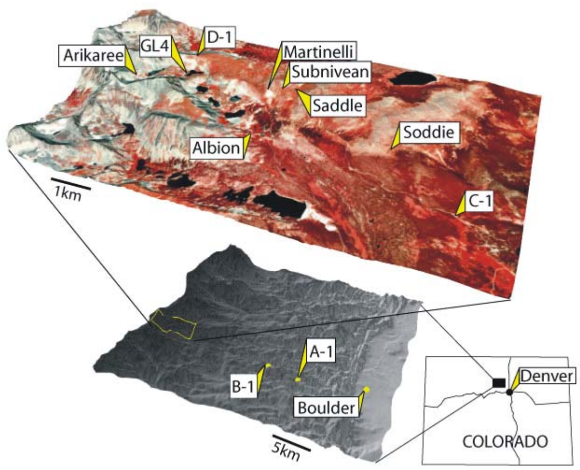

Figure 1. Composite image of Niwot Ridge, LTER site and the CU-Ameriflux tower located at C-1.

METHODS

Flux measurements

Water vapor fluxes, (latent heat flux; λE) were calculated as 30-min means of 10-Hz measurements over a 40 d mid-winter period (DOY 60 – 100, 2002) using the eddy covariance (EC) method described by Turnipseed et al. [2002]:

$$\lambda E = L_v \overline{w' \rho_v'} \tag{1}$$

where L_v is the latent heat of sublimation, w' is the deviations of vertical wind velocity (m s^{-1}) from the ½-hr mean, ρ_v' is the deviations of the water vapor density from the ½-hr mean. The over-story and under-story EC systems were mounted at a height of 21.5 m and 1.7 m above-ground, respectively, from towers separated by a distance of approximately 20-m. The over- and under-story EC systems and other meteorological instruments are summarized in Table 1.

Components of snow sublimation were computed as:

$$\lambda E_{c,t} = \lambda E_{c,s} + \lambda E_{c,i} \tag{2}$$

where $\lambda E_{c,t}$ is the total sublimation from the system measured using the over-story EC instruments (21.5 m above ground) and $\lambda E_{c,s}$ is snowpack sublimation determined from the sub-canopy EC instruments (1.7 m above ground). Water vapor fluxes associated with sublimation of intercepted snow, $\lambda E_{c,i}$ were determined as the difference of measured over-story and sub-canopy fluxes. Measurements of the above-canopy CO_2 flux were used to confirm that photosynthesis from the forest canopy was negligible (i.e. values were positive indicating canopy respiration but no carbon uptake) and therefore over-story water flux observations could be inferred to be entirely associated with snow sublimation since transpiration was insignificant.

Atmospheric stability was calculated by dividing the Monin-Obukhov length, L [*Monin and Obukhov*, 1954] into the measurement height (z):

$$L = \frac{u*^3 \times z \times \rho(T) \times c_p(e,p) \times T}{k \times z \times g \times H} \tag{3}$$

where $u*$ is the friction velocity (m s^{-1}), $\rho(T)$ is the air density as a function of air temperature (T) (Kelvin), c_p is the specific heat of dry air (kJ kg^{-1} K^{-1}) as a function of vapor pressure, e (kPa), and barometric pressure, p (kPa), k is von Karman's constant (0.41), g is acceleration due to gravity 9.81 (m s^{-2}), and H is the sensible heat flux (W m^{-2}). Negative z/L values correspond to unstable atmospheric conditions, positive values represent stable conditions, and values near 0 are neutral.

Table 1. Observations and instruments on the above and below canopy towers at the University of Colorado, Ameriflux site.

observation	measurement height, meters	instrument
relative humidity, %	21.5	HMP-35D, Vaisala, Inc.
air temperature, ºC	21.5 | 1.7	CSAT-3, Cambell Scientific
pressure, kpa	18	PT101B, Vaisala, Inc.
net radiation, W m^{-2}	26	4-component CNR-1, Kipp & Zonen
H_2O flux, mg m^{-2} s^{-1}	21.5 | 1.7	IRGA-6260, Li-Cor
CO_2 flux, mg m^{-2} s^{-1}	21.5	IRGA-6260, Li-Cor
wind speed, m s^{-1}	21.5 | 1.7	propvane-09101, RM Young Inc.
wind direction, degrees	21.5 | 1.7	propvane-09101, RM Young Inc.
precipitation, mm	12	385-L, Met One
soil heat flux, W m^{-2}	-0.07 - -0.1	HFT-1, REBS
soil moisture, % by volume	0 - -.15	CS-615, Campbell Scientific
soil temperature, ºC	0- - 0.1	STP-1, REBS

Note: Above and below canopy eddy covariance systems were located 21.5 and 1.7 m above ground, respectively.

Turbulent flux estimates were evaluated by exploring total energy balance closure; turbulent fluxes should be equal to the available energy. A linear regression between the summation of the sensible (H) and latent heat fluxes and the difference between the net radiation (R_n) and ground (G) heat flux was developed [*Blanken, et al.*, 1998; *Blanken, et al.*, 1997]. The relationship between the 30-min above canopy (λE +H) and (Rn-G) was y = 0.77x + 13 ($R^2 = 0.89$; $p < 0.01$) indicating adequate energy balance closure.

The sampling area, or flux footprint, was calculated using the method described by Schuepp et al. [*Schuepp, et al.*, 1990]. The upwind distance that the understory flux measurements were most sensitive to occurred at a distance of 23, 27, and 29-m during typical daytime, neutral, and nighttime atmospheric stability conditions, respectively (Figure 2a). The cumulative flux footprint, indicative of the upwind sampling area where 80% of the flux originated from, was 207, 243, and 263 m (daytime, neutral, and nighttime atmospheric stability conditions, respectively) (Figure 2b).

Supporting Understory Measurements

Observations of soil, snow, and air temperature from three thermistor strings were used to develop relationships between snowpack temperature and rates of snowpack sublimation. In this regard, we investigated relationships between snowpack temperature gradients and diurnal variability in snow temperature, and rates of snowpack sublimation; snowpack temperature gradients control vapor pressure gradients in the snowpack and therefore the movement of water vapor from deeper in the snowpack toward the snowpack / atmosphere interface [*McClung and Schaerer*, 1993]. The three thermistor strings were placed along a transect through a small clearing (~ 6 m in diameter) in the forest adjacent to the understory flux tower (Figure 3). The thermistor strings were buried 20 – 30 cm below the soil before snow accumulation began and extended to 80, 180, and 200 cm above the ground surface; a guy wire tied to two trees at opposite ends of the clearing was used to tether the tops of the thermistor strings. During the study period the thermistor strings provided observations of soil, snow, and air temperature.

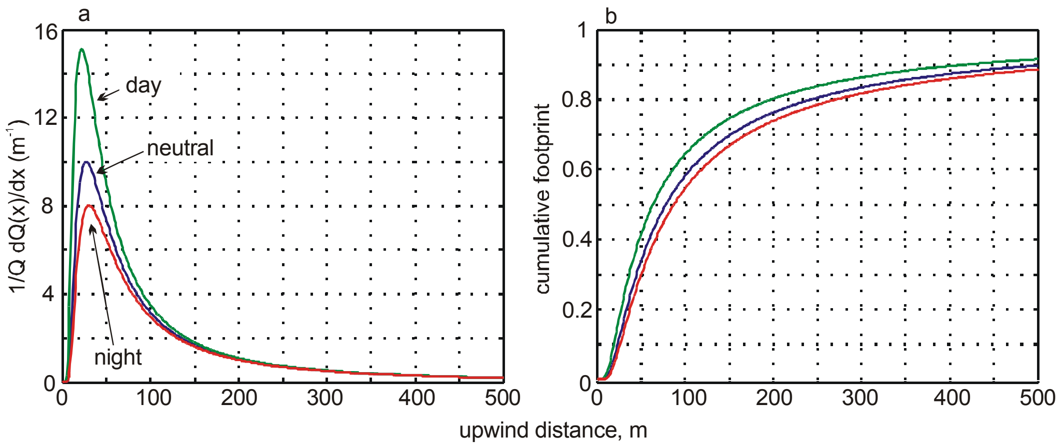

Figure 2. (a) Normalized change in the understory turbulent flux (Q) with upwind distance (x) during typical daytime (green), neutral (blue) and nighttime (red) atmospheric stability conditions. (b) Cumulative change in the understory turbulent flux (Q) with upwind distance (x) for typical daytime (green), neutral (blue) and nighttime (red) atmospheric stability conditions.

Eight water content reflectometers (Campbell Scientific model CS-615) were used to monitor soil moisture conditions surrounding the towers (Figure 3). These observations were used to ensure that latent heat fluxes were primarily allocated to sublimation as opposed to snowmelt and to confirm that water from snowmelt had not entered the soil horizon which might trigger the onset of transpiration.

Snowpack properties

Ground observations of snow depth and snow density were derived from snow pits excavated weekly at two different locations (sub-canopy and within a small clearing adjacent to the flux towers). Within each snowpit, samples were taken at 10 cm vertical intervals over the entire snowpit depth using a 1000 cc stainless steel cutter. Snow density stratigraphy and bulk density and snow water equivalent were calculated from weighted-average density values and total snowpack depth.

Observations of precipitation were used to determine the mass input between the weekly snowpit observations, allowing us to approximate sublimation losses; changes in snow water equivalent between the weekly snowpit observations result from input of mass due to snowfall and reduction in mass due to sublimation. This provides a field based technique for evaluating sublimation estimates from the sub-canopy EC system. Precipitation observations were obtained at a height of 12 m from the above-canopy EC tower; an Alter gauge shield was used to improve precipitation gauge catch efficiency.

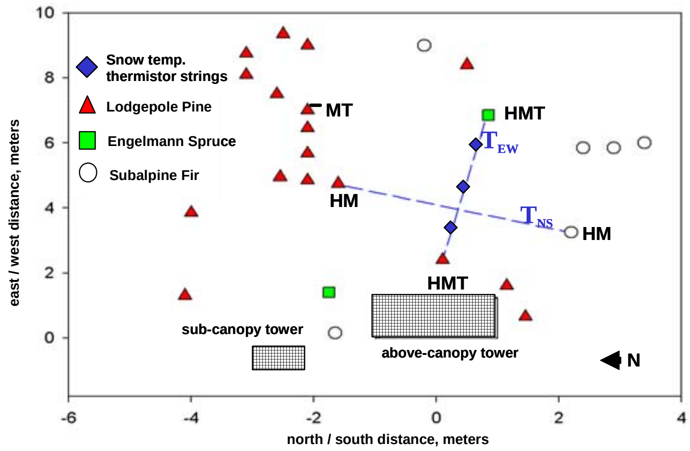

Figure 3. Location of above and below canopy flux towers and supporting ground based instruments. Water content reflectometers (M), thermistor strings, ground thermistors (T), and soil heat flux plates (H). Additional H, M and T measurements were made along east / west (T_{ew}) and north / south (T_{NS}) transects.

RESULTS

Soil temperature, moisture and ground heat flux were consistent with mid-winter conditions throughout the study period (Figure 4). Temporal variability in soil temperature (coefficient of variation = .77) and ground heat flux (coefficient of variation = 2.6) was considerably greater than that of soil moisture (coefficient of variation = 0.11). Spring onset of snowmelt percolation occurred on DOY 100; soil moisture increased by threefold over the subsequent 20-d period.

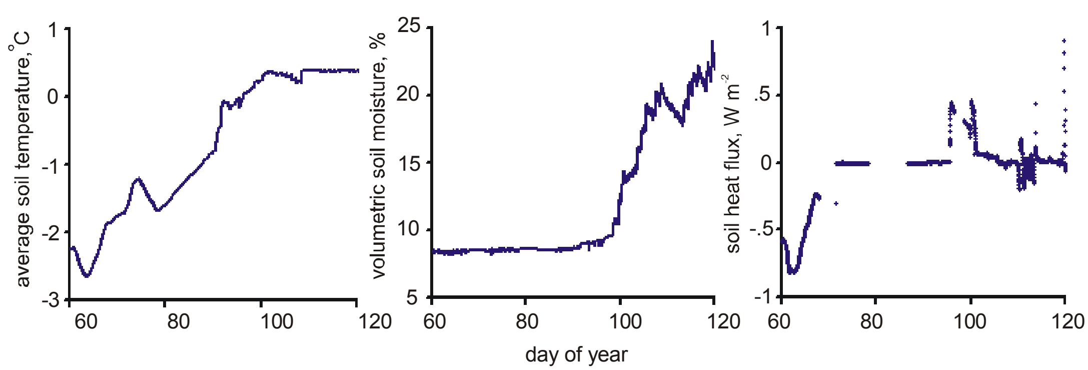

Figure 4. Time series of soil temperature, moisture, and ground heat flux from day of year 60 – 120, 2002.

The diurnal energy fluxes, R_n, λE and H above and below the canopy are shown in Figure 5, together with precipitation. The CO_2 flux above the canopy is included to show that the forests had not yet transitioned from losing to gaining carbon, and therefore transpiration at this time was negligible. The majority of the above-canopy net radiation was partitioned as H above the canopy, and as λE beneath the canopy; above-canopy ratios of the daytime mean H/Rn and λE/Rn were 0.67 and 0.16, respectively. Beneath the canopy, these ratios were 0.02 (H/Rn) and 0.06 (λE/Rn).

Although the λE/Rn fraction was on average relatively small, large increases in λE with a subsequent decrease in H occurred several times in response to snowfall events.

Average sublimation rates over the study period were 0.7 and 0.41 mm d^{-1} for intercepted snow and the sub-canopy snowpack, respectively. Both fluxes exhibited considerable variability (coefficient of variation = 0.66 for both total sublimation and snowpack sublimation), with intercepted snow sublimation rising after snowfall events (Figure 6). The ratio between sub-canopy snowpack sublimation and total sublimation averaged 0.45 during the study period, increasing with time after snowfall and approaching 1 during consecutive days without snowfall; e.g. DOY 63 – 65 and DOY 87 – 93 (Figure 6). On average snowpack to total sublimation ratios peaked 3 days after snowfall; timing to peak varied considerably with snowfall magnitude.

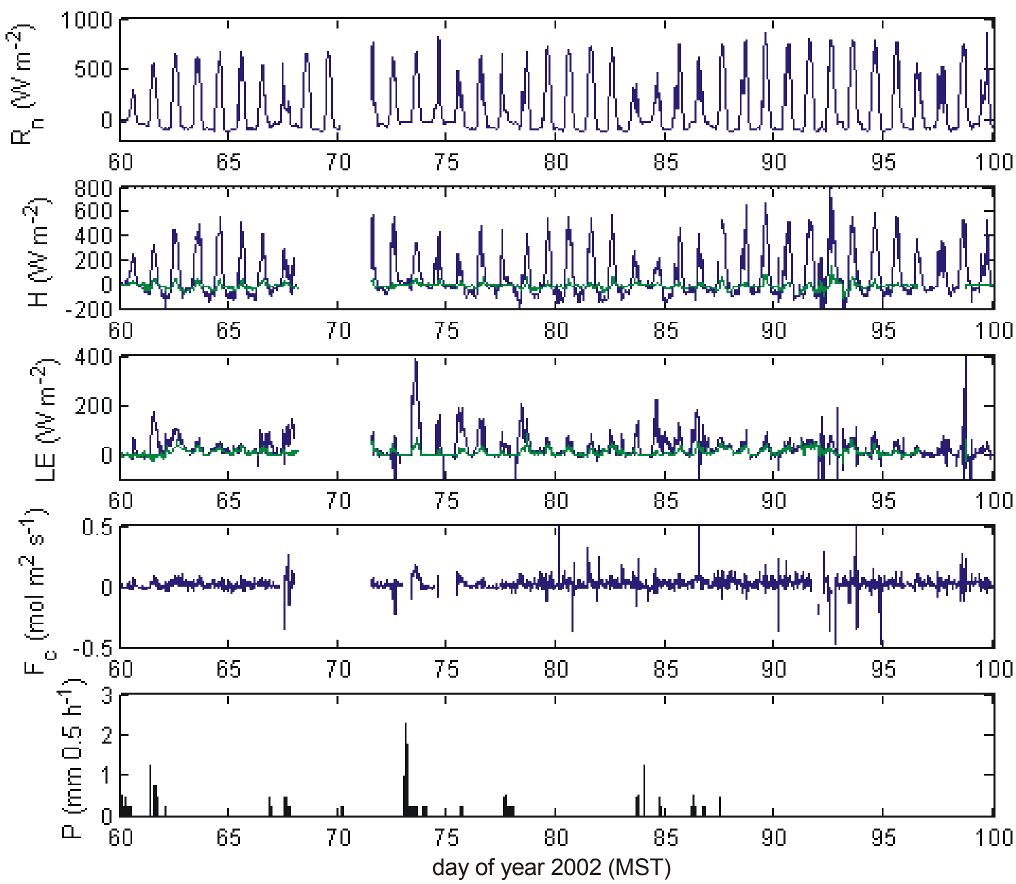

Figure 5. Diurnal variability in net radiation (R_n), sensible (H) and latent (λE) heat fluxes, carbon flux (F_c), and precipitation (P) measured above the canopy (blue lines) and beneath the canopy (green lines). Period shown is from DOY 60 – 100 2002. Positive values represent fluxes toward the surface.

A total of 34.8 mm of snow fell during the measurement period (Fig 7). 38.5 mm of sublimation was measured above the canopy over the same time period, and 14.8 mm sublimated from the snowpack at the forest floor. These correspond to sublimation to precipitation ratios of 1.11 (total) and 0.43 (snowpack), with the total ratio exceeding one due to sublimation of snow that fell prior to the start of the measurements. Subtracting the above-canopy λE measurements from that below the canopy (Figure 7) reveals that 23.7 mm of intercepted snow was sublimated from the canopy itself. This corresponds to a sublimation to precipitation ratio of 0.68.

Diurnal fluctuations in snowpack and near surface air temperatures were notably different for time periods with high snowpack sublimation rates. For example, only 0.1 mm of water sublimated from the snowpack on DOY 60 whereas over 0.6 mm sublimated on DOY 64. At 60 cm above the ground surface snow temperatures fluctuated by less than 5˚ during DOY 60 and by more than 10˚ during DOY 64 (Figure 8). Similarly, diurnal variability in snow temperature was significantly lower on DOY 74 relative to DOY 93; sublimation rates were 0.1 versus 0.6 mm d^{-1} for these two days, respectively. Temperature fluctuations in the surface layers, associated with cool nights and warm dry days potentially drive significant water vapor movement in the surface layers of the snowpack and enhance sublimation rates.

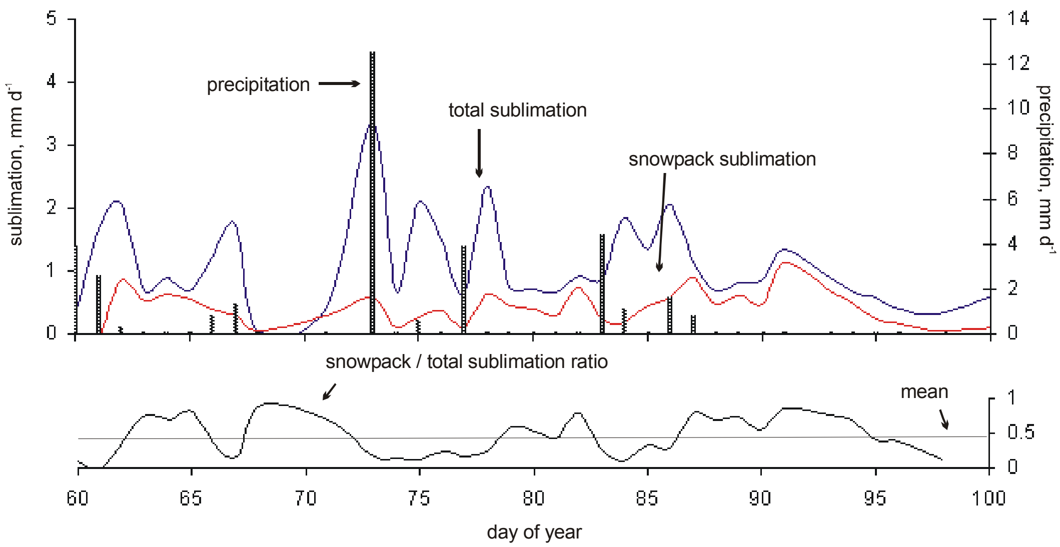

Figure 6. Time series of daily average sublimation measured above the canopy (blue line) and beneath the canopy (red line). Precipitation and the ratio of snowpack sublimation to total sublimation are also shown.

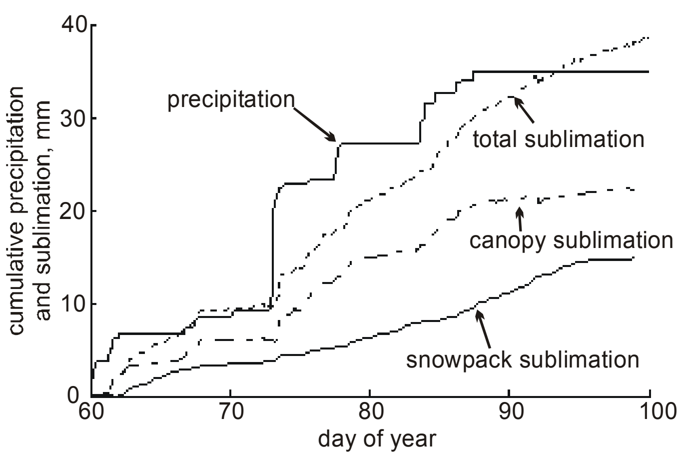

Figure 7. Cumulative sublimation from the snowpack, intercepted snow (canopy), and total sublimation throughout the study period. Cumulative precipitation is also shown.

Estimates of snow depth on snow temp profile plots were derived from coincident pit observations when available. In the case of DOY 60, the majority of precipitation was recorded on DOY 59 and early hours of DOY 60 and therefore we assume snow depth equivalent to that measured in the snowpit on DOY 64. In the case of DOY 74, snow depth was difficult to estimate as there was a large (12.45 mm) snowfall event on DOY 73. Thus, we assumed a snow depth of 80 cm, corresponding to the observed snow depth from the snowpit on DOY 84. For DOY 93, we estimated snow depth based on the 2:00 temperature curve which showed a distinct inflection point at the snow–atmosphere interface.

Above and below canopy friction velocities were considerably greater for DOY 64 and 93 relative to that on DOY 60 and 74 (Figure 9). The combination of the relatively high air temperatures on these days with sufficient turbulence lead to enhanced near-surface gradients in specific humidity and sublimation.

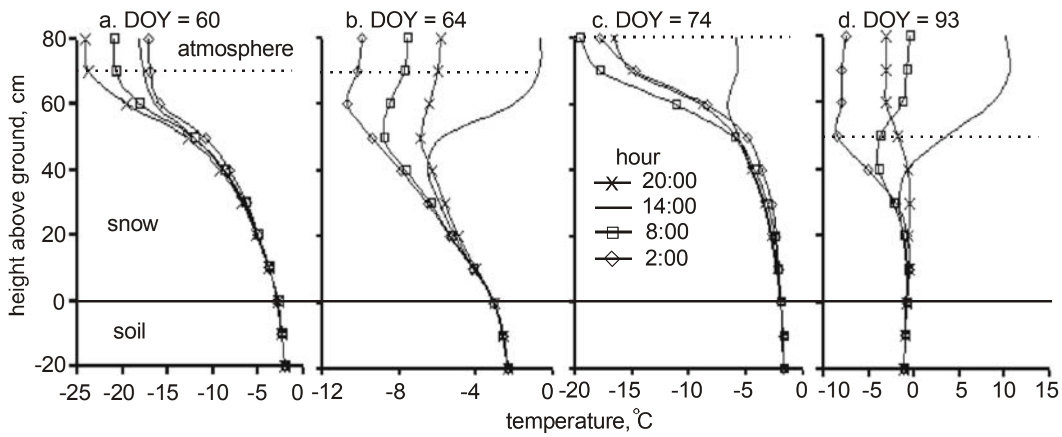

Figure 8. Snowpack, soil, and air temperature profiles from 80 cm above the ground surface to 20 cm below. Profiles are shown for 4 different days for hours 2, 8, 14, and 20 MST. Dotted horizontal lines represent the snowpack / atmosphere interface. Solid horizontal lines indicate the soil / snowpack interface.

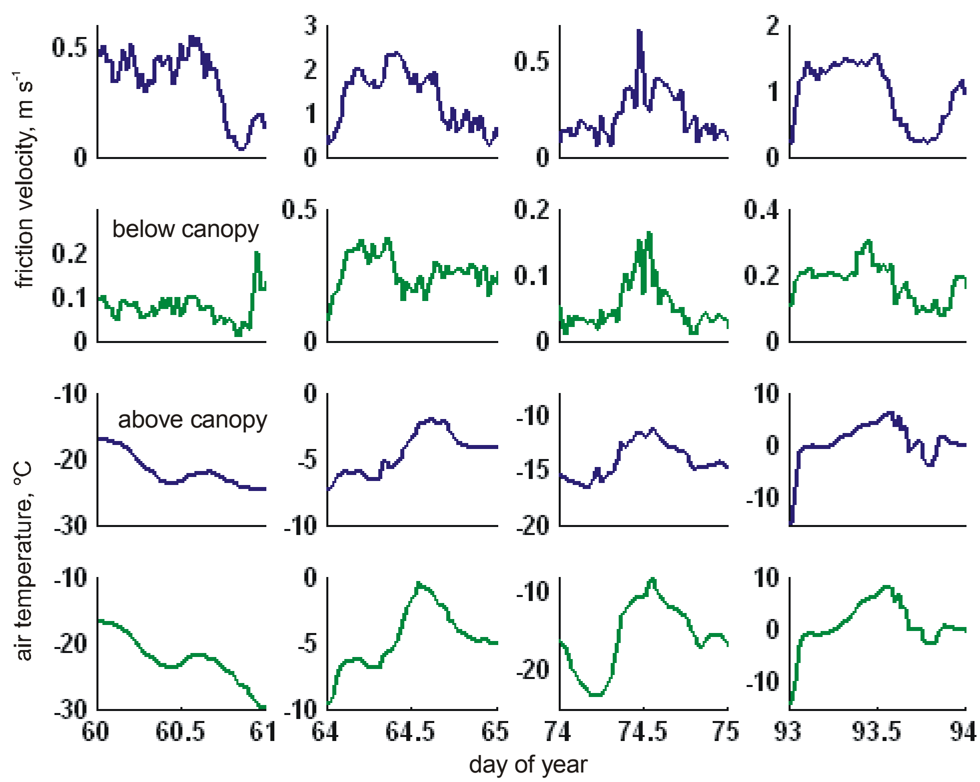

Figure 9. Above (blue lines) and below canopy (green lines) diurnal variability in friction velocity and air temperature for the same 4 days shown in Figure 8.

Unstable atmospheric conditions resulted in considerable sublimation of intercepted snow. For example, on DOY 62 measurement-height / Monin-Obukhov ratios dropped below –200 (Figure 10) and daily sublimation was 2.09 mm (Figure 6). Conversely, measurement-height / Monin-Obukhov ratios on DOY 76 were less than 0 but greater than –3, suggesting only slight atmospheric instability. Sublimation of intercepted snow on DOY 76 was 1.74 mm; only 17% lower than that of DOY 62. Precipitation magnitude is likely responsible for these differences with 12 mm of precipitation falling on DOY 73 and a combined 6 mm of precipitation falling over the course of DOY 60 and 61 (Figure 5).

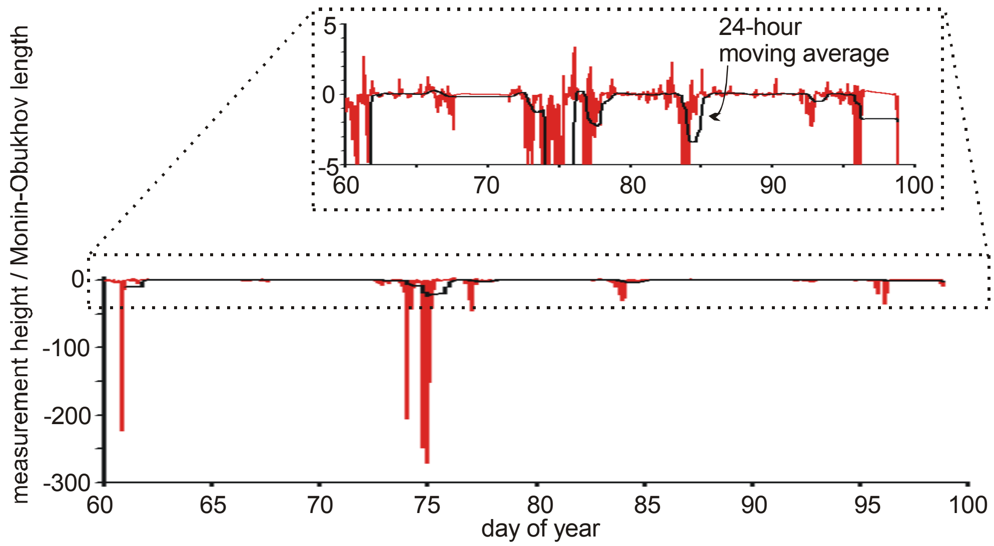

Figure 10. Measurement-height / Monin-Obukhov-length ratios calculated from 30-minute, above-canopy observations. See equation (3) for derivation of Monin-Obukhov length. Negative and positive values represent unstable and stable conditions, respectively.

DISCUSSION

A variety of techniques have been developed to estimate sublimation from snowpacks and intercepted snow [*Montesi, et al.*, 2004; *Pomeroy, et al.*, 1998]. It is especially challenging to capture the impact of vegetation on variability in turbulence and subsequent vapor fluxes. Results of previous work performed at the individual tree scale provide useful values to evaluate results of our new technique. Comparisons, however, must be made with caution as our technique integrates fluxes over the stand scale from two systems with different flux footprints (despite reasonably homogenous stand characteristics); tree scale studies provide limited information at the stand scale due to introduction of uncertainty associated with vegetation properties. Further, quantitative comparison with previous studies is difficult given that meteorological conditions and site specific attributes can have dramatic impacts on the energy balance of forested environments – in particular, variability in vegetation structure [*Sicart, et al.*, 2004]. Here we compare general observations of both snowpack and intercepted sublimation rates. Average mid-winter snowpack sublimation rates observed here (0.41 mm d^{-1}) were low relative to the highest of values found within the literature (1.2 – 1.8 mm d^{-1} [*Pomeroy and Essery*, 1999]) and are within 14% of values observed at the nearby Fraser Experimental Forest (e.g. 0.36 mm d^{-1}) [*Schmidt, et al.*, 1998]. Fassnacht [*Fassnacht*, 2004] estimated winter sublimation rates at 0.75 mm d^{-1} at an open site in Leadville, Colorado; open sites are known to exhibit substantially greater sublimation rates [*West*, 1962]. The weekly snowpits excavated in a clearing adjacent to the flux towers used in this research indicated a total sublimation rate of 0.8 mm d^{-1}. While these estimates have inherent uncertainties, these on-site observations and comparisons with previous studies indicate that sublimation rates are not being overestimated using the sub-canopy EC system. In this regard, it is important to note that the average snowpack to total sublimation ratio of 0.45 (Figure 6) represents the low-end of the contribution of sub-canopy sublimation to overall water loss; a significant finding given previous assumptions that sublimation losses in forested systems are primarily the result of intercepted snow sublimation [*Montesi, et al.*, 2004].

The assessment of sub-canopy sublimation estimates mentioned above must be considered when evaluating the EC estimates of intercepted snow sublimation as they are calculated from the residual of total sublimation and sub-canopy sublimation (equation (2)). Sublimation rates of intercepted snow estimated using our EC approach (0.71 mm d^{-1}) compared favorably with previous works. For example, Parviainen [*Parviainen and Pomeroy*, 2000] estimated intercepted snow sublimation from a boreal forest at 0.5 mm d^{-1}; at higher latitudes available energy is diminished due to higher solar zenith angles.

Montesi et al., [2004] explored the impact of elevation on sublimation rates and found that increased wind speeds, lower relative humidity and warmer air temperatures contributed to a 23% increase in sublimation rates at lower elevation. On average Montesi's results indicate considerable differences between our estimates, with sublimation losses equivalent to 20 – 30% of total snow water equivalent during the 21 storms considered. These differences may be due to an underestimate in sub-canopy sublimation from the sub-canopy EC system used here. Differences may also be due to previously mentioned sources of underestimates in sublimation using the tree-weighting method of Montesi [2004].

As previously found by Niu and Yang [2004], the relatively high over-story sensible heat fluxes lead to strong temperature differences between vegetation and the snow-surface, driving strong specific humidity gradients at the snow / atmosphere interface and elevated snowpack sublimation rates (e.g. DOY 78 – 82 & 88 – 95, Figure 6). When snowfall occurred, over-story available energy was partitioned into latent heat fluxes (e.g. Figure 5, DOY 74), leading to high within-canopy sublimation rates but diminished diurnal variability in temperatures at the snow – atmosphere interface (DOY 74, Figure 8). These results indicate that sublimation losses from the under-story snowpack is strongly dependent on the partitioning of sensible and latent heat fluxes in the canopy.

CONCLUSIONS

Sub-canopy and over-story eddy covariance systems indicated substantial losses of winter-season snow accumulation in the form of snowpack (0.41 mm d^{-1}) and intercepted snow (0.71 mm d^{-1}) sublimation. The partitioning between these over and under story components of water loss was highly dependent on atmospheric conditions and near-surface conditions at and below the snow – atmosphere interface. High over-story sensible heat fluxes lead to strong temperature gradients between vegetation and the snow-surface, driving substantial specific humidity gradients at the snow surface and high sublimation rates. Intercepted snowfall resulted in rapid response of over-story latent heat fluxes, high within-canopy sublimation rates, and diminished sub-canopy snowpack sublimation. These results indicate that sublimation losses from the under-story snowpack is strongly dependent on the partitioning of sensible and latent heat fluxes in the canopy. This compels comprehensive studies of snow sublimation in forested regions that integrate sub-canopy and over-story processes.

ACKNOWLEDGEMENTS

This research was supported by NASA grant #NNG04GO74G and a visiting research fellowship awarded to the primary author at the Cooperative Institute for Research in Environmental Sciences, University of Colorado at Boulder. Instrumentation infrastructure was developed in part through the Ameriflux network and through support of the National Science Foundation, Niwot Ridge LTER Project. Phillip Jacobson provided technical support. Andrew Rossi is acknowledged for collecting snow pit data.

REFERENCES

Bales, R., N. Molotch, T. Painter, M. Dettinger, R. Rice, and J. Dozier (2006), Mountain hydrology of the western United States., Water Resources Research, 42, W08432, doi:08410.01029/02005WR004387.

Blanken, P. D., T. A. Black, H. H. Neumann, G. Den Hartog, P. C. Yang, Z. Nesic, R. Staebler, W. Chen, and M. D. Novak (1998), Turbulent flux measurements above and below the overstory of a boreal aspen forest, Boundary-Layer Meteorology, 89, 109–140.

Blanken, P. D., T. A. Black, P. C. Yang, H. H. Neumann, Z. Nesic, R. Staebler, G. den Hartog, M. D. Novak, and X. Lee (1997), Energy balance and canopy conductance of a boreal aspen forest: Partitioning overstory and understory components, Journal Of Geophysical Research-Atmospheres, 102, 28915–28927.

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.

Davis, R. E., J. P. Hardy, W. Ni, C. E. Woodcock, J. C. Mckenzie, R. Jordan, and X. Li (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.

Essery, R., J. Pomeroy, J. Parviainen, and P. Storck (2003), Sublimation of snow from coniferous forests in a climate model, Journal Of Climate, 16, 1855–1864.

Fassnacht, S. R. (2004), Estimating Alter-shielded gauge snowfall undercatch, snowpack sublimation, and blowing snow transport at six sites in the coterminous USA, Hydrological Processes, 18, 3481–3492.

Lundberg, A., and S. Halldin (1994), Evaporation of intercepted snow: Anlaysis of governing factors., Water Resources Research, 30, 2587–2598.

McClung, D., and P. Schaerer (1993), The Avalanche Handbook, 271 pp., The Mountaineers, Seattle, Washington.

Monin, A. S., and A. M. Obukhov (1954), Basic laws of turbulent mixing in the ground layer of the atmosphere., Tr. Inst. Teor. Geofiz. Akad. Nauk SSSR, 1, 95–115.

Montesi, J., K. Elder, R. A. Schmidt, and R. E. Davis (2004), Sublimation of intercepted snow within a subalpine forest canopy at two elevations, Journal of Hydrometeorology, 5, 763–773.

Nakai, Y. T., T. Sakamoto, T. Terajima, H. Kiahara, and T. Saito (1994), Snow interception by forest canopies: Weighing a conifer tree, meteorological observation and analysis by hte Penman-Monteith formula, IAHS Publ., 223, 227–236.

Niu, G. Y., and Z. L. Yang (2004), Effects of vegetation canopy processes on snow surface energy and mass balances, Journal Of Geophysical Research-Atmospheres, 109.

Parviainen, J., and J. W. Pomeroy (2000), Multiple-scale modelling of forest snow sublimation: initial findings, Hydrological Processes, 14, 2669–2681.

Pomeroy, J. W., and R. L. H. Essery (1999), Turbulent fluxes during blowing snow: field tests of model sublimation predictions, Hydrological Processes, 13, 2963–2975.

Pomeroy, J. W., and D. M. Gray (1995), Snowcover accumulation, relocation and management, NHRI Science Report 7, National Hydrology Research Institute, Environment Canada, 134 pp.

Pomeroy, J. W., N. Parviainen, N. Hedstrom, and D. M. Gray (1998), Coupled modeling of forest snow interception and sublimation., Hydrological Processes, 12, 2317–2337.

Pomeroy, J. W., and R. A. Schmidt (1993), The use of fractal geometry in modeling intercepted snow accumulation and sublimation., Proceedings of the 50th Eastern Snow Conference., Quebec City, QC, Canada., 50, 1–10.

Schmidt, R. A. (1991), Sublimation of snow intercepted by an artificial conifer, Agricultural and Forest Meteorology, 54, 1–27.

Schmidt, R. A., R. L. Jairell, and J. W. Pomeroy (1988), Measuring snow interception and loss from an artificial conifer., Proceedings of the 56th Western Snow Conference., Kalispell, MT., 166–169.

Schmidt, R. A., and C. A. Troendle (1992), Sublimation of intercepted snow as a global source of water vapor., Proceedings of the 60th Western Snow Conference., Jackson, WY., 60, 1–9.

Schmidt, R. A., C. A. Troendle, and J. R. Meiman (1998), Sublimation of snowpacks in subalpine conifer forests, Canadian Journal Of Forest Research-Revue Canadienne De Recherche Forestiere, 28, 501–513.

Schuepp, P. H., M. Y. Leclerc, J. I. Macpherson, and R. L. Desjardins (1990), Footprint Prediction Of Scalar Fluxes From Analytical Solutions Of The Diffusion Equation, Boundary-Layer Meteorology, 50, 353–373.

Sicart, J. E., J. W. Pomeroy, R. E. Essery, J. Hardy, T. Link, and D. Marks (2004), A sensitivity study of daytime net radiation during snowmelt to forest canopy and atmospheric conditions, Journal of Hydrometeorology, 5, 774–784.

Turnipseed, A. A., P. D. Blanken, D. E. Anderson, and R. K. Monson (2002), Energy budget above a high-elevation subalpine forest in complex topography, Agricultural And Forest Meteorology, 110, 177–201.

West, A. (1962), Snow evaporation from a forested watershed in the central Sierra Nevada, Journal of Forestry, 60, 481–484.

Woo, M. K., and M. A. Giesbrecht (2000), Simulation of snowmelt in a subarctic spruce woodland: Scale considerations, Nordic Hydrology, 31, 301–316.