

# Estimating sublimation of intercepted and sub-canopy snow using eddy covariance systems

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## Abstract:

Direct measurements of winter water loss due to sublimation were made in a sub-alpine forest in the Rocky Mountains of Colorado. Above- and below-canopy eddy covariance systems indicated substantial losses of winter-season snow accumulation in the form of snowpack ( $0.41 \text{ mm d}^{-1}$ ) and intercepted snow ( $0.71 \text{ mm d}^{-1}$ ) 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 above-canopy 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 above-canopy latent heat fluxes, high within-canopy sublimation rates (maximum =  $3.7 \text{ mm d}^{-1}$ ), and diminished sub-canopy snowpack sublimation. These results indicate that sublimation losses from the sub-canopy 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. Copyright © 2007 John Wiley & Sons, Ltd.

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## INTRODUCTION

Sublimation of intercepted snow constitutes a significant component of the overall water balance in many seasonally snow-covered coniferous forests (Schmidt and Troendle, 1992; Lundberg and Halldin, 1994; Pomeroy and Gray, 1995; Essery *et al.*, 2003). In such environments, approximately 60% of total annual snowfall can be intercepted by the canopy (Hedstrom and Pomeroy, 1998) and associated sublimation losses may exceed 30% (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).

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 vapour 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 (Pomeroy and Essery, 1999; Pomeroy and Li, 2000; Fassnacht, 2004). Similarly, much work has been devoted towards estimating sublimation losses from intercepted snow (Schmidt and Troendle, 1992; Pomeroy and Schmidt, 1993; Montesi, *et al.*, 2004). Harding and Pomeroy (1996) present some of the first observations of turbulent fluxes in snow-covered forests. Similarly, differences in energy fluxes between snow-covered and snow-free canopies have been documented (Nakai *et al.*, 1999). These studies complement several works on snow-vegetation interactions in high-latitude boreal forests (Blanken *et al.*, 1997; Davis *et al.*, 1997; Hardy *et al.*, 1997; Blanken *et al.*, 1998; Hedstrom and Pomeroy, 1998; Pomeroy *et al.*, 1999; Link and Marks, 1999a,b; Blanken and Black, 2004). Lacking is a thorough analysis of the above- and below-canopy energy

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fluxes and associated differences in sublimation at an individual site—particularly, at mid-latitude.

Measurement of sublimation from intercepted snow has primarily focused on tree-weighing techniques (Schmidt *et al.*, 1988; Schmidt, 1991; Nakai *et al.*, 1994; Montesi *et al.*, 2004). Several factors lead to uncertainty in this approach and towards 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 and Schmidt, 1993; Pomeroy *et al.*, 1998). All of these limitations could be accounted for in techniques that integrate all of these processes by measuring above- and below-canopy water vapour 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.

Direct measurements of winter water loss by sublimation of snow from a sub-alpine 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 largely 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: (1) determine snow sublimation rates in a sub-alpine forest; (2) partition snow sublimation into above- and below canopy components; and (3) explore relationships between atmospheric and snowpack conditions, and snow sublimation rates.

STUDY SITE

This work was conducted at the Niwot Ridge Forest, Colorado Ameriflux site (40°1'58"N; 105°32'47"W), located at an elevation of 3050 m approximately 8 km east of the continental divide (Figure 1). The area 1 km<sup>2</sup> east of the tower is dominated by Engelmann spruce (7 trees ha<sup>-1</sup>) and lodgepole pine (27 trees ha<sup>-1</sup>). Rising at a slope of about 6–7°, the 1 km<sup>2</sup> area west of the tower contains sub-alpine fir (16 trees ha<sup>-1</sup>), Engelman Spruce (10 trees ha<sup>-1</sup>) and lodgepole pine (9 trees ha<sup>-1</sup>). Maximum leaf area index during the growing season is approximately 4.2 m<sup>2</sup> m<sup>-2</sup>. 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 twentieth century. The hydrology

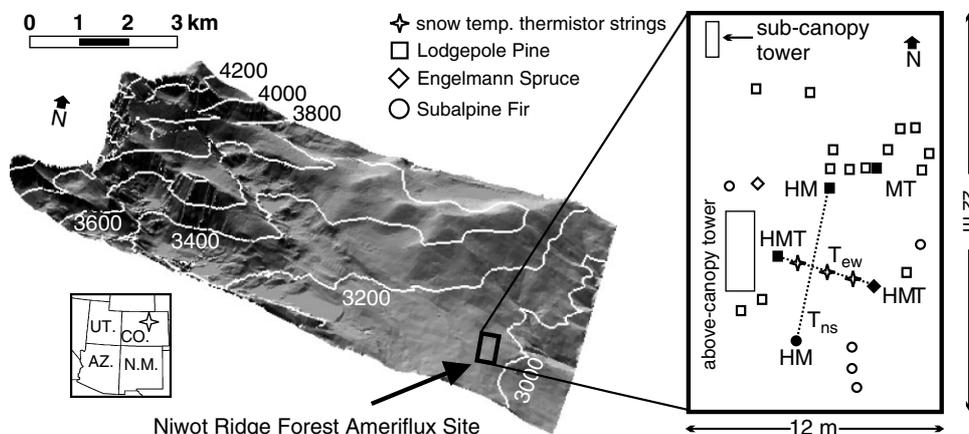


Figure 1. Hillshade relief of the Niwot Ridge, Long Term Ecological Research site, Colorado. The terrain is illuminated from the northwest and is overlain with a 200-m contour interval. The Niwot Ridge, Forest Ameriflux site is shown at lower right of image. Location of above- and below-canopy flux towers and water content reflectometers (M), thermistor strings (stars), ground thermistors (T), and soil heat flux plates (H) are shown. Additional H, M and T measurements were made along east–west ( $T_{ew}$ ) and north–south ( $T_{ns}$ ) transects

of the site is dominated by a moderate snowpack that accounts 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 speed 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).

## METHODS

### Flux measurements

Water vapour fluxes, (latent heat flux;  $\lambda E$ ) were calculated as 30-min means of 10-Hz measurements over a 40 d mid-winter period (day of year (DOY) 60–100, 2002) using the eddy covariance (EC) method (Goulden *et al.*, 1996; Turnipseed *et al.*, 2002):

$$\lambda E = L_v \overline{w' \rho_v'} \quad (1)$$

where  $L_v$  is the latent heat of sublimation,  $w'$  is the deviation of vertical wind velocity ( $\text{m s}^{-1}$ ) from the 1/2-h mean,  $\rho_v'$  is the deviation of the water vapour density from the 1/2-h mean. The above- and below-canopy EC systems were mounted at heights of 21.5 and 1.7 m above-ground, respectively, from towers separated by a distance of approximately 20-m. The above- and below-canopy EC systems and other meteorological instruments are summarized in Table I. Post-processing corrections to the EC data included mathematical coordinate rotation of the mean lateral and vertical wind velocities to zero; only the lateral component was corrected for in processing the sub-canopy data (Baldocchi and Hutchison, 1987). The sonic anemometers' virtual air temperature were corrected, accounting for wind speed normal to the sonic path and humidity effects (Schotanus *et al.*, 1983). See Turnipseed *et al.*, (2002, 2003) for complete details.

Components of snow sublimation were computed as:

$$\lambda E_{c,t} = \lambda E_{c,s} + \lambda E_{c,i} \quad (2)$$

where  $\lambda E_{c,t}$  is the total sublimation from the system measured using the above-canopy 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 vapour fluxes associated with sublimation of intercepted snow,  $\lambda E_{c,i}$  were determined as the difference of measured above- and below-canopy fluxes. In this regard, we assumed that there was no change of vapour storage in the canopy air space. Measurements of the above-canopy  $\text{CO}_2$  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 above-canopy 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 ( $z$ ):

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

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

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 soil ( $G$ ) heat flux was developed (Blanken *et al.*, 1997; Blanken *et al.*, 1998). The relationship between the 30-min above canopy ( $\lambda E + H$ ) and ( $R_n - G$ )

Table I. Observations and instruments on the above- and below-canopy towers at the Niwot Ridge Forest, Ameriflux site

Observation	Measurement height, meters	Instrument
Relative humidity, %	21.5	HMP-35D, Vaisala, Inc.
Air temperature, °C	21.5 1.7	CSAT-3, Campbell Scientific
Pressure, kpa	18	PT101B, Vaisala, Inc.
Net radiation, $\text{W m}^{-2}$	26	4-component CNR-1, Kipp & Zonen
$\text{H}_2\text{O}$ flux, $\text{mg m}^{-2} \text{ s}^{-1}$	21.5 1.7	IRGA-6260, Li-Cor
$\text{CO}_2$ flux, $\text{mg m}^{-2} \text{ s}^{-1}$	21.5	IRGA-6260, Li-Cor
Wind speed, $\text{m s}^{-1}$	21.5 1.7	Propvane-09 101, RM Young Inc.
Wind direction, degrees	21.5 1.7	Propvane-09 101, RM Young Inc.
Precipitation, mm	12	385-L, Met One
Soil heat flux, $\text{W m}^{-2}$	-0.07-0.1	HFT-1, REBS
Soil moisture, % by volume	0-0.15	CS-615, Campbell Scientific
Soil temperature, °C	0-0.1	STP-1, REBS

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

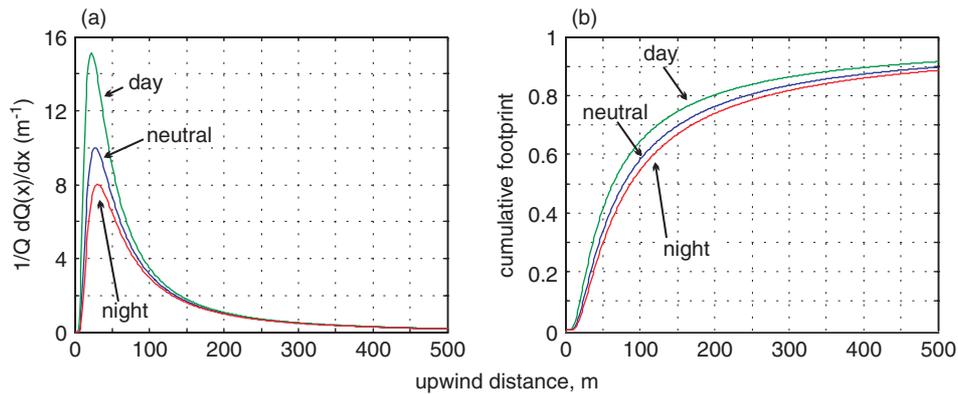


Figure 2. (a) Normalized change in the sub-canopy turbulent flux ( $Q$ ) with upwind distance ( $x$ ) during typical daytime, neutral and night time atmospheric stability conditions. (b) Cumulative change in the sub-canopy turbulent flux ( $Q$ ) with upwind distance ( $x$ ) for typical daytime, neutral and night time atmospheric stability conditions

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.* (1990). The upwind distance that the sub-canopy flux measurements were most sensitive to occurred at a distance of 23, 27, and 29-m during typical daytime, neutral, and night time atmospheric stability conditions, respectively (Figure 2(a)). 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 night time atmospheric stability conditions, respectively) (Figure 2(b)). For the above-canopy turbulent flux measurements, greater than 90% originated from within 1200 m of the tower during downslope, westerly flow, common during the winter ( $\sim 70\%$  of the time) (Turnipseed *et al.*, 2003).

#### Supporting sub-canopy 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 vapour pressure gradients in the snowpack and therefore the movement of water vapour from deeper in the snowpack towards the snowpack-atmosphere interface (McClung and Schaerer, 1993). The three thermistor strings were placed along a transect through a small clearing ( $\sim 6$  m in diameter) in the forest adjacent to the sub-canopy flux tower (Figure 1). 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.

Eight water content reflectometers (Campbell Scientific model CS-615) were used to monitor soil moisture conditions surrounding the towers (Figure 1). 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.

## RESULTS

Soil temperature, moisture and soil heat flux were consistent with mid-winter conditions throughout the study period (Figure 3(a–c)). Soil temperatures were within 0 to  $-2^\circ\text{C}$  during the winter period, indicating sufficient insulation of the soil from cold winter air temperatures (Figure 3(a)). Spring onset of snowmelt percolation occurred on DOY 100; soil moisture increased by 3-fold over the subsequent 20-d period (Figure 3(b)). Throughout the majority of the study period (i.e. DOY 60–100),

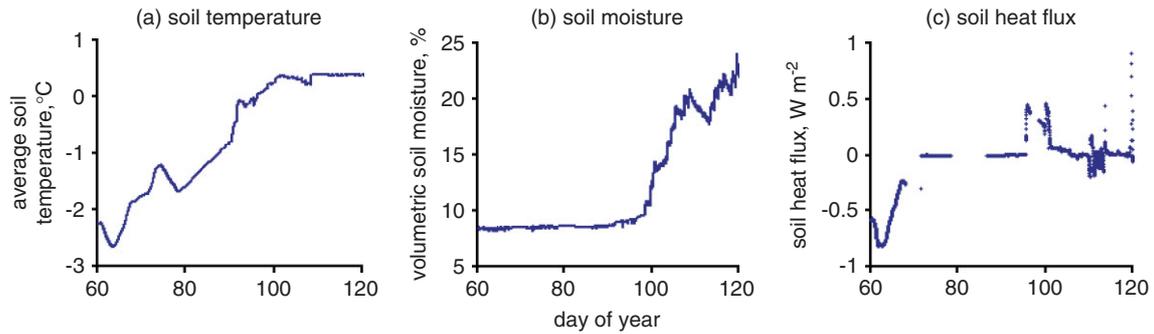


Figure 3. Time series of soil temperature (a), moisture (b), and soil heat flux (c) from day of year 60–120, 2002

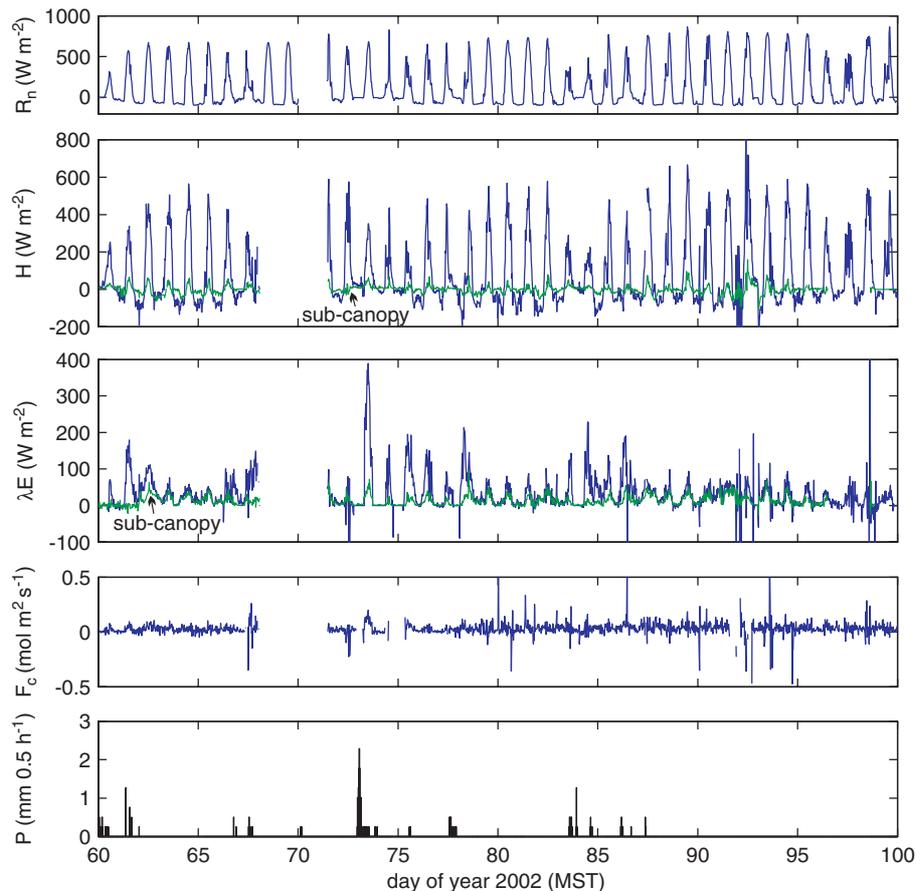


Figure 4. Diurnal variability in measured sensible ( $H$ ) and latent ( $\lambda E$ ) heat fluxes (above- and sub-canopy), net radiation ( $R_n$ ), carbon flux ( $F_c$ ), and precipitation ( $P$ ). Period shown is from DOY 60–100, 2002. Positive values represent fluxes towards the surface

soil heat flux was close to  $0 \text{ W m}^{-2}$  (Figure 3(c)) and had a negligible impact on available energy and energy input to the snowpack. Temporal variability in soil temperature (coefficient of variation = 0.77) and soil heat flux (coefficient of variation = 2.6) was considerably greater than that of soil moisture (coefficient of variation = 0.11).

The diurnal energy fluxes,  $R_n$ ,  $\lambda E$  and  $H$  above and below the canopy are shown in Figure 4, together with precipitation. The  $\text{CO}_2$  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  $\lambda E$  beneath the canopy; above-canopy

ratios of the daytime mean  $H/R_n$  and  $\lambda E/R_n$  were 0.67 and 0.16, respectively. Beneath the canopy, these ratios were 0.02 ( $H/R_n$ ) and 0.06 ( $\lambda E/R_n$ ). Although the  $\lambda E/R_n$  fraction was on average relatively small, large increases in  $\lambda E$  with a subsequent decrease in  $H$  occurred several times in response to snowfall events.

Average sublimation rates over the study period were  $0.70$  and  $0.41 \text{ mm d}^{-1}$  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 5(a)). The ratio between sub-canopy snowpack sublimation and total sublimation averaged

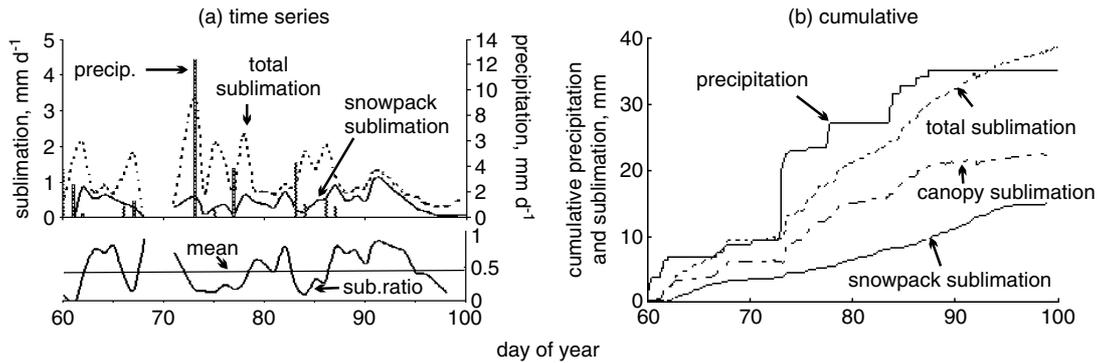


Figure 5. (a) Time series of daily average sublimation measured above the canopy and beneath the canopy. Precipitation and the ratio of snowpack sublimation to total sublimation are also shown. (b) Cumulative sublimation from the snowpack, intercepted snow (canopy), total sublimation and precipitation throughout the study period

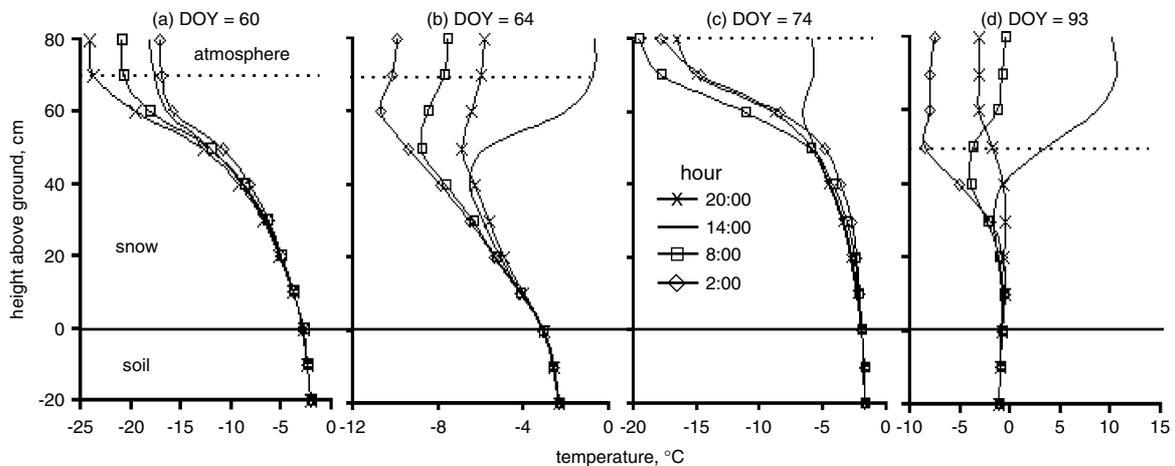


Figure 6. Snowpack, soil, and air temperature profiles from 80 cm above the ground surface to 20 cm below. Profiles are shown for 4 different days for hours 2, 8, 14, and 20 Mountain Standard Time (-7 UTC). Dotted horizontal lines represent the snowpack-atmosphere interface. Solid horizontal lines indicate the soil-snowpack interface

0.45 during the study period, increasing with time after snowfall and approaching one during consecutive days without snowfall; e.g. DOY 63–65 and DOY 87–93 (Figure 5(a)). On average snowpack to total sublimation ratios peaked 3 days after snowfall; timing to peak varied considerably with snowfall magnitude.

A total of 34.8 mm of snow fell during the measurement period (Figure 5(b)). 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 sub-canopy snow that fell prior to the start of the measurements. Subtracting the above-canopy  $\lambda E$  measurements from that below the canopy (Figure 5(b)) 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 6(a,b)). Similarly, diurnal signatures in snow temperature were considerably different on DOY 74 versus DOY 93, with little variability in morning, evening and night-time snow temperatures on DOY 74 versus a sinusoidal diurnal snow temperature signature on DOY 93 (Figure 6(c,d)); sublimation rates were 0.1 versus 0.6 mm d<sup>-1</sup> for these two days, respectively. Temperature fluctuations in the surface layers, associated with cool nights and warm dry days potentially drive significant water vapour movement in the surface layers of the snowpack and enhance sublimation rates.

Estimates of snow depth on snow temperature profile plots (Figure 6(a–d)) 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. Therefore, we assumed snow depth was equal to that measured in the snowpit on DOY 64 as no precipitation or melt occurred between DOY 60 and 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 (Figure 5(a)). Thus, we

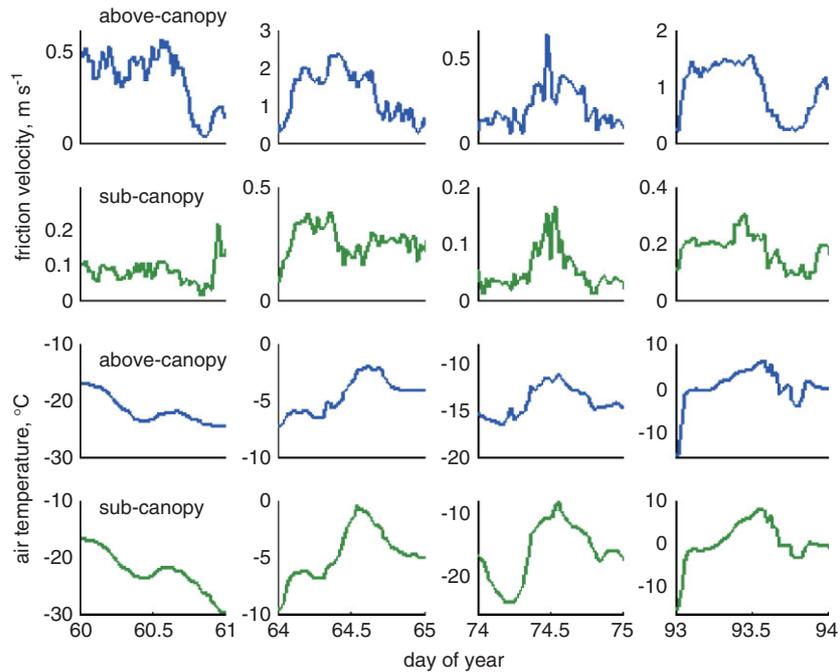


Figure 7. Above- and sub-canopy diurnal variability in friction velocity and air temperature for the same 4 days shown in Figure 6(a–d)

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 (Figure 6(d)).

Above- and below-canopy friction velocities were considerably greater for DOY 64 and 93 relative to that on DOY 60 and 74 (Figure 7). Chinook winds, which enhance latent energy exchange between the land-surface and the atmosphere (Golding, 1978), were persistent on DOY 64 and 93. On these days, down-slope, westerly winds prevailed in combination with dry atmospheric conditions; above-canopy relative humidity averaged 29 and 38%, respectively. Over the 48 h prior to DOY 64 and 93 above-canopy air temperatures rose by approximately 20 °C; consistent with Chinook conditions (Barry, 1992). The combination of the relatively high air temperatures, dry conditions, and sufficient turbulence lead to enhanced near-surface gradients in specific humidity and sublimation. Conversely, on DOY 60 and 74, calm (<3 m s<sup>-1</sup>) easterly winds prevailed and relative humidity averaged 80 and 85%, respectively.

Unstable atmospheric conditions resulted in considerable sublimation of intercepted snow. For example, on DOY 62 measurement-height/Monin-Obukhov-length ratios dropped below -200 (Figure 8) and daily sublimation was 2.09 mm (Figure 5(a)). Slight instabilities resulted in significant sublimation rates after snow-fall events; measurement-height/Monin-Obukhov-length ratios on DOY 76 were between -3 and 0 yet sublimation of intercepted snow was 1.74 mm, only 17% lower than that of DOY 62. Precipitation magnitude is likely responsible for these differences as 12 and 6 mm of precipitation fell on DOY 73 and 60, respectively.

## DISCUSSION

A variety of techniques have been developed to estimate sublimation from snowpacks and intercepted snow (Pomeroy *et al.*, 1998; Montesi *et al.*, 2004). It is especially challenging to capture the impact of vegetation on variability in turbulence and subsequent vapour 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 homogeneous 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<sup>-1</sup>) were low relative to the highest of values found within the literature observed on the Canadian Prairies (1.2–1.8 mm d<sup>-1</sup> (Pomeroy and Essery, 1999)) and those observed in open mountainous locations (0.75 mm d<sup>-1</sup> (Fassnacht, 2004)); open sites are known to exhibit substantially greater sublimation rates (West, 1962).

Our sublimation estimates were within 14% of values observed at the nearby Fraser Experimental Forest (e.g. 0.36 mm d<sup>-1</sup>) (Schmidt *et al.*, 1998). The weekly snowpits excavated in a clearing adjacent to the flux towers used in this research indicated a total sublimation rate of

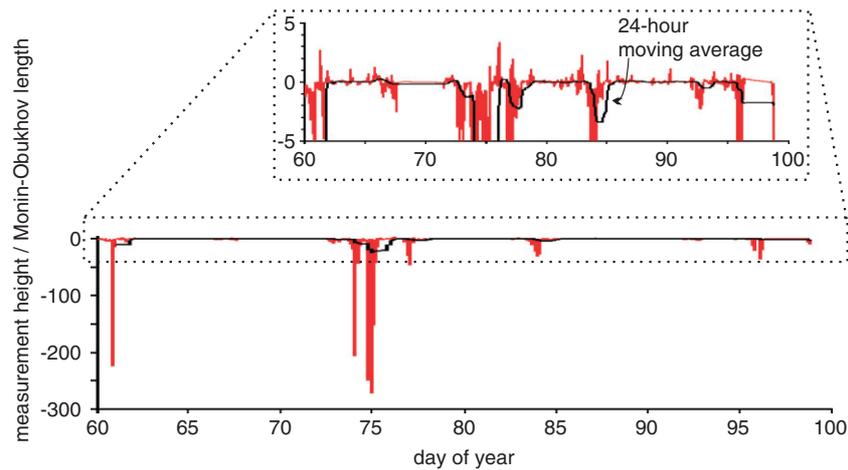


Figure 8. Measurement-height/Monin-Obukhov-length ratios calculated from 30-min, above-canopy observations. See Equation (3) for derivation of Monin-Obukhov length. Negative and positive values represent unstable and stable conditions, respectively

$0.8 \text{ mm d}^{-1}$ . 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 5(a)) 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 ( $0.71 \text{ mm d}^{-1}$ ) compared favourably with previous works. For example, Parviainen and Pomeroy (2000) estimated intercepted snow sublimation from a boreal forest at  $0.5 \text{ mm d}^{-1}$ ; 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 estimates of intercepted snow sublimation were considerably lower than 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 *et al.* (2004). Climatic differences may also be responsible for these discrepancies as the sites used by Montesi *et al.* (2004) were located on the windward side of the continental divide; our site is located on the leeward side of the divide where warm, dry chinook winds are more prevalent (Barry, 1992).

As previously found by Niu and Yang (2004), the relatively high above-canopy sensible heat fluxes lead to strong temperature differences between vegetation and the snow-surface, driving strong specific humidity gradients at the snow-atmosphere interface due to elevated snowpack sublimation rates (e.g. DOY 78–82 & 88–95, Figure 5(a)). When snowfall occurred, above-canopy available energy was partitioned into latent heat fluxes (e.g. Figure 4, DOY 74), leading to relatively low total-sublimation to snowpack-sublimation ratios (DOY 74, Figure 5(a)). This shift in partitioning of available energy is consistent with that found by Nakai *et al.* (1999). The results presented here explicitly indicate that sublimation losses from the sub-canopy snowpack are strongly dependent on the partitioning of sensible and latent heat fluxes in the canopy.

## CONCLUSIONS

Sub-canopy and above-canopy eddy covariance systems indicated substantial losses of winter-season snow accumulation in the form of snowpack ( $0.41 \text{ mm d}^{-1}$ ) and intercepted snow ( $0.71 \text{ mm d}^{-1}$ ) 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 above-canopy 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 above-canopy latent heat fluxes, high within-canopy sublimation rates, and diminished sub-canopy snowpack sublimation. These results indicate that sublimation losses from the sub-canopy 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 above- and below-canopy processes.

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## REFERENCES

- Baldocchi DD, Hutchison BA. 1987. Turbulence in an almond orchard-vertical variations in turbulent statistics. *Boundary-Layer Meteorology* **40**: 127–146.
- 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:10.1029/2005WR004387.
- Barry RG. 1992. *Mountain Weather and Climate*. Routledge: New York; 402.
- Blanken PD, Black TA. 2004. The canopy conductance of a boreal aspen forest, Prince Albert National Park, Canada. *Hydrological Processes* **18**: 1561–1578.
- Blanken PD, Black TA, Yang PC, Neumann HH, Nesic Z, Staebler R, den Hartog G, Novak MD, Lee X. 1997. Energy balance and canopy conductance of a boreal aspen forest: Partitioning overstory and understory components. *Journal of Geophysical Research-Atmospheres* **102**: 28915–28927.
- Blanken PD, Black TA, Neumann HH, Den Hartog G, Yang PC, Nesic Z, Staebler R, Chen W, Novak MD. 1998. Turbulent flux measurements above and below the overstory of a boreal aspen forest. *Boundary-Layer Meteorology* **89**: 109–140.
- Caine N. 1995. Temporal trends in the quality of streamwater in an Alpine environment: green Lakes valley, Colorado Front Range, USA. *Geografiska Annaler Series A-Physical Geography* **77A**: 207–220.
- 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.
- Essery R, Pomeroy J, Parviainen J, Storck P. 2003. Sublimation of snow from coniferous forests in a climate model. *Journal of Climate* **16**: 1855–1864.
- Fassnacht SR. 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.
- Golding DL. 1978. Calculated snowpack evaporation during Chinooks along eastern slopes of rocky Mountains In Alberta. *Journal of Applied Meteorology* **17**: 1647–1651.
- Goulden ML, Munger JW, Fan SM, Daube BC, Wofsy SC. 1996. Measurements of carbon sequestration by long-term eddy covariance: methods and a critical evaluation of accuracy. *Global Change Biology* **2**: 169–182.
- Harding RJ, Pomeroy JW. 1996. Energy balance of the winter boreal landscape. *Journal of Climate* **9**: 2778–2787.
- 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.
- Hedstrom NR, Pomeroy JW. 1998. Measurements and modelling of snow interception in the boreal forest. *Hydrological Processes* **12**: 1611–1625.
- Link TE, Marks D. 1999a. Distributed simulation of snowcover mass- and energy-balance in the boreal forest. *Hydrological Processes* **13**: 2439–2452.
- Link TE, Marks D. 1999b. Point simulation of seasonal snow cover dynamics beneath boreal forest canopies. *Journal of Geophysical Research-Atmospheres* **104**: 27841–27857.
- Lundberg A, Halldin S. 1994. Evaporation of intercepted snow: Analysis of governing factors. *Water Resources Research* **30**: 2587–2598.
- McClung D, Schaerer P. 1993. *The Avalanche Handbook*. The Mountaineers: Seattle, Washington, DC; 271.
- Monin AS, Obukhov AM. 1954. Basic laws of turbulent mixing in the ground layer of the atmosphere. *Akad. Nauk SSSR Geofiz. Inst.* **151**(24): 163–187.
- Montesi J, Elder K, Schmidt RA, Davis RE. 2004. Sublimation of intercepted snow within a subalpine forest canopy at two elevations. *Journal of Hydrometeorology* **5**: 763–773.
- Nakai YT, Sakamoto T, Terajima T, Kiahara H, Saito T. 1994. Snow interception by forest canopies: Weighing a conifer tree, meteorological observation and analysis by the Penman-Monteith formula. *IAHS Publications* **223**: 227–236.
- Nakai Y, Sakamoto T, Terajima T, Kitamura K, Shirai T. 1999. Energy balance above a boreal coniferous forest: a difference in turbulent fluxes between snow-covered and snow-free canopies. *Hydrological Processes* **13**: 515–529.
- Niu GY, Yang ZL. 2004. Effects of vegetation canopy processes on snow surface energy and mass balances. *Journal of Geophysical Research-Atmospheres* **109**: D23111. DOI: 10.1029/2004JD004884.
- Parviainen J, Pomeroy JW. 2000. Multiple-scale modelling of forest snow sublimation: initial findings. *Hydrological Processes* **14**: 2669–2681.
- Pomeroy JW, Schmidt RA. 1993. The use of fractal geometry in modeling intercepted snow accumulation and sublimation. In Proceedings of the 50th Eastern Snow Conference, Quebec, Canada, 50, 1–10.
- Pomeroy JW, Gray DM. 1995. Snowcover accumulation, relocation and management, *NHRI Science Report 7, National Hydrology Research Institute, Environment Canada*, 134.
- Pomeroy JW, Essery RLH. 1999. Turbulent fluxes during blowing snow: field tests of model sublimation predictions. *Hydrological Processes* **13**: 2963–2975.
- Pomeroy JW, Li L. 2000. Prairie and arctic areal snow cover mass balance using a blowing snow model. *Journal of Geophysical Research-Atmospheres* **105**: 26619–26634.
- Pomeroy JW, Parviainen N, Hedstrom N, Gray DM. 1998. Coupled modeling of forest snow interception and sublimation. *Hydrological Processes* **12**: 2317–2337.
- Pomeroy JW, Davies TD, Jones HG, Marsh P, Peters NE, Tranter M. 1999. Transformations of snow chemistry in the boreal forest: accumulation and volatilization. *Hydrological Processes* **13**: 2257–+.
- Schmidt RA. 1991. Sublimation of snow intercepted by an artificial conifer. *Agricultural and Forest Meteorology* **54**: 1–27.
- Schmidt RA, Troendle CA. 1992. Sublimation of intercepted snow as a global source of water vapor. In Proceedings of the 60th Western Snow Conference, Jackson, WY, 60, 1–9.
- Schmidt RA, Jairell RL, Pomeroy JW. 1988. Measuring snow interception and loss from an artificial conifer. In Proceedings of the 56th Western Snow Conference, Kalispell, MT, 166–169.
- Schmidt RA, Troendle CA, Meiman JR. 1998. Sublimation of snowpacks in subalpine conifer forests. *Canadian Journal of Forest Research-Revue Canadienne De Recherche Forestiere* **28**: 501–513.
- 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.
- Schuepp PH, Leclerc MY, Macpherson JJ, Desjardins RL. 1990. Foot-print prediction of scalar fluxes from analytical solutions of the diffusion equation. *Boundary-Layer Meteorology* **50**: 353–373.
- 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.
- 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.
- Turnipseed AA, Anderson DE, Blanken PD, Baugh WM, Monson RK. 2003. Airflows and turbulent flux measurements in mountainous terrain Part 1. Canopy and local effects. *Agricultural and Forest Meteorology* **119**: 1–21.
- West A. 1962. Snow evaporation from a forested watershed in the central Sierra Nevada. *Journal of Forestry* **60**: 481–484.
- Woo MK, Giesbrecht MA. 2000. Simulation of snowmelt in a subarctic spruce woodland: Scale considerations. *Nordic Hydrology* **31**: 301–316.