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Sublimation from a seasonal snowpack at a continental, mid-latitude alpine site

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Abstract:

Sublimation from the seasonal snowpack was calculated using the aerodynamic profile method at Niwot Ridge in the Colorado Front Range. Past studies of sublimation from snow have been inconclusive in determining both the rate and timing of the transfer of water between the snowpack and the atmosphere, primarily because they relied on one-dimensional measurements of turbulent fluxes or short term data sets. We calculated latent heat fluxes at ten minute intervals based on measurements of temperature, relative humidity and wind speed at heights of 0·5 m, 1·0 m and 2·0 m above the snowpack for nine months during the 1994–1995 snow season. The meteorological instruments were raised or lowered daily to maintain a constant height above the snow surface. At each ten minute time step, the latent heat fluxes were converted directly into millimeters of sublimation or condensation. Total net sublimation for the snow season was 195 mm of water equivalent, or 15% of maximum snow accumulation at the study site. The majority of this sublimation occurred during the snow accumulation season. Monthly losses to sublimation during the fall and winter ranged from 27 to 54 mm of water equivalent. The snowmelt season from May through mid-July showed net condensation to the snowpack ranging from 5 to 16 mm of water equivalent. Sublimation was sometimes episodic in nature, but often showed a diurnal periodicity with higher rates of sublimation during the day. Copyright © 1999 John Wiley & Sons, Ltd.

KEY WORDS seasonal snow pack; sublimation; moisture transfers

INTRODUCTION

Melting snow in mountainous regions is the source of much of the annual streamflow in the western United States. Consequently, a clear understanding of the water balance of high altitude, seasonally snow-covered basins is of particular importance both from the point of view of water resources and in terms of evaluating the hydrology and hydrochemistry of high elevation catchments (Kattelmann and Elder, 1991). One component of the alpine water balance which is still poorly understood is the amount of water exchanged between seasonal snowpacks and the atmosphere through sublimation and condensation (Lang, 1981). Moisture fluxes between the snowpack and the atmosphere are commonly calculated from measurements of the snowpack latent heat flux. However, because of the difficulties associated with measuring turbulent heat fluxes in alpine environments, detailed observations of snowpack latent heat fluxes and subsequent estimations of water fluxes are limited in these areas.

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In general, the literature reveals a wide range of opinions regarding the importance of sublimation and condensation in the water balance of high-elevation catchments. Martinelli (1960) reports that sublimation is responsible for only 1–2% of summer snowpack ablation in snowfields between 3500–3800 m in the Rocky Mountains. Similarly, measurements from snow evaporation pans at the Central Sierra Snow Laboratory in California measured annual sublimation at only 25–35 mm or 2–3% of total ablation over the snow season (West and Knoerr, 1959). More recent work in this area has focused on the snowpack energy balance over short time periods during the snowmelt season. Typically, the latent heat flux is only a small fraction of the snowpack energy balance during this time period (e.g. Harding, 1986; Kuusisto, 1986). In the Swiss Alps, Pluss and Mazzoni (1994) report daily snowpack latent heat fluxes in May to be on the order of 4 W m^{-2} which equates to only 3·7 mm of net monthly sublimation. In Norway, Harding (1986) measured net condensation to the snowpack during the month of May.

In contrast, there have been several studies that describe significant mass loss through sublimation in alpine snowpacks. In an extreme example, Beaty (1975) reported that sublimation was responsible for 80% of the ablation of fresh snow and 60% of the ablation of older snow during springtime conditions in the White Mountains of California. Measurements made using snow evaporation pans indicate that net sublimation for the five month winter period from December–April was 135 mm of water equivalent at Pingree Park (2740 m) in the Colorado Front Range (Meiman and Grant, 1974). Kattelmann and Elder (1991) estimated sublimation from snow to be 18% of total precipitation over two water years based on data collected in the Emerald Lake Basin in the Sierra Nevada. At Niwot Ridge in the Colorado Front Range, Berg (1986) has previously estimated sublimation losses from blowing snow to be between 30–51% of precipitation for the two year period 1973–1975.

The most common technique for measuring the snowpack latent heat flux is the bulk aerodynamic or mean profile method (Moore, 1983). This method has the advantage of requiring meteorological measurements at only one height above the snowpack. However, a primary assumption of the bulk profile method, that snow surface temperature effectively tracks the air temperature, is often inaccurate below 0 °C and will result in the over-estimation of sublimation (Bernier and Edwards, 1989). A more accurate method for calculating the snowpack latent heat flux is the aerodynamic profile method (APM) which requires the measurement of wind speed, temperature, and relative humidity at multiple heights above the snowpack (Cline, 1997b). In a comparison of the bulk aerodynamic method and the APM, Pluss and Mazzoni (1994) report that the logistically simpler bulk method is not a satisfactory substitute for the APM in terms of calculating latent heat fluxes. The third, and most accurate, method for measuring latent heat fluxes is the eddy correlation method. However, this procedure requires a high frequency sonic anemometer that is too fragile to use for extended periods in an alpine environment (Hood and Williams, 1998; Male and Granger, 1981).

In this study, we evaluate the importance of moisture fluxes to and from a seasonal snowpack for the entire snow season using the aerodynamic profile method. Specific hypotheses we evaluate include (1) water loss by sublimation and evaporation from the snowpack is more important than water gain from condensation, (2) water losses from a seasonal snowpack on the time scale of a snow season are an important component of the annual hydrologic cycle of alpine areas located in continental sites, (3) water losses from the seasonal snowpack are driven by high-magnitude, short-duration events, and (4) wind speeds are the primary meteorological variable driving water fluxes between the atmosphere and the snow surface.

SITE DESCRIPTION

The data for this study were collected during the 1994–1995 snow season at Niwot Ridge, Colorado (Figure 1). Niwot Ridge is a broad interfluve located on the eastern slope of the Front Range of the Rocky Mountains (3517 m a.s.l., 40°03′N, 105°35′W). The Front Range of Colorado is characterized by a dry, continental climate. The mean annual temperature range is 21 °C while daily temperatures show a range of 6–8 °C (Barry, 1992). The high elevation and exposure of Niwot Ridge result in high incoming solar radiation, low air temperatures and vapor pressures, and high wind velocities. Wind speed during the snow

Figure 1. Aerial photo showing the Niwot Ridge saddle study site in the Colorado Front Range. Inset shows location of meteorological tower, index snowpit and Belfort precipitation gage

season averages 10–13 m/s (Greenland, 1989). Westerly flow dominates during the winter months with over 75% of recorded wind directions between 255 and 300 degrees. Additionally, the unobstructed upwind fetch in this direction is approximately 2·5 km (Berg, 1986).

Meteorological instrumentation for this study was located in a broad, flat saddle on Niwot Ridge approximately 150 m above treeline (Figure 1). The surface of the ridge is made up of alpine tundra vegetation communities interspersed with patchy snowfields, many of which last into the summer months. Snow accumulation is governed chiefly by an interaction between persistent westerly winds and topography with most of the deposition occurring in leeward depressions and sheltered areas. Snow accumulation typically begins in September with consistent snowcover at the study site in October. In the saddle area, snow depths generally exceed 2 m at the end of winter (Cline, 1997a).

METHODS

Detailed information on meteorological instrumentation and error analysis have been presented previously (Cline, 1997b). Wind speed is measured at three levels above the surface using R.M. Young 05103 anemometers. Relative humidity and temperature are similarly measured at three levels using Vaisala HMP35C temperature/relative humidity probes. All of the instrumentation is mounted on a movable support which was attached to a 10 m fixed mast (Figure 2). The support was adjusted regularly to ensure consistent instrument heights of 0·5 m, 1·0 m and 2·0 m above the snow surface. Additionally, a REBS Q7 net

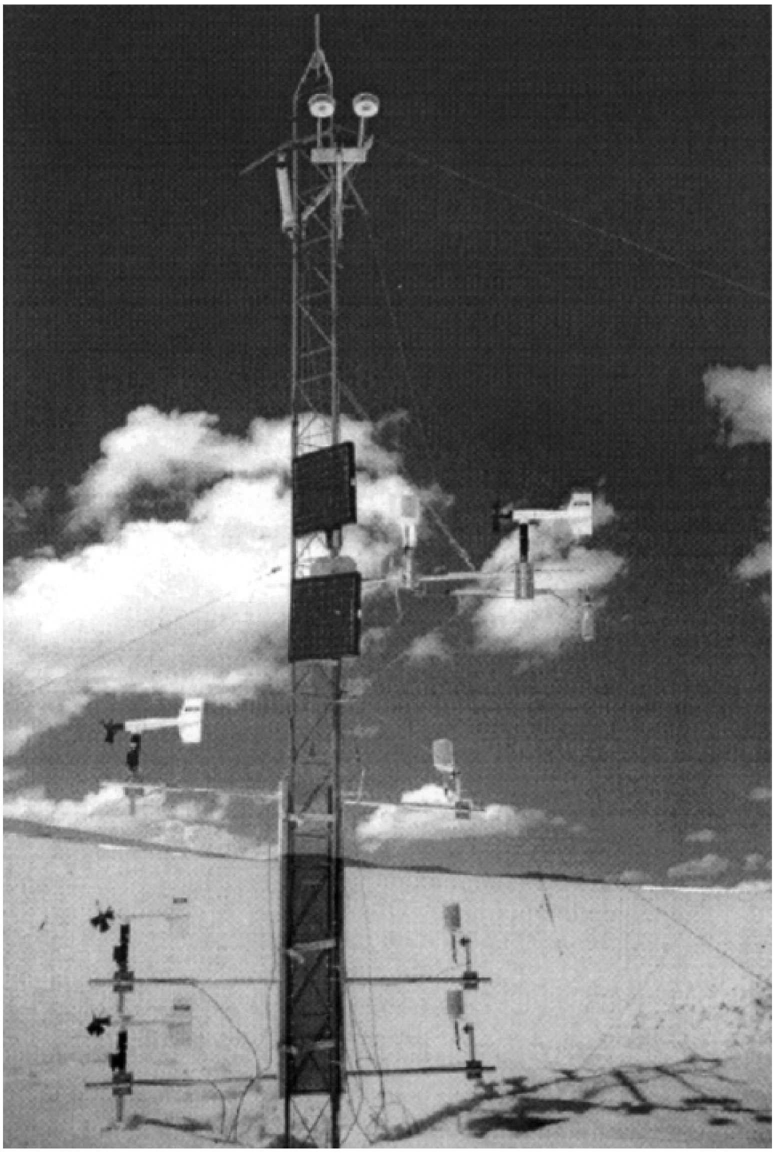

Figure 2. The meteorological tower at the subnivean laboratory on Niwot Ridge. Temperature, relative humidity and wind speed are collected at 0·5, 1·0 and 2·0 meters above the snowpack to calculate latent heat fluxes via the aerodynamic profile method. Instrumentation is mounted on a fixed mast which can be adjusted according to snow height

radiometer attached to the same tower at the 6 m height recorded net all-wave radiation. The instruments were monitored each weekday to ensure continuous and accurate data collection.

Data were collected at ten second intervals and means recorded every ten minutes. Data logging was done with a Campbell CR21x data logger in a subnivean laboratory located 20 m away and stored on a solid state storage module. The instruments were cross-calibrated with each other for one week before the period of measurement. One of the three instruments was used as a standard and the other two were calibrated to fit the standard using linear regression equations. R^2 values for the regression fits were all greater than 0·99. The RMSE of the regression fits were less than 0·1 °C for air temperature, about 0·5% for relative humidity, and less than 0·1 m/s for wind speed. The nominal instrument error for the relative humidity measurements is $\pm 2\%$. A comparison of the humidity measurements at the three heights shows that slopes of linear regression analyses ranged from 2 to 11%, equal to or greater than the nominal instrument accuracy of 2% and much greater than the 0·5% difference from the field calibration.

Latent heat fluxes

We followed the protocol developed by Cline (1995, 1997a, 1997b) to calculate latent heat fluxes between the snowpack and the atmosphere. The latent heat fluxes throughout the surface boundary layer were estimated using aerodynamic formulae with corrections for stability. The latent heat flux (Q_E) is expressed as:

$$Q_E = \bar{\rho}(L_V)\left(\frac{k(q_2 - q_1)}{\phi_E[\ln(z_2/z_1)]}\right)\left(\frac{k(u_2 - u_1)}{\phi_M[\ln(z_2/z_1)]}\right) \tag{1}$$

where $\bar{\rho}$ is the density of air calculated as a function on barometric pressure, L_V is the latent heat of vaporization of water calculated as a function of air temperature, k is von Karman's constant (0·4), ϕ_E is the stability function for water vapor, ϕ_M is the stability function for momentum, z_1 and z_2 are the instrument heights in the profile, q_1 and q_2 are the specific humidities at the given profile height, and u_1 and u_2 are the horizontal wind speeds at the given profile height.

The specific humidity at each profile level is calculated by Saucier (1983):

$$q = \frac{0{\cdot}622(e)}{P - 0{\cdot}378(e)} \tag{2}$$

where P is the atmospheric pressure and e is the vapor pressure from the equation:

$$e = \left(\frac{e_s(RH)}{100}\right) \tag{3}$$

RH is the relative humidity at a given level and e_s is the saturation vapor pressure over water calculated from the equation: $e_s = 6{\cdot}11 \text{ mb} \times 10^{aT/(T+b)}$. In this case, T is the air temperature (°C) at each level and a and b are constants.

The stability functions over the snowpack were calculated as a function of the Richardson number (Ri) as described by Ohmura (1982) for use over arctic snowcovers using the equations shown in Table I.

$$Ri = \frac{g}{\bar{\theta}}\left(\frac{\delta\theta/\delta z}{(\delta u/\delta z)^2}\right) \tag{4}$$

and g is the acceleration due to gravity.

The principle difficulty with the aerodynamic profile method is that it assumes the latent heat flux is constant throughout the atmospheric layer being measured. However, this assumption breaks down in conditions of blowing snow because vertical fluxes of heat and water vapor should vary with height as transported snow particles sublimate in the boundary layer (Morris, 1989). In an attempt to compensate for

Table I. Richardson Criteria used for calculating stability functions in the aerodynamic profile method

Stability function	Richardson Criteria (Ri)		
	$Ri < -0.03$	$-0.03 \leqslant Ri < 0$	$0 < Ri < 0.19$
ϕ_M	$(1 - 18Ri)^{-0.25}$	$(1 - 18Ri)^{-0.25}$	$(1 - 5.2Ri)^{-1}$
ϕ_E	$1.3\ (\phi_M)$	ϕ_M	ϕ_M

this problem in the measurement program, the latent heat flux was calculated between 2·0 m and 1·0 m, 1·0 m and 0·5 m, and 2·0 m and 0·5 m. The average of these three individual flux calculations is reported here for each time step. A more appropriate approach for calculating temperature and humidity profiles for short time periods during blowing snow conditions is described by Schmidt (1982), however that approach was not feasible for this long-term study.

The measured latent heat fluxes were then used to calculate the mass transfer of water between the snowpack and the atmosphere for each ten minute time step by converting the latent heat flux directly into millimeters of water equivalent lost to sublimation or gained by condensation. During the snow accumulation season, the latent heat fluxes were converted to millimeters of water lost or gained using the latent heat of sublimation. Following the onset of snowmelt, the latent heat of vaporization was used for the conversion because of the presence of liquid water in the snowpack. The starting date for snowmelt was determined by snowpack temperature measurements at a snowpit adjacent to the meteorological tower (Cline, 1997b).

Data quality

Data collection was extremely reliable for nine of the ten months in the study period. A malfunction in the data logger caused data loss from midnight on 29 March until noon on 1 May. Consequently, latent heat fluxes for April could not be calculated. For the nine months of record outside of April, 97% of the latent heat flux estimates were made using measured meteorological variables. The few missing values in the nine months of record were interpolated from existing data as follows. When three or less consecutive ten minute time steps were missing, the values were filled in with the average of the individual time steps immediately before and after the gap. If more than three consecutive time steps were missing, the missing values were taken to be the average of the latent heat fluxes for the twelve hours before and twelve hours after the data gap.

Regression analysis on a selection of the data collected indicates that meteorological measurements at the various heights tracked each other extremely well. A least squares regression between the ten minute average wind speeds measured at 0·5 m and 2·0 m above the snowpack demonstrated an R^2 value of 0·98 ($p < 0.001$). The analysis was based on a random sampling of 5% of the time steps in the four month period from October through December to reduce the potential effects of serial correlation. A similar regression on measured relative humidity values resulted in an R^2 value of 0·95 ($p < 0.001$). This is important because the similarity hypothesis, which assumes that the latent heat flux is constant throughout the atmospheric boundary layer, is a primary assumption of the aerodynamic profile method. A failure by the instruments at the three heights to effectively track each other could indicate a breakdown in the similarity hypothesis.

Missing meteorological data

The missing meteorological data for the month of April was replaced with data from the meteorological station at the tundra laboratory located 100 m northeast of the study site (Figure 1). Records from this station provide hourly average wind speed, relative humidity and temperature for the entire snow year. Wind speed was recorded at 8 m, while relative humidity and temperature were recorded at the 1·5 m height. These meteorological data were used to calculate latent heat fluxes for April using the SNTHERM model (Jordan, 1991). SNTHERM is a physically based point energy and mass balance snowcover model. The methods used by SNTHERM to calculate latent heat fluxes are fundamentally different from those used for the APM, but

Cline (1997b) has shown that the two methods agree well at Niwot Ridge. SNTHERM was initialized using snowpack conditions from the index snowpit (Figure 1) and latent heat fluxes were calculated on an hourly time step for the month of April.

Snowpack properties and annual precipitation amount

The presence of snow beneath the meteorological tower in the fall was determined by reflected shortwave radiation measurements using a Kipp and Zonen CM14 Albedometer and verified with field observations. Snow water equivalence (SWE) was measured weekly throughout the snow season at an index snowpit approximately 40 m north of the meteorological tower (Williams *et al.*, 1996) (Figure 1). Summer precipitation was collected in a shielded Belfort recording gauge located about 100 m east–northeast of the meteorological tower (Figure 1). Net annual precipitation amount was calculated by adding the summer precipitation total to the maximum SWE measured at the index snowpit and subtracting the calculated net sublimation for the snow season.

RESULTS

Meteorological conditions

The study period began on 1 October 1994 and ended on 16 July 1995. There was intermittent snowcover at the study site in September with continuous snowcover beginning on 4 October. Maximum snow accumulation was on 1 June and was 1·308 m of water. Snowmelt began on 2 June and ended with the complete ablation of snow beneath the meteorological tower on 16 July. The snow accumulation season is therefore defined as 1 October to 1 June while the snowmelt season lasted from 2 June until 16 July.

The 1994–1995 snow year was characterized by heavy precipitation concentrated in the late spring. This resulted in significantly greater than average snow accumulation at the instrumentation site (Cline, 1997a). Records from a USDA SNOTEL site located 5 km southeast of the subnivean lab indicate that the snowpack was 134% of normal at the end of April, before record snowfall events in May and June. As a result, snowmelt, which typically begins in May, was delayed until the beginning of June and snow persisted beneath the instrument tower well beyond the normal ablation date.

Average daily wind speeds were generally greatest in the fall and winter during the snow accumulation season and lower in the spring (Figure 3). The mean daily average wind speed for the season was 7·3 ms^{-1} with a maximum of 20·2 ms^{-1} in late October and a minimum of less than 2 ms^{-1} in early February (Table II). Average daily air temperatures were consistently below 0 °C from mid-October until mid-May with a low of −20·5 °C in late November. After the middle of May, average daily temperatures fluctuated but showed a consistent upward trend until the end of the study period with a high of 13·0 °C in July. The mean air temperature for the snow season was −4·9 °C. Specific humidity was less than 2 g kg^{-1} for the bulk of the snow accumulation season and increased sharply during snowmelt season. The minimum specific humidity was 0·46 g kg^{-1} in early January and the maximum was 7·76 g kg^{-1} in mid-July. December had the lowest monthly average specific humidity at 1·35 g kg^{-1}, although every month from November through April had an average specific humidity of less than 2 g kg^{-1}.

Table II. Mean and range of average daily wind speed, air temperature and specific humidity for the 1994–1995 snow season

Variable	Mean	Range
Wind speed	7·3 m/s	<2·0 to 20·21 m/s
Air temperature	−4·9 °C	−20·5 to 13·0 °C
Specific humidity	3·15 g kg^{-1}	0·46 to 7·76 g kg^{-1}

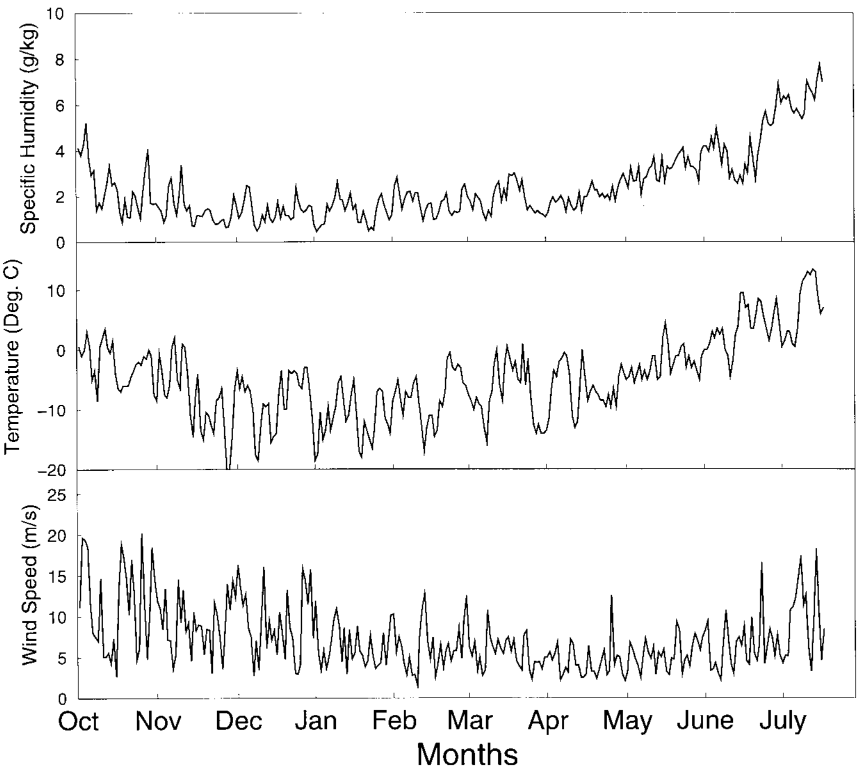

Figure 3. Mean daily wind speed, temperature and specific humidity at the Niwot Ridge saddle for the 1994–1995 snow season. Tick marks denote the beginning of each month. Wind speeds are highest in the fall and early winter corresponding with the periods of greatest net sublimation. Air temperature is consistently above freezing beginning in May. Similarly, specific humidity increases throughout the spring and peaks during snowmelt season

Seasonal water fluxes

Sublimation losses from the seasonal snowpack dominated water gain from condensation over the majority of the snow season. The breakdown of monthly contributions to the net seasonal sublimation total shows that sublimation is heavily biased towards the snow accumulation season (Figure 4). During the period from October through March, 226 mm of water equivalent sublimated from the snowpack. And 75% of this sublimation occurred relatively early in the season, between October and the end of January. Trends in net sublimation correspond reasonably well with prevailing wind speeds which were greatest in the fall and winter. November is the only single month with greater than 50 mm of sublimation (53·8) and overall it accounts for nearly 28% of the net seasonal sublimation. After November, monthly sublimation totals remain between 28–40 mm through the end of March.

As the snowpack warmed towards isothermal at 0 °C in May, there was a shift from net water loss to net water gain at the snow surface. From the beginning of May through snowpack ablation at the study site on 16 July, 30·4 mm of water equivalent was returned to the snowpack in the form of condensation. The bulk of this condensation occurred in May (17 mm). June showed only 5·0 mm of net condensation. The condensation total for July was similar to that seen in May with a net total of 9 mm of condensation in only 16 days. The net condensation during May, June, and July does not come close to compensating for sublimation losses during the snow accumulation season. It does, however, reduce the net seasonal moisture flux to the atmosphere total by nearly 14%.

Seasonal trends in net sublimation and condensation correspond with changes in meteorological conditions at the site over the snow season, particularly prevailing wind speeds and specific humidity. Wind speeds

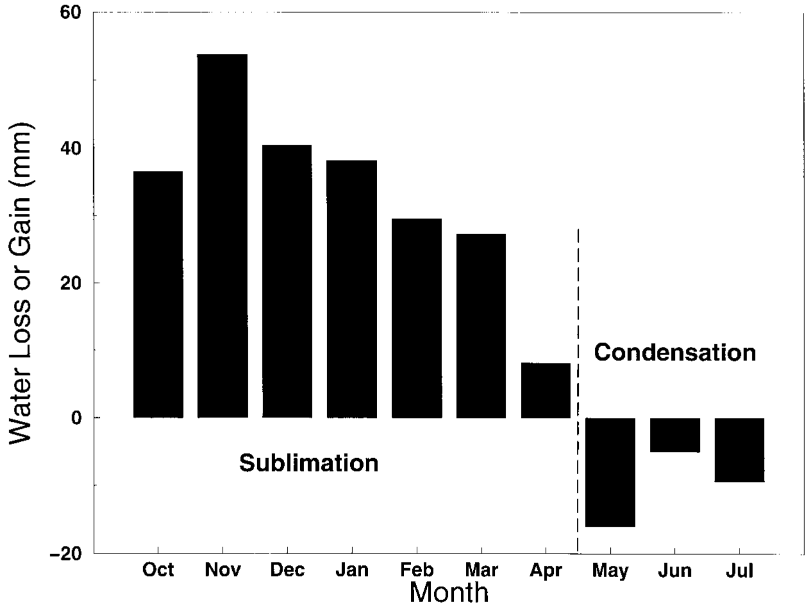

Figure 4. Monthly net sublimation and condensation for the 1994–1995 snow season. Most of the sublimation occurs during the snow accumulation season. During snowmelt season in the early summer, there is net condensation from the atmosphere to the snowpack

at the site were greatest in the fall and winter and relatively low in the spring and early summer. The average mean daily wind speed for the months which demonstrated net sublimation (October–April) was 1·3 ms^{-1} higher than for the months which showed net condensation (May–July). At the same time, specific humidity remained low ($<$2–3 kg/g) for the bulk of the snow accumulation season and then showed an upward trend beginning in April (Figure 3). The result of these conditions is that the snow accumulation season was characterized by high levels of turbulent mixing and a relatively high water vapor gradient between the snowpack and the atmosphere, both of which are conducive to sublimation.

After 1 May, daily mean specific humidity was commonly above 4 g kg^{-1}. The temperature and specific humidity of the snowpack have upper limits of 0 °C and 4·85 g kg^{-1} respectively. Consequently, during the late spring and summer the specific humidity of the atmosphere was commonly higher than that of the snow surface. The combination of increased atmospheric specific humidity and decreased wind speeds made condensation much more likely during the snowmelt season.

Sublimation losses and the annual hydrologic budget

The net seasonal moisture flux was calculated by summing the sublimation and condensation amounts for each ten-minute time step over the entire snow season. In total, there was a net loss to sublimation of 195 mm of water equivalent from the seasonal snowpack (Figure 5). Maximum snow accumulation as determined by regular snow surveys was 1308 mm of water equivalent on 8 June. Consequently, the total water equivalent in the seasonal snowpack at the site was reduced 15% through sublimation.

Total summer precipitation at Niwot Ridge was 203 mm of water equivalent based on measurements made with a shielded Belfort gage from 9 June until the end of the water year on 30 September. Therefore, total annual precipitation for the 1994–1995 water year, excluding net losses to sublimation, was 1511 mm of water equivalent. For the entire water year, water losses to sublimation were equal to 13% of annual precipitation at Niwot Ridge. Interestingly, the net seasonal sublimation total of 195 mm was only slightly less than the 203 mm of water equivalent the area received as rainfall over the entire summer season.

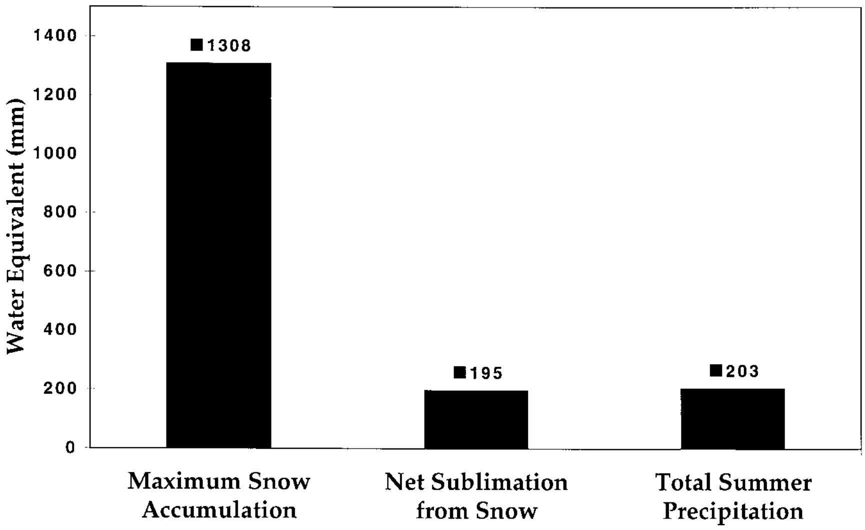

Figure 5. Mass loss from the snowpack in the context of other precipitation totals. Sublimation from the seasonal snowpack is an important component of the alpine hydrologic cycle. The 195 mm of water equivalent lost to sublimation was greater than 15% of the maximum snow accumulation and nearly equivalent to the total precipitation for the entire summer season

Importance of episodic events

We evaluated moisture fluxes from three representative months in order to understand whether the movement of water between the snowpack and the atmosphere was driven primarily by episodic events. November was chosen because it had the highest net sublimation of any month, February was typical of mid-winter snow accumulation conditions, and June demonstrated net condensation. Time series of hourly sublimation amounts from these three months provide insight into the trends in sublimation/condensation over short time intervals. We report hourly fluxes rather than ten minute fluxes because of the low magnitude of the ten minute fluxes.

The records from November and February show that sublimation is a somewhat episodic phenomenon (Figure 6). There are periods in the record where sublimation rates continue below 0·05 mm/hr for several days (e.g. 24–29 November and 20–28 February). However, these low sublimation losses are often punctuated by short 'sublimation events' during which sublimation rates remain greater than 0·02 mm/hr for longer than 24 hours (e.g. 22–24 November and 16–17 February). These sublimation events are a result of Chinook conditions which are typified by warm, dry downslope winds and are common in the Colorado Front Range, particularly in fall (Barry, 1992).

Sublimation events typically last between one and two days and account for a large portion of the total seasonal sublimation. Three such events in November (9–10, 22–24, and 29–30 November) totaling six and a half days accounted for 21·5 mm of sublimation which was 40% of the monthly net sublimation. In particular, the two day period from noon on 22 November until noon on 24 November accounted for 11·1 mm or 21% of the net monthly sublimation. A similar effect is evident in February. The seven day period between 13–19 February accounted for 17·4 mm of net sublimation or 58% of the monthly total.

The highest rates of sublimation seen in these records are on the order of 0·4 mm/hr during November. However, these periods of heightened sublimation rarely last longer than 12 hours and often only two or three hours. Sublimation rates in February exceed 0·3 mm/hr only once. For both months, sublimation rates are less than 0·1 mm/hr for a majority of the time. Many of the periods which exhibit low rates of sublimation appear to demonstrate an underlying diurnal periodicity (for example 4–8 November and 20–28 February). During these periods, sublimation rates peak during the day and decline to near zero during the night. In November, these periods of low sublimation (1–8, 11–21, and 25–28 November)

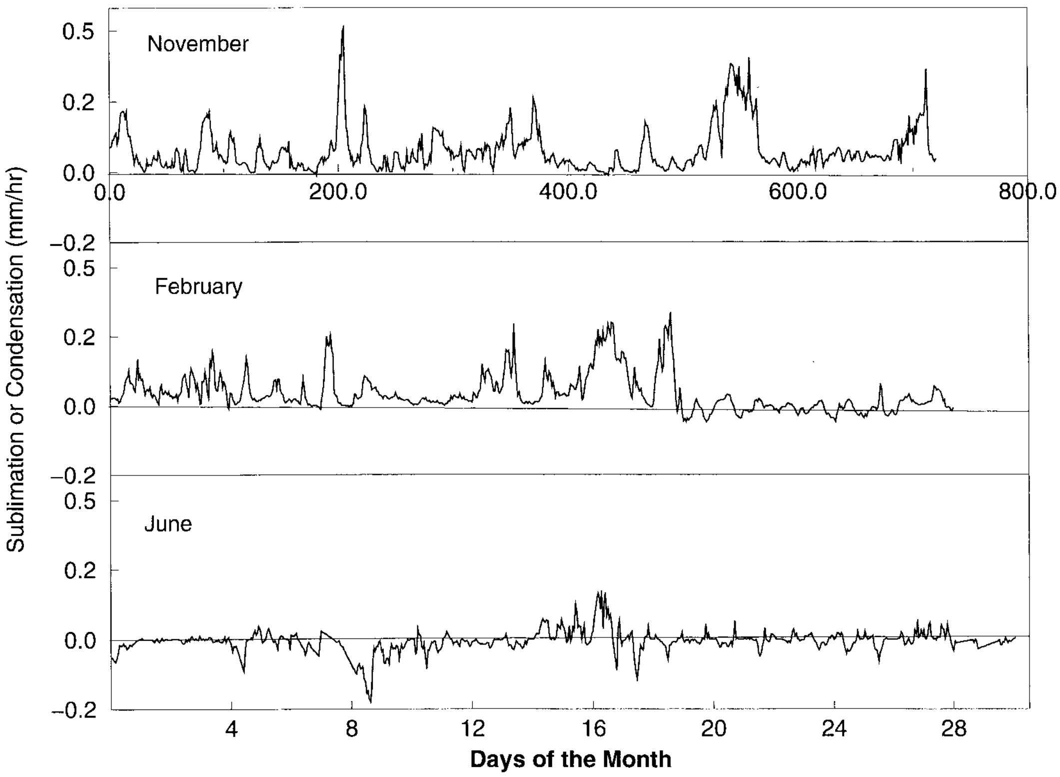

Figure 6. Hourly moisture fluxes between the snowpack and atmosphere for November, February and June. Positive fluxes indicate sublimation while negative fluxes indicate condensation. November and February are dominated by sublimation, whereas June features both sublimation and condensation as well as significantly lower moisture flux rates

account for 60% of the net sublimation for the month. So the small diurnal fluctuations are responsible for more monthly sublimation than the more obvious 'sublimation events'. This pattern does not hold true for February. The two prolonged periods with low sublimation rates and an apparent diurnal periodicity (1–12 and 20–28 February) account for only 42% of the net monthly sublimation.

The hourly record for the month of June shows a fairly even split between sublimation and condensation. The rate of moisture transfer between the atmosphere and the snowpack was below 0·1 mm/hr for most of the month. There was one prolonged period of elevated condensation rates (7–8 June), as well as a small sublimation event on 16 June. As with November and February, there are periods, such as 18–24 June, which appear to exhibit a small diurnal periodicity with higher sublimation rates during the day.

Meteorological driving forces

In an effort to better understand the meteorological controls on water fluxes between the snowpack and the atmosphere, we compared a time series of ten-minute averages for wind speed, temperature, specific humidity, and net radiation with the moisture flux records for two one-week periods during the snow season. The first time period, 20–27 November, encompasses the largest sublimation episode seen in our records. The second time period, 22–28 February, demonstrates a diurnal pattern in the moisture flux record.

For the first time period, the elevated rates of sublimation seen on 23–24 November were associated with wind speeds that were consistently above 10 ms^{-1} (Figure 7). During the same period, air temperatures rose from well below $-10\,^{\circ}C$ to between -4 and $-8\,^{\circ}C$ and specific humidity remained below 2 g kg^{-1}. In contrast, the specific humidity in the pore spaces at the snow surface would have been higher than 2 g kg^{-1}

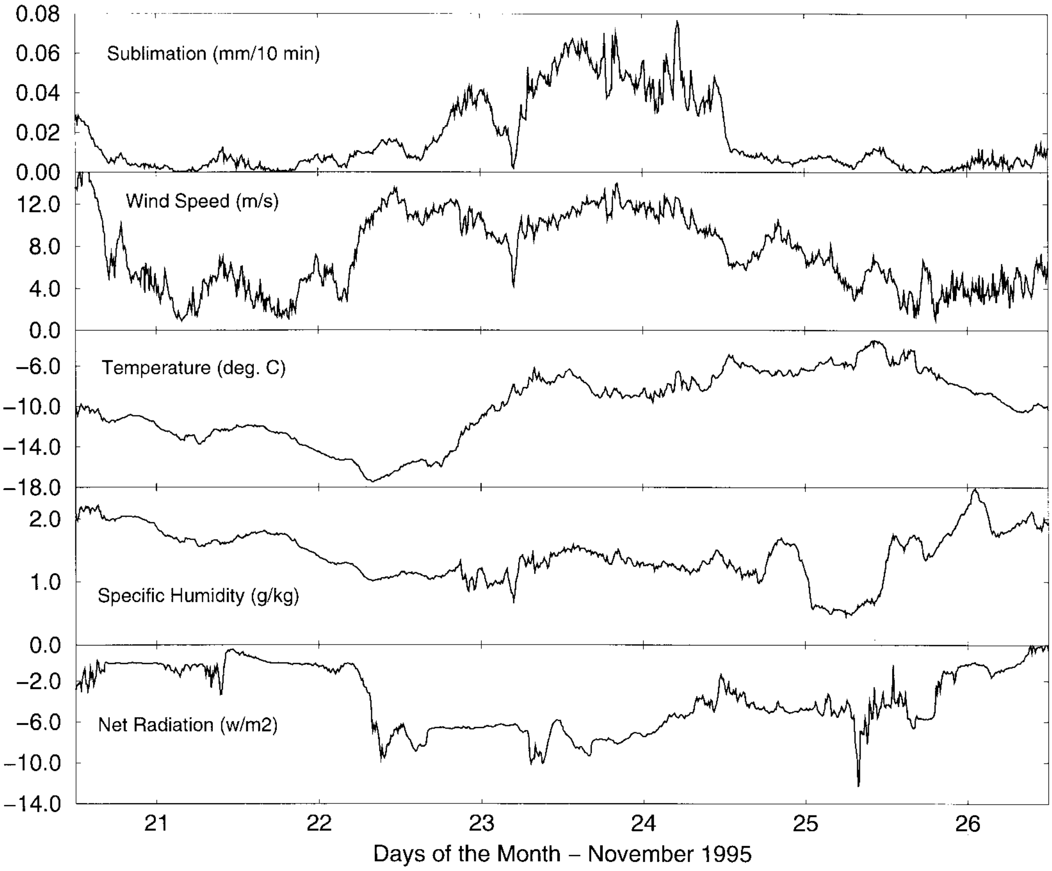

Figure 7. Times series of ten-minute averages for sublimation, wind speed, temperature, specific humidity, and net radiation during six days in November. Tick marks denote the beginning of each day. All measurements were taken from the 1·0 m height in the instrument profile except net radiation which was measured at 6·0 m. Sublimation was most closely correlated with wind speed ($r^2 = 0{\cdot}56$, $p \ll 0{\cdot}001$, $n = 865$), particularly during the sublimation event on 22–24 November ($r^2 = 0{\cdot}66$, $p \ll 0{\cdot}001$, $n = 239$)

within the observed temperature range. Observed trends in net radiation showed little connection to either air temperature or sublimation rates.

Sublimation during this period appears to have been most closely correlated with wind speed, however there are periods with high wind speeds and very low levels of sublimation (e.g. 22 and 24 November). A linear regression analysis between wind speed and sublimation for the whole time period showed that the two variables were reasonably well correlated ($r^2 = 0{\cdot}56$, $p \ll 0{\cdot}001$, $n = 865$). During the sublimation event, wind speed and sublimaton were even more highly correlated ($r^2 = 0{\cdot}66$, $p \ll 0{\cdot}001$, $n = 239$). Temperature, specific humidity, and net radiation were all uncorrelated with sublimation during this period ($r^2 < 0{\cdot}05$). Additionally, the highest recorded air temperatures and lowest recorded specific humidities (25 November) were associated with low sublimation rates ($<0{\cdot}02$ mm/10 min).

During the second time period in February, moisture fluxes were much smaller than those seen in November, absolute magnitudes rarely exceeded 0·01 mm/10 min (Figure 8). The moisture flux record showed a diurnal periodicity with low rates of sublimation during the day and low rates of condensation during the night. High wind speeds were associated with both sublimation (e.g. 22 and 26 February) and condensation (e.g. 25 and 27 February). However, both net radiation and air temperature tracked the moisture flux more closely than wind speed. Higher air temperatures were generally associated with sublimation (with the exception of 28 February). Similarly, the sharp decreases seen in temperature during

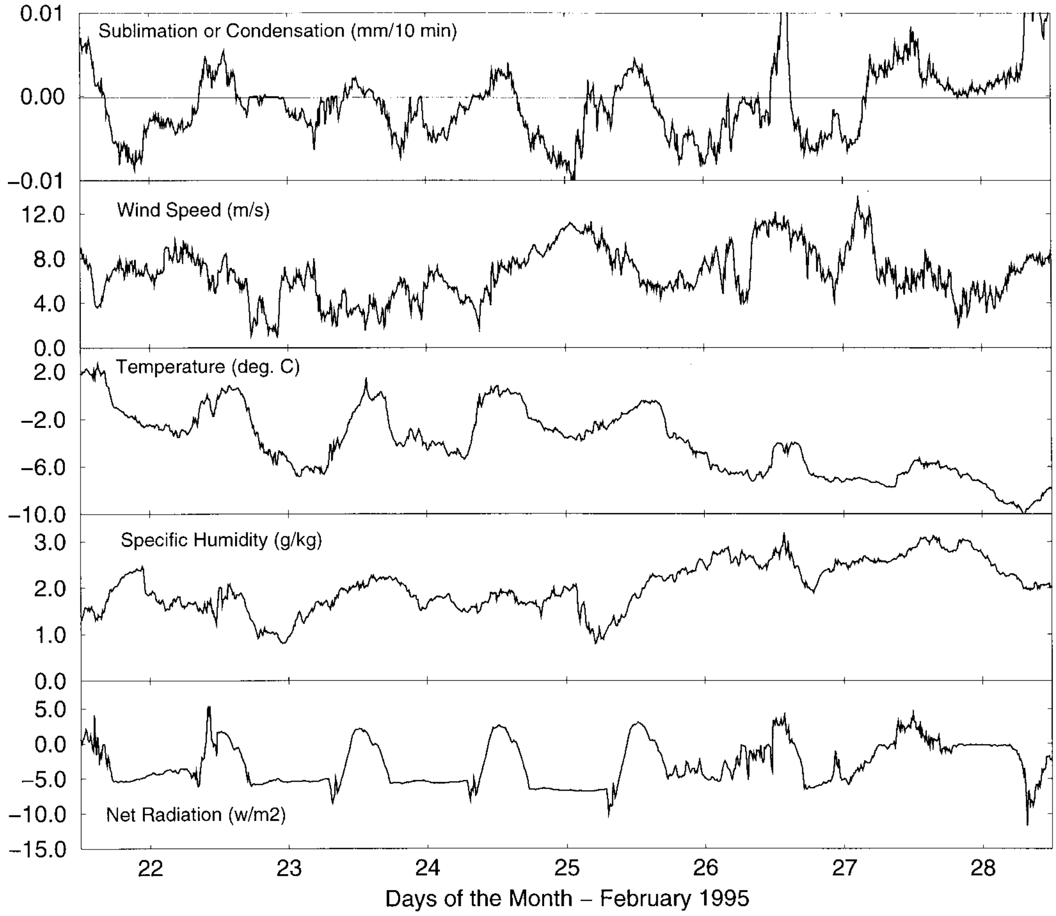

Figure 8. Times series of ten-minute averages for sublimation, wind speed, temperature, specific humidity, and net radiation during seven days in February. Tick marks denote the beginning of each day. All measurements were taken from the 1·0 m height in the instrument profile except net radiation which was measured at 6·0 m. Moisture fluxes were most closely correlated with net radiation ($r^2 = 0{\cdot}32$, $p \ll 0{\cdot}001$, $n = 1009$). During the first five days of this period, temperature also showed a moderate correlation with the moisture flux ($r^2 = 0{\cdot}29$, $p \ll 0{\cdot}001$, $n = 648$)

the night are associated with condensation from the atmosphere onto the snowpack. As in November, atmospheric specific humidity was relatively constant, ranging between approximately 1 and 3 g kg^{-1} and therefore showed little correlation with the moisture flux record, although there are several cases where decreased levels of specific humidity were associated with sublimation (e.g. 22, 25, and 28 February).

Regression analyses between moisture fluxes and the four meteorological variables showed that net radiation was the most closely correlated with trends in sublimation and condensation ($r^2 = 0{\cdot}32$, $p \ll 0{\cdot}001$, $n = 1009$). Over the seven day period, temperature showed little correlation to the moisture flux ($r^2 = 0{\cdot}005$, $p = 0{\cdot}03$, $n = 1009$). However, between noon on 21 February and the end of 25 February, the diurnal temperature signal tracked the moisture flux more closely ($r^2 = 0{\cdot}29$, $p \ll 0{\cdot}001$, $n = 648$). Unlike in November, wind speed showed little correlation with the moisture flux ($r^2 = 0{\cdot}01$, $p = 0{\cdot}001$, $n = 648$).

DISCUSSION

The trend at Niwot Ridge towards low rates of sublimation and increased rates of condensation during the melt season has significant implications for past studies which drew conclusions on seasonal sublimation

Table III. Previous studies on sublimation rates from snow. Mass transfer is described both as the amount of the energy budget that went towards sublimation instead of snowmelt and as a monthly rate of sublimation or condensation in millimeters of water equivalent. Positive numbers indicate losses to sublimation and negative numbers indicate gains from condensation. From Kuusisto (1986) and Pluss and Mazzoni (1994)

Reference	Location	Elev. (m)	Period	Daily latent heat flux ($W m^{-2}$)	Average monthly moisture flux (mm)
Dewalle and Meiman (1971)	open forest (Colorado)	3260	June (1968)	−6	5·5
de la Casinere (1974)	open mtn. field (France)	3550	July (1968)	−11	10·1
Moore and Owens (1984)	open mtn. field (New Zealand)	1450	melt season (1982)	34	−30·9
Harding (1986)	Finse (Norway)	1000	15 days in May (1986)	0·3	−0·3
Marks and Dozier (1992)	Emerald Lake	2800	May (1986)	−54	49·5
Pluss and Mazzoni (1994)	Swiss Alps	2600	18 days in May (1992)	−4	3·7
Fohn (1973)	Peyto Glacier, Canada	2510	14 days in July (1973)	14·5	−13·3
Martin (1975)	St. Sorlin Glacier, France	2700	11 days in summer	−3·5	3·2
Calanca and Heuberger (1990)	Urumqi Glacier (Tien Shan, China)	3900	summer (1990)	−15	13·8
Meiman and Grant (1974)	Colorado Front Range	2740	winter (1973–1974)	N/A	27·0

rates based entirely on data from the melt season. Many recent studies of moisture fluxes over snow measured sublimation only in the spring or early summer (Table III). As a result, their estimates of sublimation are biased toward the prevailing spring conditions and may not reflect the rate or magnitude of moisture transfers taking place throughout the winter. In a continental, alpine climate similar to Niwot Ridge, using data from the melt season to predict sublimation over the whole snow season will most likely result in an underestimation of sublimation, perhaps to quite a large extent. This will be particularly true in areas where spring and early summer are the primary precipitation season.

It is somewhat difficult to make comparisons between sublimation magnitudes recorded in this study to those in previous studies because of the short periods of record seen in many of the previous studies. However, the figure of 15% of total seasonal snow accumulation lost to sublimation compares favorably with the percentage of the snowpack estimated lost to sublimation in the Sierra Nevada by both Marks and Dozier (20%) and Kattelmann and Elder (1991) (18%) (Table III). Further, we report 154 mm of net sublimation for the five month period covering December–April, which is consistent with the 135 mm of water equivalent reported lost to sublimation for the same period in the Pingree Peak area by Meiman and Grant (1974).

The highest monthly condensation rate among the studies in Table III is 30 mm, recorded by Moore and Owens (1984) in New Zealand. This is about twice the highest monthly condensation rate we documented at Niwot Ridge of 17 mm in May. At the continental study sites, sublimation ranged from 3·2–13·8 mm/month which is 4–9 times lower than the 27–54 mm/month for the snow accumulation season at Niwot Ridge. However, because these studies were conducted in the spring and early summer, the low rates of sublimation are generally consistent with the melt season rates in the Niwot record. Overall, it is unclear whether or not the rates of sublimation and condensation reported in Table III are indicative of any trends in the transfer of water between the snowpack and the atmosphere over an entire snow season because of the lack of data from the snow accumulation season. At Niwot Ridge, the bulk of net sublimation occurred during the snow accumulation season. Additionally, it is possible that in a drier spring than the one seen in 1995, this net sublimation could continue through the snowmelt season which would be more consistent with the late spring and summer sublimation rates documented in Table III.

On a shorter time scale, an examination of hourly sublimation rates reveals two contrasting patterns in the record. First, there are clearly short periods where rates of mass loss from the snowpack remain elevated for a period of a day or more. King *et al.* (1996) have documented similar sublimation events which contribute disproportionately to the surface energy and mass budgets of an Antarctic snowcover. Observed sublimation events at Niwot Ridge were most closely correlated with wind speed. Thus it appears that strong turbulent mixing is necessary to enhance the movement of water vapor along prevailing gradients during sublimation events. In a broader sense, multi-day periods with elevated rates of sublimation are probably driven by specific synoptic weather patterns. This hypothesis is supported by Cline (1997a), who demonstrated a link between trends in the overall snowpack energy balance and prevailing synoptic weather patterns at Niwot Ridge. In particular, the Colorado Arch pattern, which results in warm air flows into Colorado from the southwest, is associated with warm, dry conditions conducive to sublimation. If sublimation is indeed associated with particular synoptic patterns, it may be possible to predict when elevated rates of sublimation are likely to occur based on weather forecasts.

The second pattern evident in the hourly moisture flux record is that, at times, there appears to be a diurnal periodicity in the movement of water between the snowpack and the atmosphere (Figure 6). McKay and Thurtell (1978) have previously reported that the latent heat flux over a snow cover does not display a marked diurnal periodicity because it is not dependent on incoming solar radiation. However, similar to the pattern seen in periods of the Niwot record, Meiman and Grant (1974) report that afternoon sublimation rates at an open site were on average three times as high as those recorded at night.

In contrast to periods of elevated sublimation, the small diurnal fluctuations evident in the moisture flux record at Niwot Ridge (Figure 8) appear to be driven more by temperature than by wind speeds. During periods when the specific humidity of the snowpack and the atmosphere are close in magnitude it appears that diurnal fluctuations in air temperature can reverse the direction of the water vapor gradient at the snow surface, causing a switch between sublimation and condensation. During the day, warming at the snow surface raises the specific humidity in the snowpack pore spaces. When the amount of water vapor in surface pore spaces rises above the ambient specific humidity in the atmospheric boundary layer, sublimation will occur. Similarly, cooling of the snow surface at night depresses the specific humidity in snow surface pore spaces which can lead to condensation from the atmosphere onto the snowpack. The water vapor gradients during periods showing a diurnal fluctuation between sublimation and condensation are generally small in magnitude. As a result, sublimation rates are low even when wind speeds are high.

The somewhat episodic nature of the movement of moisture between the snowpack and the atmosphere highlights the need to sample data on short time intervals over a long period. Sampling during 'sublimation events' where sublimation rates remain elevated for an extended period, will undoubtedly lead to an overestimation of the importance of sublimation in the hydrologic cycle of alpine areas. For example, it is possible that the high rates of sublimation documented in the White Mountains by Beaty (1975) are primarily a function of sampling during a period which was heavily biased toward sublimation. Similarly, sampling sublimation during different seasons will yield very different results depending on the prevailing meteorological conditions. Clearly caution must be used when attempting to extrapolate estimates of sublimation between seasons or years.

As a general error analysis, there are two main sources of error in our calculations. The first of these is sublimation from blowing snow. The occurrence of blowing snow is variable, even at high wind speeds, because it is closely correlated with snow surface conditions. Work on the Canadian Prairies using a physically-based blowing snow model has shown that the percentage of annual snowfall removed from a 1-km fetch by saltation and suspension ranges from 4·5 to 39·4% (Pomeroy and Male, 1995). More than half of this wind-transported snow can sublimate before reaching the field edge (Pomeroy and Gray, 1994). These estimates indicate that sublimation from blowing snow can account for 2–20% of the annual snowfall amount.

The occurrence of blowing snow at Niwot Ridge clearly has the potential to influence estimates of sublimation over the snow season. Paradoxically, the consistently high wind speeds at Niwot Ridge result in rapid increases in surface hardness after precipitation events. This increased surface hardness dramatically decreases the likelihood of blowing snow. As a result, it is likely that at Niwot Ridge the percentage of the annual snowpack lost to sublimation from blowing snow would be on the lower end of the range estimated by Pomeroy and Gray (1994) and Pomeroy and Male (1995).

Sublimation from blowing snow also has the potential to confound latent heat fluxes estimated by the APM by serving as an outside source of water vapor at the measurement site. However, daily visual records from November indicate that more than half of the largest sublimation events did not occur during periods of blowing snow. This finding agrees with work done by Berg (1986), who showed that November has the lowest frequency of hours with blowing snow of any month during the snow accumulation season at Niwot Ridge. In addition, January, which consistently demonstrates the highest frequency of blowing snow at the site (Berg, 1986), had less net sublimation than either November or December. These facts support the conclusion that the occurrence of blowing snow did not consistently bias the largest measured latent heat fluxes. Overall, the error caused by blowing snow is most likely less than 20% of the measured water vapor fluxes at our site.

The second potential source of error is the accuracy of the aerodynamic profile measurements. Recent work at Niwot Ridge has shown that the aerodynamic profile method tends to underestimate the magnitude of latent heat fluxes above snowcover by 36% when compared with high frequency eddy correlation instrumentation (Hood and Williams, in review). Therefore, it is possible that the net sublimation at the site is actually greater than the 195 mm we report.

The establishment of a comprehensive, season-long data set for sublimation at Niwot Ridge is an important step towards both understanding the role of sublimation in the alpine hydrologic cycle and estimating its importance on the catchment scale. Further work looking at sublimation in lower snow years will reveal the extent to which sublimation rates vary on an annual basis. This information will be useful in predicting sublimation rates both at other alpine sites and under conditions of changing climate parameters.

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