Estimating Alter-Shielded Gauge Snowfall Undercatch, Snowpack Sublimation, and Blowing Snow Transport at Six Sites in the Conterminous United States

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ABSTRACT

Computing the monthly and winter water balance for cold regions can be difficult due to data scarcity. Historically, the spatial and temporal resolution, and the number of variables measured have been limited. Currently these data are once again becoming limited. To estimate the net snowpack accumulation, measured precipitation must be adjusted to consider precipitation underestimation due to gauge undercatch, the snow lost to sublimation, and blowing snow transport. Using existing formulations, hourly meteorological data were used to estimate snowpack sublimation and blowing snow transport losses for three winters at six National Weather Service (NWS) Automated Surface Observation Stations across the conterminous United States. Wind-induced undercatch was estimated from daily data for the colocated NWS Alter-shielded gauges. For the average wind speed sites (the average wind speed was from 2.4 to 4.3 m/s), 70% of the snow that fell was caught, while at the low wind site (1.3 m/s), 90% was caught and only 46% was caught at the high wind site (5.6 m/s). Average snowpack sublimation ranged from 7 mm per month at either low wind or low precipitation sites to over 20 mm per month at average wind sites with either average precipitation and low humidity or high precipitation and moderate humidity. Blowing snow transport was only important at higher wind sites (>4 m/s). A distinct relationship was not obvious for average monthly meteorology for undercatch versus snowpack sublimation plus blowing snow losses. Seasonally, they are approximately equal for more snowy and wet environments.

Keywords: snowfall undercatch, snow sublimation, blowing snow transport, water balance, meteorological stations

INTRODUCTION

Hydrological data are sparse in regions where snow processes are dominant, such as alpine and polar areas. The scarcity of these data is due to low populations in the often harsh environments and, in recent years, budget constraints. Limitations and reductions in data collection have decreased the number of operating stations (Shiklomanov et al., 2002), and in some cases decrease the types of data collected. While automated stations that record observations on a hourly basis are currently being operated at most airports in many countries, the location of such stations are not necessarily representative of the more extreme hydrological environments, such as in mountainous areas.

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Some hydrological and environmental studies require the calculation of the monthly, seasonal or annual water balance studies. For cold regions, the snowpack often accumulates until ablation starts to generate water for runoff. To determine net accumulation, water balance design studies require good estimates of the amount of snow added to the pack and the amount of snow lost to sublimation and from blowing snow transport. While these processes can occur at hourly or shorter time intervals, observations are often scaled up to monthly, seasonal or annual time steps. If observations do not exist, calculations are performed using hourly or daily data, but results are required for the longer time intervals. Much historical data were only collected daily and for some computations may be too coarse in temporal resolution to adequately capture sub-daily fluctuations.

The difference between the actual amount of snow and the amount measured by a precipitation gauge is called undercatch. Snowfall undercatch is greater than rainfall undercatch (Larson and Peck, 1974); mostly because snow particles have significantly larger drag forces than do rain drops, enabling wind to more easily blow snow particles away from the mouth of a precipitation gauge. To improve the measurement of solid precipitation, various wind shielding devices have been developed to alter the turbulent air flow around the gauge mouth in order to increase the gauge particle catch efficiency (e.g., Alter, 1937; Larson, 1971; Rechard and Wei, 1980; Goodison et al., 1983; Goodison et al., 1998). The first study by Alter (1937) introduced the Alter Shield that is now the standard in the United States. However, at wind speeds faster than 2 m/s at gauge height, enough turbulence is still produced that gauge catch efficiencies are reduced (Goodison et al., 1998). A number of studies have attempted to quantify the undercatch of different shield and gauge configurations (e.g., Larson and Peck, 1974; Goodison, 1978; Legates and DeLiberty, 1993; Metcalfe and Goodison, 1993; Yang et al., 1993; Goodison et al., 1998; Yang et al., 1998a; 1998b; 1999). Undercatch can be estimated relatively well as a function of only temperature and wind speed (Goodison et al., 1998).

The potential snowpack sublimation rates are a function of vapour pressure deficits, wind speeds and surface pressure (Sverdrup, 1936), while blowing snow transport has been estimated as a function of wind speed (e.g., Pomeroy et al., 1991). These rates change on an hourly or even sub-hourly basis. However, historical meteorological data are typically available only for daily time steps and many stations only measure precipitation and temperature, especially stations that started recording meteorological conditions fifty or more years ago.

Besides meteorological conditions (e.g., temperature, vapour pressure, surface pressure) and physical characteristics (e.g., terrain features such as surface and obstacle roughness, gauge and shield shape and size), sublimation and solid precipitation undercatch are both a function of wind speed. Considering that undercatch would increase the observed mass added to a snowpack while sublimation and blowing snow transport would decrease the snowpack storage, it is hypothesized that the proportion of these two quantities for the monthly water balance can be predicted as a function the local climate conditions. There are two objectives for this paper: 1) to determine if the quantities of snow losses (i.e., precipitation undercatch, snowpack sublimation and blowing snow transport) are significant in terms of the monthly and seasonal water balance; and 2) if the snow losses are significant, how comparable are they to each other and in what environments?

METHODOLOGY

Gauge Undercatch

The most significant problems associated with the measurement of solid precipitation according to the World Meteorological Organization were the systematic errors caused by wind undercatch, wetting, and evaporation losses (Goodison et al., 1998). To correct for the underestimation due to wind, best fit regression equations were derived as a function of gauge height wind speed, using the Double Fence Intercomparison Reference (DFIR) gauge as the standard (Goodison et al., 1998). Yang et al. (1998a) derived the following relationship for the snowfall (T # 0^{o}C) catch ratio for an Alter-shielded 8" NWS Standard precipitation gauge with respect to the DFIR gauge:

$$\frac{ALTER}{DFIR}(\%)_{snow\,(T<0C)} = \exp\left(4.606 - 0.036\, U_a^{1.75}\right) \quad (1).$$

where U_a is the average daily wind speed (in m/s) at the gauge height. Yang et al. (1998a) defined mixed precipitation by average daily air temperatures between 0 and +3°C, not by phase, and the catch ratio was a fitted equation:

$$\frac{ALTER}{DFIR}(\%)_{mixed\,(0<T<3C)} = (101.04 - 5.62\, U_a) \quad (2).$$

Equation 1 and 2 are computations using average daily data; temperatures will vary over a day and the phase of precipitation can also vary when air temperatures are slightly warmer than freezing. Upadhyay (1995) illustrated that 98% of precipitation < 0 NC falls as snow but at temperatures from 0 to 3°C a large proportion of precipitation fell as rain or snow with a minority percentage falling as mixed precipitation. Fassnacht et al. (2001) summarized a number of relationships between the probability of snow and near surface air temperature and mentioned that humidity is the most important factor in determining the shape of relationships. Specifically precipitation can fall as snow at near surface air temperatures warmer than 6°C for dry environments (e.g., Wyoming) while this threshold can be 4°C or colder for wet, maritime environments (e.g., California). The relationship developed by Yang et al. (1998a) uses data from three different environments, specifically, Valdai, Russia, Reynolds Creek ID, and Danville VT. The maximum snow threshold temperature ranged from –4.1°C to 2.6°C, mixed precipitation was observed between –8.6 and 7.3°C, while the minimum rain threshold was between –1.6 and –0.3°C. Therefore, equations 1 and 2 will be used for the suggestions temperature ranges.

Daily average wind speeds can exceed 6.5 m/s on occasion. Based on equations 1 and 2, this would result in a computed gauge collection of only one-fifth of the actual snowfall. This undercatch is a result of the fitting of equations 1 and 2 and limited observed data (Yang et al., 1998a) in the tails of the formulations. The gauge undercatch equation is a regression fit for 5 gauges, compared to the World Meteorological Organization Double Fence Intercomparison Reference (DFIR), based on 107 observations with an r^2 of 0.72 for snow, and on 75 observations with an r^2 of 0.59 for mixed precipitation (Yang et al., 1998a). Therefore, the maximum permissible wind speed was set at 6.5 m/s to consider the error associated with curve fitting. Also, it should be noted that the DFIR gauge has been shown to underestimate precipitation by up to 20%, compared to a gauge situated in bushes (Yang et al., 1993).

Trace Precipitation Events

The National Weather Service (NWS) classifies precipitation amounts less than 0.01 inches or 0.254 mm as trace events (NWS, 2003). Since the amount of precipitation that occurs during a trace event is uncertain, especially considering the undercatch potential, it was assumed that a daily trace event was 0.127 mm.

Snowpack Sublimation

Sublimation can be computed from mass and heat transfer equations that express the latent heat required for phase change and the energy available from the environment in terms of sensible heat. Sverdrup (1936) derived amount of water transported away from or towards the surface as a function of the eddy conductivity and the change in the specific humidity of air (q) with respect to the height above the surface (z). This mass transport (F_E with units of mass per time per area) can explain the vertical distribution and exchange of water vapour, and is given as:

$$F_E = \rho_a\, k_o^2\, \frac{U_a}{\ln\left(\frac{z_a + z_d}{z_o}\right)} \frac{q_b - q_o}{\ln\left(\frac{z_b + z_d}{z_o}\right)} \quad (3),$$

where ρ_a is the density of water vapour, k_o is the von Karman roughness coefficient, U_a is the wind velocity at height z_a, z_o is the roughness height, z_d is the zero-plane displacement, z_b is the measurement height of q, and q_o is the specific humidity at the surface of the snow. This vapour flux is a latent heat when F_E is multiplied by the latent heat of sublimation. The sensible heat transported towards the surface, defined as Q_H (energy per unit time per area), is a function of the specific heat capacity of air (c_p), the eddy conductivity under stable air, and the change in the potential temperature of air (θ) with respect to the height above the surface. It takes a similar form as the vapour flux, and considers the potential temperature gradient between the air moving over the snowpack and the snowpack surface. For this paper, it was assumed that the vapour pressure gradient can be used to estimate the sublimation rates.

Light (1941) assumed no zero-plane displacement, i.e., z_d was 0, and that all meteorological variables were measured at the same height above the surface (a). The mass transfer equation (3) was rewritten as a function of vapour pressures at height (e_b), at the surface (e_o), and the air pressure (P) as follows:

$$F_E = \frac{0.623\,\rho_a\,k_o^2}{P}\,\frac{U_a}{\ln(z_a / z_o)}\,\frac{e_b - e_o}{\ln(z_b / z_o)} \qquad (4).$$

To use the vapour flux to estimate sublimation (equation 4), the density of air was computed as a function of the air temperature, the von Karman constant was assigned a value of 0.40 (Oke, 1987), the air pressure was measured (in mb), and the roughness height was approximated from Fassnacht et al. (1999) at 0.5 cm using a relationship developed by Lettau (1969). Vapour pressure above the snow surface (e_b) was computed as a function of air temperature and humidity that are measured at the same height as the wind speed. The vapour pressure at the snow surface (e_o) was assumed to be the saturated vapour pressure at the air temperature. Equation (4) was used to estimate the hourly sublimation losses when the air temperature was at or colder than 0^{o}C.

Blowing Snow Transport

To estimate the amount of blowing snow, the 3-m wind speed was converted to the 10-m wind speed (U_{10}) using the logarithmic wind profile equation. The U_{10} wind speed was then compared to the threshold wind speed to commence blowing snow ($U_{10\text{-}threshold}$). For temperatures colder than 0^{o}C, Li and Pomeroy (1997) estimated $U_{10\text{-}threshold}$ as a function of temperature:

$$U_{10\,(threshold)} = 9.43 + 0.18\,T + 0.0033\,T^2 \qquad (5).$$

For temperatures between 0 and +5^{o}C, Li and Pomeroy (1997) estimated the threshold to be 9.9 m/s. The hourly quantity of blowing snow was computed from Pomeroy et al. (1991) as:

$$q_{TOTAL} = 0.00792\,U_{10}^{4.04} \qquad (6).$$

Climate Variables

Hourly and average daily meteorological data were used to compute the gauge undercatch. Sublimation and blowing transport were computed from hourly data. While only a portion of blowing snow is sublimated, any snow that is blown from the meteorological station is snow that is lost from that snowpack at that location. The net quantity of snow lost is thus a function of sublimation directly from the snowpack surface and blowing snow away, with the exception of snow blowing into the area. It is assumed that no snow is blown into the area. The net quantity of snow lost from the pack is limited to the corrected amount of precipitation.

The data were used in the computation were unedited (NCDC, 2003). To reduce potential measurement errors, the monthly average, standard deviation, maximum and minimum were computed and examined to identify erroneous measurements that were subsequently removed.

Extensive time periods of missing data can yield net underestimates (Ginn, 1980), especially for monthly and seasonal totals. The monthly data were examined and less than 3% were identified as missing. Thus it was assumed underestimation due to missing data was minimal. For cold environments, accurate humidity measurements are not possible below –30^{o}C, as the wet bulb thermometer does not function. When the dry bulb temperature approached –30^{o}C, wet bulb temperatures and the relative humidity were examined to identify error. This did not occur for the study sites examined in this paper.

Snowfall gauge undercatch is computed when the air temperature is colder than 0^{o}C, while the mixed precipitation formulation (equation 2) is used for air temperatures between 0 and 3^{o}C. Sublimation is only computed when the temperature is colder than 0^{o}C. Blowing snow occurs at temperature colder than 0^{o}C based the threshold wind speed as a function of air temperature in equation 5, or when greater than 9.9 m/s when the air temperature was between 0 and +5^{o}C.

STUDY SITE

Six meteorological stations across the coterminous U.S. that have different climate conditions were chosen for this study (Table 1). Hourly meteorological data for the winter (October–June) of three water years (2000–2002) were retrieved from the National Climatic Data Center online database (NCDC, 2003). Data were only available for water years 2001 and 2002 for the South Lake Tahoe CA and Rawlins WY stations. Since snow depth data were not available, the time series of temperature and precipitation data were examined for each year for each station to determine when snow started to accumulate and when it ablation was likely complete. These dates were rounded to the nearest month (Table 1). While the phase of precipitation was not known for the study sites, it was observed that only in the later winter months did precipitation occasionally occur at air temperatures warmer than 0^{o}C. This also determines the start and end of ablation.

Table 1. Summary of stations used in analyses, and the periods considered winter for each of the three water years of interest (2000–2002). All stations are located at an airport, with the exception of Stanley, which is located at the Ranger Station.

station	state	call sign	WBAN	elevation (m)	latitude (N)	longitude (W)	winter period 2000	2001	2002
South Lake Tahoe	CA	TVL	93230	1925	38E54'	120E	*no data*	Nov – Apr	Dec – Apr
Stanley	ID	SNT	4112	1980	44E10'	114E56'	Dec – Apr	Nov – Apr	Nov – Apr
Rawlins	WY	RWL	24057	2053	41E48'	107E12'	*no data*	Nov – Mar	Nov – Mar
Leadville	CO	LXV	93009	3029	39E14'	106E19'	Dec – Apr	Nov – Apr	Nov – Apr
Rhinelander	WI	RHI	4803	487	45E38'	89E28'	Dec – Feb	Nov – Mar	Dec – Apr
Syracuse	NY	SYR	14771	125	43E7'	76E6'	Jan – Feb	Dec – Mar	Dec – Feb

The average monthly precipitation, temperature, wind speed and vapour pressure for the winter months are summarized in Table 2, for the winter months given in Table 1. Rawlins is dry, while Syracuse is wet. The monthly variation in precipitation is small for Stanley, Rawlins and Leadville (10–15 mm), somewhat larger for Rhinelander (20 mm), and large for South Lake Tahoe (30.7 mm) and Syracuse (42.5 mm). Syracuse and South Lake Tahoe are warmer than the four other stations, while Stanley and Leadville saw larger temperature variations. Similarly Leadville was dry, having low humidity, while Syracuse and South Lake Tahoe were more humid. The variation in humidity was limited. South Lake Tahoe and Leadville have the largest average vapour pressure deficit, with Leadville having the largest variation. The vapour pressure deficit is almost half at Stanley and Rhinelander, with the smallest variation being observed at Rhinelander. The vapour pressure deficit at Rawlins and Syracuse are between the high and low values. Rawlins was the windiest location having the largest variation in wind speed, while wind speeds were low in Stanley and moderate to low in South Lake Tahoe. The variation in wind speed was similar for all stations except Rawlins.

Table 2. The average (mean) and coefficient of variation (COV) of the station meteorology for the winter periods listed in Table 1. Note: † precipitation is corrected using daily data

station	precipitation (mm)†		temperature (°C)		humidity (mb)		vapour pressure deficit (mb)		wind (m/s)	
	mean	COV	mean	COV	mean	COV	mean	COV	mean	COV
South Lake Tahoe	34.6	0.887	–0.44	–6.14	4.01	0.155	1.86	0.301	2.4	0.142
Stanley	31.4	0.439	–5.83	–0.81	3.25	0.246	1.04	0.577	1.3	0.269
Rawlins	19.2	0.552	–4.71	–0.62	3.12	0.218	1.36	0.294	5.6	0.170
Leadville	27.0	0.552	–5.24	–0.74	2.44	0.250	1.76	0.449	3.6	0.111
Rhinelander	26.7	0.749	–6.64	–0.51	3.19	0.251	0.97	0.278	3.4	0.097
Syracuse	89.9	0.473	–1.66	–1.69	4.32	0.183	1.48	0.250	4.3	0.086

RESULTS

For the winter period, the computed average monthly precipitation undercatch (hereinafter called undercatch), snowpack sublimation, and blowing snow transport are presented in Table 3, along with the corrected precipitation that considering wind-induced undercatch. At South Lake Tahoe, undercatch is moderate and comparable to the sum of sublimation and blowing snow transport (hereinafter called snowpack loss). Undercatch and snowpack loss are low at Stanley. At Rawlins, Leadville and Rhinelander undercatch is moderate, and it is high at Syracuse. Sublimation is moderate at Rawlins since blowing snow is substantial. Syracuse is the other location with substantial blowing snow. Sublimation averages 0.5 mm per day at Rhinelander, and is approximately 0.75 mm per day at Leadville and Syracuse.

For each month, the undercatch is plotted versus the snowpack loss in Figure 1. With the exception of two of the nine winter months at Syracuse (December 2000 and March 2001) and one of the 11 South Lake Tahoe months (December 2001), snowpack loss is either equal to or larger than undercatch. The snowpack loss and undercatch are plotted as a percentage of the corrected precipitation in Figure 2. These results are similar to Figure 1. With the exception of Stanley, both undercatch and snowpack losses are substantial with respect to the monthly precipitation.

To consider the relative quantity of undercatch, sublimation and blowing snow, the percentage of each with respect to the sum of the three components is plotted in Figure 3. Overall, monthly sublimation rates are the largest of the three, as also summarized in Table 3. Blowing snow is most important at Rawlins, while undercatch and sublimation are almost equally important at Syracuse. Blowing snow is rarely of consequence at Stanley.

Table 3. For the number of winter months in the three water years, the average corrected precipitation, catch ratio (as per equation 1 and 2), gauge undercatch for an Alter-shielded NWS precipitation gauge, snowpack sublimation, and blowing snow transport. The average month total snowpack losses from sublimation and blowing snow transport are also included.

station	no. of months	corrected precipitation (mm)	catch ratio (% of actual precipitation)	gauge undercatch (mm)	snowpack sublimation (mm)	blowing snow transport (mm)	sublimation + blowing snow (mm)
South Lake Tahoe CA	11	34.6	73.1	9.3	10.0	1.9	11.9
Stanley ID	17	31.4	91.4	2.7	7.1	0.1	7.2
Rawlins WY	10	19.2	46.9	10.2	7.9	11.3	19.2
Leadville CO	17	27.0	68.5	8.5	21.6	2.6	24.2
Rhinelander WI	12	26.7	69.3	8.2	14.2	1.9	16.1
Syracuse NY	9	89.9	70.2	26.8	22.7	12.6	35.3

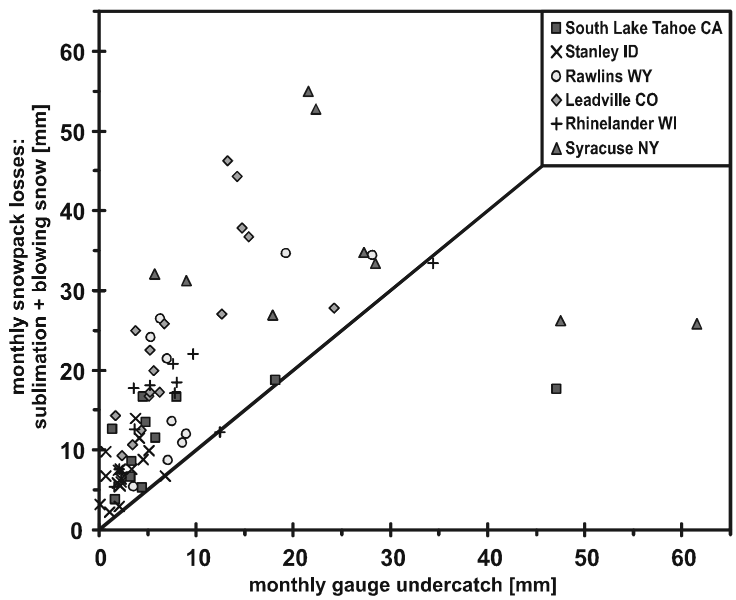

Figure 1. Monthly snowpack losses (sublimation and blowing snow) versus wind induced precipitation gauge undercatch for the six stations for winter months of water years 2000 to 2002

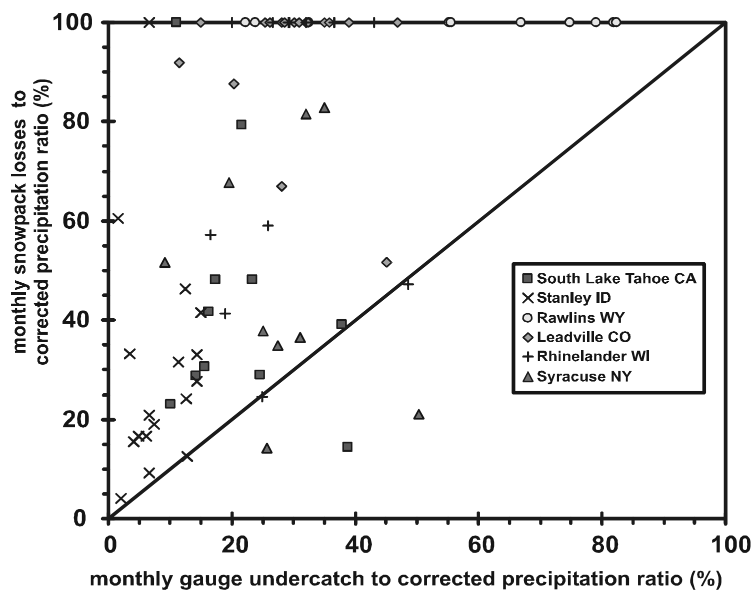

Figure 2. Monthly ratios with respect to corrected precipitation for snowpack losses (sublimation and blowing snow) versus wind induced precipitation gauge undercatch for the six stations for winter months of water years 2000 to 2002

The monthly data have been summed for each season at each site and these results are illustrated in Figure 4. For South Lake Tahoe and Stanley, seasonal snowpack losses are consistent, while undercatch varies. For the other stations, as seasonal snowpack losses increase, generally so does undercatch. As for the monthly estimates, snowpack losses are substantially larger than undercatch for Rawlins and Leadville.

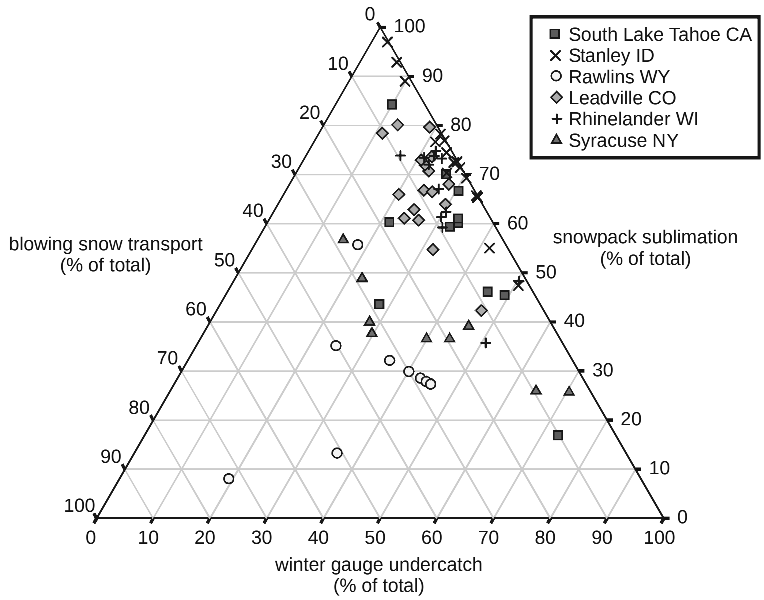

Figure 3. Distribution of monthly precipitation gauge undercatch, snowpack sublimation and blowing snow transport as a function of the sum of the three components for the three winters at the six study sites

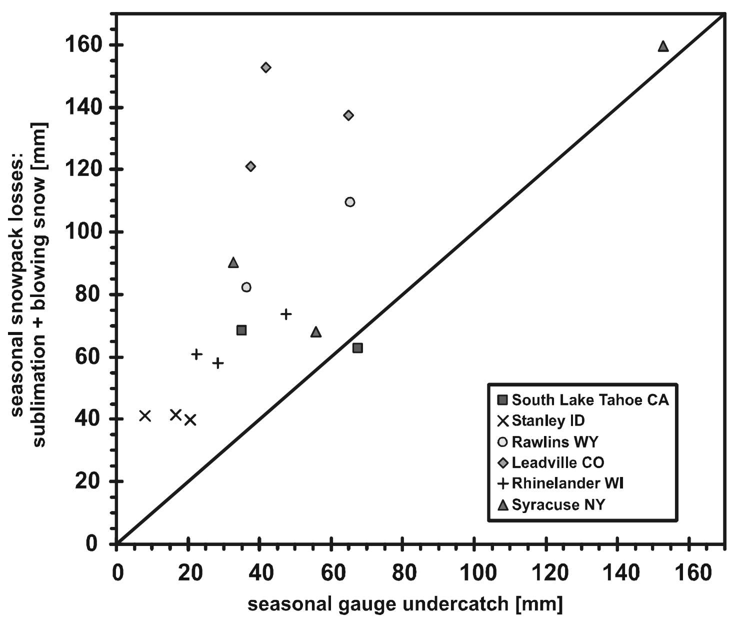

Figure 4. Winter season snowpack losses (sublimation and blowing snow) versus wind induced precipitation gauge undercatch for the six stations for entire cumulative winter months of water years 2000 to 2002

DISCUSSION

Precipitation gauge undercatch, snowpack sublimation and blowing snow transport are important to adjust estimates of snowpack accumulation, especially for monthly and seasonal water balance calculations. Since wind is the most important meteorological variable, undercatch and snowpack losses are small at Stanley (Table 2).

For the monthly comparison, there is no obvious pattern for the different sites. For very windy environments, i.e., Rawlins WY, computed snowpack sublimation and blowing snow transport can result in no net snowpack accumulation (Figure 2). Blowing snow is a major winter transportation problem in Wyoming, and estimated rates (Figure 3) are likely quite realistic. For other drier environments (precipitation and humidity), i.e., Leadville CO, computed sublimation is large (Figure 3). This potential exists, but actual sublimation rates can be significantly smaller as the sensible heat flux may not exist to fulfill the available latent heat flux. Therefore, undercatch estimates would more closely approximate snowpack losses for environments that are not consistently very windy, i.e., Wyoming. Where winds are not low (Stanley ID) or high (Rawlins WY) (Table 2), the average catch ratio is about 70% (Table 3) and undercatch is a function of the quantity of precipitation. Average winds over 3 m/s and more snowfall (Leadville CO, Rhinelander WI and Syracuse NY) yield high snowpack sublimation estimates. Average winds over 4 m/s are a result of high hourly winds (Rawlins WY and Syracuse NY) that yield significant blowing snow estimates (Table 3)

Seasonally for more snowy and wet environments, i.e., South Lake Tahoe CA and Syracuse NY, undercatch and snowpack losses roughly balance each other. The length of the winter and quantity of snowfall should also be considered.

There is more variability in undercatch than snowpack losses when monthly precipitation varies more than humidity or wind speed (coefficient of variation in Table 2). Undercatch is computed daily as a function of precipitation and wind speed (equation 1 and 2), while sublimation is computed hourly as a function of vapour pressure deficit and wind speed (equation 4) and blowing snow transport is computed solely as a function of hourly wind speed (equation 6). The longer daily time steps could dampen extremes in wind speed and reduce undercatch, but precipitation variability was much greater than wind speed variability. The wind speed during precipitation events is on average 10–20% stronger than the wind in general. The extreme of this is South Lake Tahoe CA, which was 50% stronger. The estimates also vary spatially, especially undercatch due to particle characteristics, gauge location, etc., (Yang et al., 1998a) and blowing snow due to particle characteristics and snowpack surface metamorphism (Li and Pomeroy, 1997).

The errors with estimating the three quantities come from the data, the equations, and the associated assumptions. The data were unedited, but attempts were made to identify erroneous values. Estimating the total precipitation during trace events was more suitable for daily time steps than hourly time steps, reducing but not eliminating the need to quantify the nature and magnitude of trace events. Trace events are less than 0.254 mm and contribute a small amount to the net snowfall, but are important for areas where they are persistent, such as Alaska.

Precipitation undercatch for six Alter-shield NWS gauges has been computed. Since gauge shape and shield configuration alter the aerodynamics around the mouth of a gauge, computation for other shield (and gauges) could yield different quantities of undercatch (see Goodison et al., 1998). This must be considered, especially for other countries that have other standards. Other mechanisms of undercatch, such as evaporative losses (sublimation in the winter), wetting, and bridging of accumulated snow across the mouth of the gauge, have not been considered in this paper. However, at the daily time step considered by the wind induced gauge undercatch equation, sublimation from snow inside the gauge is likely smaller than evaporative losses that are estimated at 0.15 mm (Goodison et al., 1998). Bridging occurs at low wind and, while some of the bridged snow can sublimate, the remaining snow will fall into the gauge or to the sides. Bridging may only be significant for wetter and warmer environments and it is likely small for a daily time step.

The sublimation estimates assume that the vapour gradient, i.e., the latent heat flux, is the controlling factor for sublimation. This is typically true for humid environments, but for less humid areas, especially colder areas, the sensible heat flux can be limiting. Over time, the two fluxes usually balance. Sublimation for the snowpack has been estimated by numerous researchers. However, while the work of Sverdrup (1936) set the stage to quantify sublimation rates, the estimation snowpack sublimation from meteorological data is not complete (Hood et al., 1999). The latent heat flux equation was used to estimate sublimation rates from the snowpack. For drier environments, such as Rawlins (Wyoming) and Leadville (Colorado), it is possible that the sensible heat flux is unavailable to sublimate the snow that vaporized based on the vapour pressure deficit. This would reduce the snowpack sublimation estimates.

Of the three computations considered in this paper, blowing snow may be the most difficult to quantify. The threshold for blowing snow is based on meteorological conditions and observations of blowing snow by the Meteorological Service of Canada data for several locations in Saskatchewan, Canada in mid-1970s. This relationship (equation 5) attempts to consider the variations in the threshold due to changes in the characteristics of the surface of the snowpack, which will enable easier movement for fresh snow and more difficult movement for older more dense snow. Once the 10-m wind speed exceeds the threshold wind speed, the equation for blowing snow transport (equation 6) was used. While this equation is solely a function of the 10-m wind speed, it attempts to consider various conditions, such as crystal characteristics, transportation in the form of saltation and suspension, and fetch direction of wind. These specifics are not known for the NWS gauges examined in this study. Blowing snow is not always a loss; it can accumulate after blowing from upwind. The computations may overestimate snow lost due to blowing snow transport. However, meteorological gauges are typically established in open areas, reducing the potential for accumulation on site. As well, once the distances over which snow is transported increase, sublimation of the transported snow become significant, i.e., much of the blowing snow is lost to the atmosphere.

Only the 3-m wind speeds are measured at the NWS sites, and these are used to estimate the 10-m wind speeds. This translation of wind speeds were computed using the logarithmic wind profile and an assumption of roughness lengths for snow, specially a z_{snow} of 0.01 m. This is based on Goodison et al. (1998), yet z_{snow} may vary. For example, Fassnacht et al. (1999) computed it to be 0.005 m based on Lettau (1969). This will effect both snowpack sublimation and blowing snow transport.

The computations performed in this paper are based on equations formulated from field observations from specific sites together with various assumptions. No observations were made of undercatch, sublimation or blowing snow at the six gauges. Field measurements would supplement the computations, especially since they vary with location and climate. For precipitation undercatch, gauge and shield types also are important. The sublimation and blowing snow equations are based on hourly meteorology while the undercatch computations are derived from daily data. The sensitivity of the different temporal resolution data should also be examined to consider evaluation of longer time periods.

The snowpack components estimated in this paper are for a point location. If the results are to be scaled, the extent of scaling will become an important factor. For example, Schroeter (1988) found that in Southern Ontario, at large enough modelling grid sizes, such as greater than 1 to 5 km, the net outflux of blowing snow is equal to the net influx of blowing snow. In that case, only the loss of blowing snow to sublimation is important, and this is a function of fetch length (Pomeroy et al., 1991). Various researchers have examined the influence of scale related to snow and snow cover (e.g., Woo, 1998; Derksen and LeDrew, 2000); however, this is beyond the scope of this paper.

CONCLUSIONS

Monthly and seasonal precipitation gauge snowfall undercatch, snowpack sublimation, and blowing snow transport losses are significant if average wind speeds are greater than 2 m/s. None were significant for the low wind Stanley, Idaho station.

There are no consistent monthly patterns for specific environments, but wind speed and precipitation quantities are the most important variables in estimating the three components. On a seasonal basis, undercatch and snowpack losses balance for more snowy and wet environments. There is often limited net snow accumulation in very windy environments.

The estimates computed in this paper are based in part on equation derived from specific locations. No dataset exists with measurements of all three components exist, thus this dataset is required at specific sites to further evaluate how they balance. As well, the magnitude and significance of trace precipitation events need to be measured as trace events can be occur often in some locations such as Alaska (e.g., Yang et al., 1998b).

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