Evaporation of intercepted snow: Analysis of governing factors

A. Lundberg

Division of Water Resources Engineering, Luleå University of Technology, Luleå, Sweden

S. Halldin

Department of Hydrology, Uppsala University, Uppsala, Sweden

Abstract. Insufficient understanding of winter hydrology conditions still hampers progress in predicting springtime discharge. The least known term in the winter water balance is evaporation, particularly of intercepted snow. Recent studies have shown that the evaporation from intercepted snow can be important. This paper elaborates factors governing evaporation of intercepted snow. Measurements with a cut tree–weighing device combined with a method to measure throughfall and drip gave a maximum evaporation rate of 0.3 ± 0.06 mm/h or 3.3 ± 0.06 mm/24 hours from a 6-m-high spruce. Calculations of evaporation with a combination equation and different ways to calculate the aerodynamic resistance and the evaporation from a snow-intercepted canopy during melt and sleet events showed that the most important factors for calculating the evaporation were the relative humidity, the aerodynamic resistance, the wind speed, and the intercepted mass. Less important factors were the energy to melt the intercepted snow, the method for calculating reduction in evaporation caused by a partly snow-covered canopy, accuracy in measurement of wind speed, air temperature, and net radiation.

Introduction

What happens to snow intercepted by forest canopies? Is it only a question of delay of the precipitation, or is a substantial amount lost by evaporation? Water regulation companies depend on a reliable answer to this question in order to correctly predict springtime discharge. The size of this discharge is also important in areas where snow from highlands provides water to drier lowlands. Timber harvesting in areas with substantial snow cover has been seen to increase snowmelt runoff [e.g., *Troendle and King*, 1985]. With the present trend in Europe to change agricultural land into forests, the hydrological effects of afforestation are becoming an important topic. Global climatic change, which might alter precipitation patterns, might also alter evaporation patterns since evaporation is larger from a wet than from a dry forest.

The first snow interception investigations indicated that the reduction of snowpack water equivalent found in forests compared to fields and small clearings was caused by the evaporation of intercepted snow [*Wilm and Dunford*, 1948; *Goodell*, 1959]. Many papers during 1950–1980 questioned the evaporation approach and stressed aerodynamic effects such as redistribution of intercepted snow and differences in deposition [e.g., *Hoover and Leaf*, 1967; *Gary*, 1974]. Measurements during recent years show that evaporation of intercepted snow can be important [*Schmidt et al.*, 1988; *Schmidt*, 1991; *Calder*, 1985, 1990; *Lundberg*, 1993].

Process studies become important when a substantial amount of the snow precipitation is lost by evaporation of intercepted snow. Recent studies of this type are presented by *Calder* [1990] and *Schmidt* [1991]. Schmidt treats subzero

events and calculates the sublimation as the product of the sublimation of a 1-mm ice sphere and a proportionality factor indicating the exposed snow area. Calder uses a combination equation, with different aerodynamic resistances for snow and liquid conditions, to calculate the evaporation.

The aim of this paper is to elaborate which factors govern evaporation of intercepted snow. Evaporation from intercepted snow was measured with a cut tree–weighing approach combined with a method to measure throughfall and drip from the tree. These measurements were analyzed with relation to the aerodynamic resistance, the amount of intercepted snow, and the accuracy of measured interception storage and weather variables.

Measurements

Weather variables as well as snow interception and evaporation from a sparse coniferous forest stand were measured during December to April 1990–1991 and 1991–1992.

Experimental Site

The experimental site was located 10 m above sea level in Luleå (65°37'N, 20°09'E), Sweden (Figure 1 (left)). The surrounding flat area consisted of a sparse stand of 5- to 6-m-high trees (average trunk diameter 10 cm and average projected radius of 0.8 m, stand density approximately 2000 stems per hectare, age around 30 years), with two stands with 12- to 15-m-high trees 80 m away to the northwest and 40 m away to the east (Figure 1 (middle)). The pine (*Pinus silvestris*) stand was intermixed with a few spruce (*Picea abies*) and birch trees (*Betula pubescens*).

Tree-Weighing Device

A device was constructed to measure intercepted mass and throughfall every 20 min. A brief description of the setup

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and evaporate the water. It was tilted in order to drain the water and equipped with a wind shield.

Theoretical Concepts

Evaporation From a Snow-Covered Stand: Basic Theory

The energy balance of a snow-covered forest stand can be written as

$$R_N = H + \lambda E + Q + M \quad (1)$$

Notation following standard conventions is listed at the end of the paper. This equation tells us that the available radiant energy at a reference height above the surface is shared between convective fluxes of sensible heat, H , and latent heat, λE , between the surface and the atmospheric boundary layer, time rate of change of heat content in the masses of air, biomass, snow, and soil below the reference height, Q , and a residual term, M . The M term is the sum of normally insignificant components of the balance, e.g., the photochemical energy storage. L.-C. Lundin et al. (Photochemical storage in the land surface energy balance, submitted to *Journal of Hydrology*, 1994) have elaborated on this term, which might be of significance during the growth season but not in the winter. For the analysis of snow interception, it is important to distinguish between the parts of the convective heat fluxes emanating from the snow cover on the ground (S) and from the snow covered stand canopy (I):

$$H + \lambda E = H_I + H_S + \lambda E_I + \lambda E_S \quad (2)$$

There is a considerable body of literature discussing when the total fluxes of sensible and latent heat primarily emanate from the canopy and only marginally from the ground. Lhomme [1991] discusses when it is possible to analyze a vegetation-covered ground with a single-level approach and when it is necessary to account for the various sources and sinks of the vegetation cover between the reference height and the ground. He proves that the vegetation-covered surface can be analyzed analogously to a physically well-defined surface as long as the convective fluxes from the ground are negligible. Shuttleworth and Wallace [1985] show that it is formally possible to establish a similar analogy even when a significant fraction of the fluxes comes from the ground. The analyses of both Lhomme [1991] and Shuttleworth and Wallace [1985], which are carried out as analogies with Ohm's law, rest on the assumption that mathematically precise definitions can be given to resistances against the convective fluxes at the canopy and ground surfaces. Shuttleworth and Wallace [1985] clarify that the analysis depends on the assumption that the aerodynamic mixing within the stand is sufficiently good to allow the hypothetical existence of a "mean canopy airstream" [Thom, 1972]. This condition is normally met by all sparse forest stands. Shuttleworth and Wallace [1985] deduce that the fraction of the total evaporation emanating from the canopy rapidly approaches 100% in cases where the surface resistance of the ground is equal to zero, as is the case for, e.g., a snow-covered ground. Already for a leaf area index of unity, the canopy evaporation fraction is close to 90%. No measurements were made of the leaf area index for the studied stand, but the age, average tree height, and the tree density marginally exceeded those of a young pine stand at Jädraås, central

Sweden, which was intensively studied during the 10-year-long interdisciplinary Swedish Coniferous Forest Project. For this reason it is reasonable to assume that the leaf area index of our stand was around or above two. In such a case, the snow cover contribution to total evaporation would be around 5% or less. The Shuttleworth and Wallace [1985] analysis is based on the assumption that around 30% of the radiant energy reaches the ground when the leaf area index is around two, which is highly unlikely for the stand used in our study. Net radiation was normally very small, and when beam radiation contributed to nonnegligible intensities, the direct sunshine, because of its low angle, primarily hit the canopy. For all these reasons we assumed that convective fluxes from the snow cover on the ground was negligible and that the canopy could be treated as a hypothetical "big leaf" surface with well-defined surface characteristics such as temperature, humidity, and aerodynamic roughness.

There are several terms relating the time rates of change in the sensible (s) and latent (l) heat storage of the various system compartments:

$$Q = Q_{gs} + Q_{gl} + Q_{ss} + Q_{sl} + Q_{is} + Q_{il} + Q_{vs} + Q_{as} + Q_{al} \quad (3)$$

Most of the sensible heat terms may safely be neglected since evaporation of intercepted snow normally occurs at or close to the freezing point when the time rate of change in temperature is minimal. The only sensible heat storage term which might be significant would be Q_{vs} , the one relating to aboveground tree biomass. An approximative calculation of this term, based on the hypothesis of a 5°C heating of the total stem volume in the studied stand, leads to a change in sensible biomass heat storage of $3.4 \times 10^4 \text{ J/m}^2$, roughly equivalent to an evaporation of 0.015 mm. Such heating could perhaps occur over the course of a day, so the corresponding rate would be minimal. It is, therefore, a good first-order approximation to neglect all sensible heat storage terms in the study of evaporation from intercepted snow.

It is more difficult to give orders of magnitude for the latent heat storage terms. Stewart and Thom [1973] present maximum values of the latent heat storage rate of change in the air below the reference height to be $+15/-7 \text{ W/m}^2$ for a reference height of 20.5 m. With our reference height, 11.5 m, this would correspond to $+8/-4 \text{ W/m}^2$. It is reasonable to decrease this maximum by 50–75% according to the largest amount of water vapor that the air can hold at the freezing point compared to the British summer air temperatures prevailing in the study of Stewart and Thom [1973]. With such a magnitude, it is reasonable to neglect the latent air heat term. The latent heat storage may be considerable in a snowpack of typically half a meter's depth. If one assumes, as done above, that evaporation from the snow cover is negligible and that almost no radiant energy is absorbed by it, it is likely that this latent term could be disregarded, in the first approximation, for situations when evaporation of intercepted snow is an important process. As long as there is a snow cover of significant depth, both latent and sensible heat changes in the ground are small, and their sum is negligible [Johnsson and Lundin, 1991]. The latent heat storage of the intercepted snow, on the other hand, is intimately related to the process of evaporation and cannot be disregarded a priori. This term can be given as

Figure 4. Quotient of actual evaporation E_A from a partly snow-covered canopy to evaporation E_I from a canopy covered with snow above its threshold value plotted against canopy storage C . The quotient is calculated in curve I from the method of *Rutter et al.* [1971, 1975] (equation (10)), in curve II from the method of *Hancock et al.* [1983] (equation (11)) under the assumption that $T = 0^\circ\text{C}$, and in curve III from a power relationship (equation (12)). The solid circles show measured intercepted rain evaporation rates E_A divided by calculated evaporation rate E_I [after *Calder and Wright*, 1986]. The evaporation is calculated with (9) using a constant $r_a = 3.5$ s/m.

$$Q_{il} = \frac{dmL}{A_t dt} \quad (4)$$

The application of (4) requires simplified assumptions to be made: The maximum energy to melt the intercepted snow, $Q_{il}(\text{max})$, can be calculated by assuming that all the intercepted mass is snow at the beginning of the period and that it is totally melted at the end of the period. The melt energy is exaggerated with this assumption since precipitation sometimes falls as sleet or rain and not all snow melts before it leaves the tree. The true value of energy for melt Q_{il} lies somewhere between zero and the calculated $Q_{il}(\text{max})$ value.

On the basis of the assumptions given above, the relevant form for the energy balance of a snow-covered forest stand is

$$R_N = H_I + \lambda E_I + Q_{il} \quad (5)$$

This energy balance must be complemented with equations governing the fluxes of latent and sensible heat from the snow-covered canopy:

$$H_I = \rho c_p \frac{T_c - T_a}{r_a} \quad (6)$$

$$\lambda E_I = \frac{\rho c_p}{\gamma} \frac{e_c - e_a}{r_a} \quad (7)$$

It is assumed in these equations, as shown to be reasonable by *Lhomme* [1991] and *Shuttleworth and Wallace* [1985], that it is possible to identify unique values of both temperature and humidity at the assumed canopy surface. It is also assumed that the resistances to fluxes of water vapor and sensible heat are the same, which inherently means that evaporation takes place at the surface of the canopy and is not subjected to stomatal or other restrictions. Under these circumstances it is possible to combine (5), (6), and (7), according to the method originally described by *Penman* [1948], in order to arrive at an equation for the evaporation, which depends only on air properties measured at the reference level:

$$\lambda E_I = \frac{\Delta(R_N - Q_{il}) + \rho c_p \frac{\delta e}{r_a}}{\Delta + \gamma} \quad (8)$$

In order to visualize the relative importance of the three terms in the numerator of (8) the equation may be rewritten as

$$\lambda E_I = \frac{\Delta R_N}{\Delta + \gamma} - \frac{\Delta Q_{il}}{\Delta + \gamma} + \frac{\rho c_p}{\Delta + \gamma} \frac{\delta e}{r_a} \quad (9)$$

The first term represents the part of evaporation driven by net radiation, the second term is similarly related to the heat required to melt the intercepted mass, and last is the ventilation term, which depends on the aerodynamic resistance, thus reflecting the efficiency of the turbulent transport.

Evaporation From a Partly Snow-Covered Canopy

There is no generally accepted theory for snow interception evaporation, whereas the theory of *Rutter et al.* [1971, 1975] has gained a fairly widespread acceptance for analysis of rain interception evaporation. This theory postulates that rain interception evaporation is reduced when a canopy is unsaturated. *Rutter et al.* [1971, 1975] assume that the reduction can be made in direct proportion to the relative size of wet and dry areas:

$$\begin{aligned} E_A &= E_I(C/S) & C < S \\ E_A &= E_I & C \geq S \end{aligned} \quad (10)$$

Hancock et al. [1983] criticize the above approach since it ignores the exchange of energy between the wet and dry patches via the canopy air space. Instead, they propose

$$\begin{aligned} E_A &= E_I \frac{(\Delta + \gamma)}{\Delta + (S/C)\gamma} & C < S \\ E_A &= E_I & C \geq S \end{aligned} \quad (11)$$

This equation is physically reasonable since it predicts a rise in the evaporation rate per unit wetted area for a drying surface (Figure 4). The evaporation values E_A measured by *Calder and Wright* [1986], however, show an E_A/E_I quotient for unsaturated canopies lower than that predicted from the *Rutter et al.* assumption. In order to investigate a dependence of this type, the data of *Calder and Wright* were approximated by a power function (Figure 4):

$$\begin{aligned} E_A &= E_I(C/S)^2 & C < S \\ E_A &= E_I & C \geq S \end{aligned} \quad (12)$$

The equations above, normally used to describe a reduction of rain interception evaporation, were tested as descriptions also of the reduction of snow interception evaporation. Because of the different maximal canopy storage capacities for rain and for snow, the concept of rain interception saturation was replaced by the concept of snow interception threshold. The threshold amount of intercepted snow, below which evaporation is reduced, was assumed equal to the rain saturation value S .

Schmidt [1991] studied evaporation of snow at subzero

temperatures. He found that the quotient between measured sublimated mass E_A from a 1-m-high, freely standing artificial tree and the calculated sublimated mass from a 1-mm ice sphere E_{SP} for new snow is roughly twice the value of that for old snow (Figure 5). The quotient increases up to a maximum value (at approximately twice the rain saturation value) and then decreases as the intercepted mass increases. This indicates that the evaporation rate decreases when the intercepted snow storage becomes considerably larger than the threshold value S . The greater sublimation rate from new snow is attributed to the fact that the exposed microscopic surface area for the same intercepted mass is greater for new snow than for old snow. The decrease in the quotient for large intercepted mass is attributed to the fact that the exposed macroscopic area of the snow decreases when bridging of snow between branches starts to occur. When the tree is completely covered with snow, the exposed macroscopic area is that of a cone. Schmidt expresses the empirically found quotient between the measured and calculated evaporation as a function of intercepted mass m and a constant c :

$$E_A = E_{SP} c (1000m)^B \quad B = (0.7 - 0.05m) \quad (13)$$

In order to make the expression independent of the tree size, the quotient is expressed here as a function of canopy storage:

$$C = m/A_t \quad (14)$$

Schmidt's [1991] formula was modified in this study by replacing the ice sphere evaporation with the combination equation value. The constant c was chosen so that the maximum quotient E_A/E_{SP} became unity:

$$E_A = E_I (0.0103) (320C)^B \quad B = (0.7 - 0.016C) \quad (15)$$

Aerodynamic Resistance

The aerodynamic resistance r_a is commonly calculated according to Monteith [1965]:

$$r_a = \frac{\left(\ln \frac{z-d}{z_0} \right)^2}{\kappa^2 u(z)} \quad (16)$$

Rutter *et al.* [1971] relate displacement height d and roughness length z_0 to stand height as $d = 0.75h$ and $z_0 = 0.1h$. Thom [1975] comments that (16) should be restricted to neutral conditions and suggests empirical stability functions

Figure 5. Quotient of measured sublimation E_A to calculated sublimation E_{SP} plotted against intercepted mass m and canopy storage C . The sublimation is measured from an artificial 1-m-high tree, and the calculated value equals the sublimation from a 1-mm ice sphere [after Schmidt, 1991]. The dashed line represents (15) suggested in this paper.

Figure 6. Estimated stand evaporation E as a function of z_0/h and d/h . The points show estimated values of z_0/h and d/h from studies of 14 coniferous forest canopies reviewed by Jarvis *et al.* [1976]. (After Gash *et al.* [1980].)

to compensate for the thermal buoyancy. Gash *et al.* [1980] examine the sensitivity of evaporation to the choice of the empirical relations between vegetation height h and displacement height d , on the one hand, and h and roughness length z_0 , on the other. It can be seen from the analysis of Gash (Figure 6) that the evaporation estimates are very sensitive to the choice of z_0 whereas the choice of d is less important. The data reviewed by Jarvis *et al.* [1976] indicate the uncertainty of the estimates and where the true values should lie (Figure 6).

Calder and Wright [1986] find no correlation between wind speed and aerodynamic resistance. They use a constant value of $r_a = 3.5$ s/m for rain and melted conditions. Their value is more than one third lower than the aerodynamic resistance for mean wind speed calculated with (16), which gives $r_a = 5.4$ s/m. Calder [1990] uses a constant aerodynamic resistance 10 times as large as the rain value for nonmelting snow conditions: $r_a = 35$ s/m. This value was found by optimization, and the arguments for such a high value are primarily that the surface of the snow is much smoother and occupies a much smaller total area than that of a rainy canopy. Calder further argues that atmospheric conditions generally are more stable and wind speeds lower for snow than for rain conditions. With respect to melting conditions he deduces the melt from the energy balance. When the maximum quotient of liquid to solid storage is exceeded, the excess melted mass is treated as rain.

Evaporation per Ground Area

The common practice in hydrology to express precipitation, evaporation, and canopy storage as depth (millimeters) in place of (the numerically identical) mass per ground area (kilograms per square meter) is adopted here, but it should be noticed that the evaporation values (millimeters) were defined as (measured mass)/(projected tree area).

Results

Losses of 0.3–3.3 mm/24 hours were found on roughly 50 days out of 250 during the two winter seasons. Roughly half of these occasions were associated with snowmelt or sleet events. This study was restricted to these occasions. This limitation was caused by the difficulty in ascertaining that no

Figure 7. Meteorological data, measured (E_M) and calculated (E_C) accumulated evaporation, canopy storage C , and energy terms in (9) on March 8 and March 22–23, 1991. (a) R_N (net radiation) and u , (wind speed). (b) T (air temperature), δe , (vapor deficit), and rh (relative humidity). (c) C and E_M . (d) P (precipitation flag, indicating occurrence of rainfall or snowfall). (e) E_C and E_M . (f) Term 1 (radiation term in (9)) and term 3 (ventilation term in (9)).

blowoff from the weighing system occurred with dry light snow in the tree. Occasions with wind velocities greater than 5 m/s were also excluded to make sure that all drip really hit the tray. Nineteen events remained for the analysis. Evaporation was calculated for these events with (8), but neglecting energy to melt the intercepted snow. The reduction of evaporation caused by a partly snow-covered canopy was calculated with the Rutter et al. approach, (equation (10) and Figure 4), and the aerodynamic resistance r_a was calculated with the standard equation (16). The displacement height d was assumed to be $0.75h$, and the roughness length z_0 to be $0.1h$. An evaporation reduction threshold value equal to a canopy rain saturation value of $S = 2$ mm was used. The equations and parameter settings above were used as a reference (REF) in the following calculations. During the two events shown in Figure 7 the net radiation was small (<200 W/m²) both days and the wind speed was less than 5 m/s (Figure 7a). The relative humidity was rather low for winter conditions ($\sim 80\%$), and the air temperature was below zero (Figure 7b). There was wet snow in the tree during both events (Figure 7c). On March 8, the total 1.4 mm of intercepted mass was evaporated. On March 22–23, 2.8

out of 3.8 mm intercepted snow evaporated during one precipitation-free period (Figures 7c and 7d). The calculated evaporation and measured evaporation during precipitation-free periods for the two days agree rather well (Figure 7e) and so does the accumulated evaporation for all 19 events (Figure 8a). The two presented events are representative for all 19 events with respect to the relative importance of the three terms (equation (9)) in the combination equation. The aerodynamic term was much greater than the net radiation term, and the net radiation term was often zero or slightly negative as on March 8 (Figure 7f). The water equivalent of the maximum energy to melt the intercepted snow was 0.1 mm on March 8 and 0.25 mm on March 22–23.

The measured evaporation values are subject to errors, and the attempts made to reduce and quantify these errors are discussed by Lundberg [1993], who recognizes the following errors: (1) temperature-induced change of the weight caused by temperature change of the balance; (2) desiccation of the tree; (3) wet snow stuck on the outside of the tray; (4) snow blown off the tree; (5) evaporation from the tray; (6) wind-induced weight changes; and (7) transpiration. The errors caused by temperature-induced weight

Figure 8. Calculated accumulated evaporation E_C plotted against measured accumulated evaporation E_M . For the calculations the REF equations (9), (10), and (16) with $d = 0.75h$, $z_0 = 0.1h$, and $S = 2$ mm are used. Two events lacked measurements of R_N , and R_N was set to zero for these events. (a) REF, the size of the maximum water equivalent to melt the intercepted snow, is indicated by the bars. (b) Relative error in E_C caused by error in rh : solid circles denote rh , inverted triangles denote $rh - 5\%$, and upward pointing triangles denote $rh + 5\%$ (but $rh \leq 100$). (c) Relative error in E_C caused by error in R_N : solid circles denote R_N , inverted triangles denote $R_N = 0$. (d) Relative error in E_C caused by uncertainties in C : solid circles denote C ; inverted triangles denote $C = C + 0.3$ mm, upward pointing triangles denote $C = C - 0.3$ mm. (e) Relative error in E_C caused by uncertainties in S : solid circles denote $S = 2$ mm, inverted triangles denote $S = 1$ mm, and upward pointing triangles denote $S = 3$ mm. (f) Data for r_a calculated with (16) using $d = 0.85h$ (inverted triangles), $d = 0.75h$ (solid circles), and $d = 0.65h$ (upward pointing triangles). (g) Data for r_a calculated with (16) using $z_0 = 0.15h$ (inverted triangles), $z_0 = 0.10h$ (solid circles), and $z_0 = 0.05h$ (upward pointing triangles). (h) Data for $r_a = 5$ s/m. (i) Data for $r_a = 10$ s/m. (j) Data for $r_a = 15$ s/m. (k) Value of $dT (= T_{10m} - T_{2m})$ is indicated by the size of the solid circles. (l) Events divided according to the size of canopy storage: inverted triangles denote canopy storage greater than S , open squares denote canopy storage less than S , solid circles denote canopy storage both greater and less than S . (m) E_A calculated from E_I with the *Hancock et al.* [1983] approach (equation (11)) when $C < S$. (n) E_A calculated from E_I with a power relationship (equation (12)) when $C < S$. (o) E_A calculated from E_I with a modified Schmidt approach (equation (15)): open circles denote new snowfall events, and solid circles denote old snowfall events.

changes of the balance as well as the desiccation of the tree were negligible during the studied events. The wet snow stuck on the outside of the tray was moved into the tray, and occasions with dry light snow or high wind velocities were omitted. The evaporation from the tray was small compared to the evaporation from the tree during all the studied events, and the transpiration is always small during midwinter besides its being hampered when there is snow or water on the tree. The average wind-induced change was small. The average error was approximately 0.1 kg ($= 0.06$ mm), so changes smaller than that could not be studied. This meant

that small changes during short time intervals could not be studied. Most values (11 events) reported here refer to 24-hour periods, whereas the remaining eight events occurred with precipitation during the day and vary in time from 5 to 17 hours.

The beginning and/or end of an event with high evaporation was, thus, sometimes determined by the start or end of precipitation events. The precipitation detector will be dry within 5 min after precipitation stops according to the manufacturer's specification. Observations were made of events with small amounts of dry, light snow during windy

Figure 9. Determination of canopy storage C ; $TW(t)$ is the weight of dry tree.

conditions, when the detector failed. The precipitation gauge measurements were used to confirm that there was no precipitation during an event and to help determine if the precipitation had fallen as snow, sleet, or rain.

The calculated evaporation values were affected by uncertainties in the weather variables, the measured canopy storage C , and the estimated canopy storage threshold S . They were also affected by the choice of equations to calculate the aerodynamic resistance and the reduction in evaporation when the intercepted mass was below the threshold value. The following calculations were made to investigate how the evaporation values were affected by the uncertainties in the variables and the choices of equations.

Changes in calculated evaporation caused by an error of 1°C in measured air temperature was small, approximately 5%. The error in the measured air temperatures could be expected to be smaller than 1°C , and the average temperature difference within the stand during the studied evaporation events was smaller than 1°C . The maximum observed temperature difference between the 10 m and the 2 m heights was 2.45°C , and the average difference was 0.61°C . An error in wind speed of 10% corresponded to an error in aerodynamic resistance of 10%. The uncertainties in the calculations of aerodynamic resistance were far greater than this. Neglecting the maximum energy to melt the intercepted snow did not affect the calculated values much (Figure 8a).

A slight trend with decreasing maximum values was observed for the relative humidity in 1991–1992. This was probably caused by corrosion in a connector. The values were corrected for this. Errors with this kind of sensor are encountered when measuring in very cold and humid environments (B. Norén, personal communication, 1991; C. Fahleson, personal communication, 1991). Condensation can take place in the sensing element or/and in the assembly (this does not change the calibration, and the probe works again after drying). Determination of the relative wintertime humidity with the Assman hygrometer (and with other methods depending on temperature differences between wet and dry thermometers) is difficult since an error in the temperature of 0.2°C causes errors in relative humidity of 4, 6, and 11% at air temperatures of 0° , -10° , and -20°C respectively [Fritschen and Lloyd, 1979]. Errors of 5% can be expected at temperatures with high evaporation rates. An error of 5% in the relative humidity caused errors in the calculated evaporations of up to 0.91 mm, or 36% of the total for individual events (Figure 8b). The greatest sensitivity to errors in relative humidity was found for the highest evaporation rates. At 0°C a change in air temperature of 1°C alters

the vapor pressure ($\Delta e \sim 50$ Pa) more than a change of 5% in the relative humidity ($\Delta e \sim 30$ Pa). Since the relative humidity was rather high ($\sim 85\%$) during the studied events, an error in the relative humidity of 5% would have caused an error in the saturation deficit δe of 30%, while a temperature change of 1°C would have caused an error of only 8%.

The net radiation could be expected to be accurate to within 3–4% or ± 10 W/m^2 (whichever is worst [Halldin and Lindroth, 1992]). Visual observations of the radiometer confirmed their finding that the ventilation arrangement was capable of keeping the radiometer functioning even during climatically severe situations. Calculations of evaporation for which the net radiation was neglected showed that the accuracy of the net radiation measurements was not critical (Figure 8c).

The canopy storage C was calculated as $C = (TW(t) + C(t)) - TW(t)$, where $(TW(t) + C(t))$ is the weight of the tree and the intercepted mass as registered by the balance and $TW(t)$ is the weight of the dry tree. Since the weight of the dry tree changed with time t because of desiccation, breaking of small twigs, and loss of needles (Figure 9), an uncertainty was introduced into the value of C .

The weight of the tree was determined by taking notes when visual inspection of the tree showed it to be dry. The weight values for the dry tree were then interpolated between these occasions. Visual inspection was a rather coarse tool and may have introduced an error in the determination of $TW(t)$ of 0.5 kg, equal to an error of 0.33 mm in the value of C . Analysis of this error showed that only evaporation values, where the canopy storage C was smaller than the threshold value S , were influenced by uncertainties in C (Figure 8d). Some calculated values with canopy storage in the vicinity of the threshold value were greatly influenced by errors in C . The largest error in evaporation caused by an error of 0.3 mm in C was 0.92 mm or 76% of the reference value. Calculated evaporation was moderately influenced when the canopy storage was well below S .

No observations from the weighed trees were available for determination of the threshold amount of intercepted snow below which evaporation was assumed to be reduced. A value of 2 mm was taken from literature on rain interception [Calder, 1990]. A sensitivity analysis performed with values of S varying from 1.0 to 3.0 mm (Figure 8e) showed that the S term influenced the calculated evaporation in roughly the same way as the C term.

The effect on calculated evaporation of possible ways to calculate the aerodynamic resistance was investigated. Calculations with different values of d and z_0 within the area of values indicated as reasonable by Jarvis *et al.* [1976] in Figure 6 showed that the estimates were very sensitive to the choice of roughness length and that changes in the calculated evaporation were approximately proportional to changes in roughness length (Table 1 and Figures 8f and 8g). The increased displacement heights d or the roughness lengths z_0 increased the calculated evaporation and vice versa as expected. There were no indications that any set of d and z_0 other than the reference would improve the agreement between calculations and measurements.

Calculations with a constant r_a (Figure 8h–8j) in accordance with Calder and Wright [1986] gave worse results than the REF calculation. The r_a value calculated with (16), using $d = 0.75h$ and $z_0 = 0.1h$, is $29.2/u$. The average value of the wind speed u during the studied events was approxi-

Table 1. Relation Between Measured Evaporation E_M and Evaporation E_C , Calculated With Different Assumptions About the Aerodynamic Resistance r_a and the Quotient Between Actual and Maximal Evaporation E_A/E_I

Values and Equations	Figure	r^2	Slope
REF	8a	0.58	1.003
<i>Assumption About r_a</i>			
$d = 0.85h$	8f	0.57	1.127
$d = 0.65h$	8f	0.59	0.913
$z_0 = 0.15h$	8g	0.55	1.548
$z_0 = 0.05h$	8g	0.64	0.547
$r_a = 5$	8h	0.36	2.112
$r_a = 10$	8i	0.38	1.011
$r_a = 15$	8j	0.39	0.646
<i>Assumption About E_A/E_I</i>			
E_A/E_I from (11) (Figure 4)	8m	0.61	1.016
E_A/E_I from (12) (Figure 4)	8n	0.49	0.983
E_A/E_I from (15) (Figure 5)	8o	0.59	0.910

mately 3 m/s, resulting in an r_a value of 10 s/m. A constant $r_a = 10$ s/m gave the right average evaporation but with large deviations for individual events (Table 1). The calculated evaporation was roughly proportional to the inverse of the selected r_a value. This also means that the calculated evaporation was roughly proportional to the wind velocity (equation (16)). The average wind speed during the studied events varied from 1.4 to 3.6 m/s.

Neutral stability is inherent in (16) for aerodynamic resistance. Stability determination would have required measurements that were not made in this study, but a hint of the stability of the air was given by the temperature difference between 2 and 10 m. When the temperature above the stand was lower than the temperature in the stand, the air could be assumed unstable. When the temperatures were roughly the same, neutral conditions could be assumed, and when the temperature increased with height, the air could be assumed stable. The largest measured temperature difference ($dT = T_{10m} - T_{2m}$) was 2.45°C, and the smallest was 0.07°C, which indicated a stable atmosphere during all events. There is a slight tendency (Figure 8k) for calculated evaporation to be overestimated when the temperature difference dT indicated stable conditions. Events with smaller temperature gradients often had evaporation estimates smaller than the measured values. This indicated that stability considerations might have improved the estimates.

Calder [1990] uses an aerodynamic resistance 10 times higher for snow than for liquid conditions. Determination of the transition from snow and sleet to liquid condition was not done in this study, but the size of the canopy storage was used as a coarse tool to distinguish between snow and liquid conditions. Snow conditions were assumed when canopy storage exceeded the rain saturation value during the whole event. There is a slight tendency (Figure 8l) for calculated evaporation to be overestimated when the canopy storage indicated that intercepted mass was mainly snow. Most evaporation events with low canopy storage were underestimated. It is possible, therefore, that roughness length decreases when a canopy is intercepted with snow. This strengthens the findings of Calder [1990] of an aerodynamic resistance larger for snow than for liquid conditions. It was not possible, however, from this study to draw any qualita-

tive conclusions about the quotient between the aerodynamic resistances in the two cases.

The different ways to calculate the reduction of evaporation caused by a partly snow-covered canopy (Figures 8m–8o and Table 1) show that the Hancock *et al.* [1983] relation (equation (11)) in Figure 8m gave a slightly better correlation between the measured and calculated values than the Rutter *et al.* approach (REF, Figure 8a), whereas the power relationship (equation (12)) gave a worse result (Figure 8n). The modified Schmidt [1991] approach (equation (15)), with decreasing evaporation velocity when canopy storage was well above the threshold value, gave roughly the same result as the REF calculations (Figure 8o). There were no indications that evaporation during events with new snow was underestimated or that it was overestimated during events with old snow.

Discussion

The analysis of our data rests heavily on the assumption that the studied tree is representative for the studied stand and that this stand can be treated as theoretically analogous to a “big leaf” surface, displaced a certain height above the ground and with properties reflecting the real stand. One prerequisite for this assumption to hold is that evaporation from the snow on the ground is negligible [Lhomme, 1991]. Experimental evidence of Doty and Johnston [1969] indicates that evaporation from the snow-covered ground in a coniferous forest is small. Bernier and Swanson [1993], on the other hand, report relatively high evaporation rates. The conditions prevailing in the study of Bernier and Swanson [1993] included much lower humidity values than in this study. They report daily average minimum values of relative humidity of 25% compared to 76% in this study. The analysis also rests on the findings of Shuttleworth and Wallace [1985], who illustrate their theory with purely numerical examples. Their predicted influence on evaporation of stand density is to some degree confirmed by the experimental findings of Teklehaimanot *et al.* [1991], who studied evaporation from canopies with different tree spacing. They found that the aerodynamic resistance of the stand depended on the ventilation and hence on the tree spacing. They studied evaporation during heavy rains with a traditional volume balance method to determine the evaporation and a truncated version (neglecting net radiation) of the combination equation to determine the boundary layer conductance (the reciprocal of the aerodynamic resistance). The boundary layer conductance per tree was found to increase as the spacing of the trees increased, but the boundary layer conductance per unit area decreased as the spacing of the trees increased.

Several factors turned out to be less important for the calculation of evaporation of intercepted snow, but care should be taken in extrapolating these findings. It does not seem necessary to include the energy to melt the intercepted snow in the calculations. The small changes observed in the estimates when net radiation was neglected cannot be extrapolated into areas where radiation plays a larger role than in the studied area. The observed small changes in the calculated evaporation caused by possible errors in temperature could be much greater if different aerodynamic resistances should be used for snow and liquid conditions. The fact that both very wet and fairly dry snow was used in the

analysis of the modified Schmidt equation may have masked any possible differences between old and new snow. The evaporation of intercepted snow depends on the frequency of precipitation and melt events. It is important that the results be interpreted in the context of the precipitation and temperature regime under which the observations were made. *Gash and Stewart* [1977] studied interception loss from forests, and they emphasize that extrapolation of results from individual studies must be limited to areas of similar precipitation climate.

Measurement of relative humidity is difficult at low temperatures, and the relative humidity is the most important factor, besides the aerodynamic resistance, when calculating evaporation of intercepted snow. This means that investigations of measurement techniques for relative humidity during winter conditions should have high priority.

The large changes in calculated evaporation for canopy storage values close to the threshold value are arguments for using an artificial tree instead of a real one to make sure that the canopy storage can be determined accurately. But the stiff branches of an artificial tree support more snow than natural branches of similar size [*Schmidt and Gluns*, 1991], and an artificial tree might introduce errors caused by nonrepresentative radiation conditions and heat characteristics.

The observed evaporation was normally much larger than the corresponding net radiation. Where does the energy to evaporate the intercepted snow come from in this case? It has been observed in several studies [*Calder*, 1985, 1986, 1990; *Stewart*, 1977] that evaporation from a rain-wetted canopy can be larger than the net radiation. *Morton* [1984, 1985] and *Calder* [1985, 1986] discuss the possibility that evaporation may exceed net radiation and the sources of energy to drive such high evaporation rates. *Morton* argues that the energy must be an edge effect operating on scales of the order of a few hundred meters, while *Calder* attributes the energy to large-scale advection of sensible heat. *Stewart* [1977] assumes that the additional energy must come from large-scale advection from areas upwind of a wet forest, where all net radiation is not used for evaporation. Studies of *de Bruin and Jacobs* [1989] show that evaporation from a small forest plot is influenced by the upwind conditions, but only to a small degree. They studied the micrometeorological behavior of a forest on a regional scale, i.e., 10–50 km with a one-dimensional planetary boundary layer (PBL) model coupled to a simple “big leaf” vegetation model and with the aerodynamic conductance corrected for stability. They suggest that the required energy (summertime) comes from entrainment of warm air from above the PBL.

Are there still doubts about the existence of high evaporation rates from intercepted snow? High evaporation rates of intercepted snow have been reported by several investigators [*Schmidt et al.*, 1988; *Schmidt*, 1991; *Calder*, 1985, 1990; *Lundberg*, 1993] in recent years. These investigators all use different methods, all of which have their individual drawbacks. The investigation by *Calder* [1985, 1990], where gamma ray attenuation was used to measure interception and heated plastic sheets were used to measure through-fall, may have exaggerated the evaporation caused by the heating of the sheets. *Schmidt et al.* [1988] report high evaporation rates during sublimation conditions. They use the cut-tree method combined with a very sensitive balance to measure sublimation events without drip. Some weight changes are

interpreted as events with blowoff, but events with very light snow might have been reported as evaporation. They were also using an isolated tree. The sublimation rate from an isolated tree differs from that of a tree in a forest. The method of *Lundberg* [1993], used in this study, cannot ascertain that no blowoff occurs, even if no traces of blowoff were observed at low and moderate wind speeds with wet snow. An investigation with a method to measure evaporation that is not associated with those weaknesses would be desirable. The eddy correlation method used successfully over both snow [*McKay and Thurtell*, 1978] and forests [*McNeil and Shuttleworth*, 1975; *Gash et al.*, 1989; *Högström et al.*, 1989] might be suitable for such an investigation.

Conclusions

Analysis of measured evaporation from intercepted snow showed that the most important factors for calculation of the evaporation were the relative humidity, the aerodynamic resistance, and the intercepted mass.

Fair agreement between measured and calculated evaporation showed that the process could be described with (1) a standard, one-dimensional combination equation for a snow-covered canopy, (2) a reduction in evaporation linearly related to a quotient between actual amount of intercepted snow and a threshold value, for interception amounts below this threshold, and (3) an aerodynamic resistance calculated with a roughness length of one tenth of the stand height. The calculations were very sensitive to errors in measured relative humidity, while the accuracies of wind speed, air temperature, and net radiation was not critical. Development of techniques for relative humidity measurement during winter conditions should thus have high priority.

Since the ventilation term was dominant in the equation for evaporation of intercepted snow, the evaporation values were very sensitive to the way the aerodynamic resistance is calculated. The calculated values were roughly proportional to the inverse of the chosen constant value for the aerodynamic resistance. With the standard equation for the aerodynamic resistance, the calculated values became roughly proportional to the roughness length. No improvement could be achieved by using any alternative relationships for the aerodynamic resistance. There were indications in this material that a larger aerodynamic resistance should be used for snow than for liquid conditions and that stability considerations would improve the calculations. Studies combining the gamma ray techniques [*Calder and Wright*, 1986] to measure the total intercepted mass and the microwave techniques [*Bouten et al.*, 1991] to measure the liquid intercepted mass would be desirable to confirm or refute the existence of different aerodynamic resistance for snow and liquid conditions.

Care should be taken when extrapolating results of this study into other geographical regions since the winter water balance will depend on both climate and vegetation characteristics. More research is needed concerning edge effects, the source of energy for evaporation of intercepted snow, and the aerodynamic resistance during intercepted snow conditions. Studies with eddy correlation techniques would be desirable to confirm the existence and occurrence of high interception evaporation rates.

Notation

A_t	projected tree area (m^2).
B	exponent in (13) (dimensionless).
C	equivalent depth of water (snow + water) held on canopy (mm).
c	constant in (13) (dimensionless).
c_p	specific heat of air ($\text{J kg}^{-1} \text{ } ^\circ\text{C}^{-1}$).
d	zero plane displacement height (m).
dt	time increment (s).
dm	melted intercepted mass (kg).
E	interception evaporation flux from a stand ($\text{kg m}^{-2} \text{ s}^{-1}$).
E_A	actual interception evaporation flux from a partly snow-covered/wet canopy ($\text{kg m}^{-2} \text{ s}^{-1}$).
E_I	evaporation flux from snow/rain intercepted in a canopy when the intercepted amount exceeds its threshold/saturation value ($\text{kg m}^{-2} \text{ s}^{-1}$).
E_S	evaporation flux from snow on the ground ($\text{kg m}^{-2} \text{ s}^{-1}$).
E_C	accumulated evaporation, calculated (mm).
E_M	accumulated evaporation, measured (mm).
E_{SP}	evaporation flux from a 1-mm ice sphere ($\text{kg m}^{-2} \text{ s}^{-1}$).
e_c	water vapor pressure at the canopy surface (Pa).
e_a	water vapor pressure at reference height z (Pa).
$e_s(T)$	saturation vapor pressure at temperature T , calculated as $0.61078 \exp [17.269T/(T + 273.30)]$ (Pa).
H	sensible heat flux from a snow/rain-intercepted stand (W m^{-2}).
H_I	sensible heat flux from a snow/rain-intercepted canopy (W m^{-2}).
H_S	sensible heat flux from snow-covered ground (W m^{-2}).
h	stand height (m).
L	latent heat of melt (J kg^{-1}).
M	residual term in energy balance, including normally negligible terms (W m^{-2}).
m	intercepted mass (kg).
P	occurrence of precipitation (dimensionless).
Q	time rate of change of heat storage below the reference height z (W m^{-2}).
Q_{gs}	time rate of change of sensible heat storage in the ground (W m^{-2}).
Q_{gl}	time rate of change of latent heat storage in the ground (W m^{-2}).
Q_{ss}	time rate of change of sensible heat storage in the snow on the ground (W m^{-2}).
Q_{sl}	time rate of change of latent heat storage in the snow on the ground (W m^{-2}).
Q_{is}	time rate of change of sensible heat storage in the intercepted snow (W m^{-2}).
Q_{il}	time rate of change of latent heat storage in the intercepted snow (W m^{-2}).
Q_v	time rate of change of sensible heat storage in the aboveground vegetation (W m^{-2}).
Q_{as}	time rate of change of sensible heat storage in the atmosphere below height z (W m^{-2}).
Q_{al}	time rate of change of latent heat storage in the atmosphere below height z (W m^{-2}).
R_N	net radiation at height z (W m^{-2}).

r_a	aerodynamic resistance to transport of vapor (s m^{-1}).
rh	relative humidity (dimensionless).
S	threshold value of snow or maximum depth of rain retained by a canopy (mm).
T_c	temperature at the canopy surface ($^\circ\text{C}$).
T_a	air temperature at reference height z ($^\circ\text{C}$).
TW	weight of dry tree (kg).
t	time (s).
$u(z)$	wind speed at height z (m s^{-1}).
z_0	roughness length (m).
z	reference height above surface (m).
δe	vapor pressure deficit ($e_s(T_a) - e_a$) at height z (Pa).
Δ	change of saturation vapor pressure with temperature ($\partial e_s / \partial T$) ($\text{Pa } ^\circ\text{C}^{-1}$).
γ	psychrometric "constant" ($\text{Pa } ^\circ\text{C}^{-1}$).
λ	latent heat of vaporization (J kg^{-1}).
κ	von Karman's constant (taken as 0.41) (dimensionless).
ρ	density of air saturated with water vapor at T (kg m^{-3}).

The physical properties c_p , γ , λ , and ρ , which change less than 1% per degree Celsius, were estimated from standard tables.

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S. Halldin, Department of Hydrology, Uppsala University, S-75309, Uppsala, Sweden.

A. Lundberg, Division of Water Resources Engineering, Luleå University of Technology, S-971 87 Luleå, Sweden.

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