Modeling snow accumulation and ablation processes in forested environments

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Abstract. The effects of forest canopies on snow accumulation and ablation processes can be very important for the hydrology of mid- and north-latitude basins. A mass and energy balance model for snow accumulation and ablation processes in forested environments was developed utilizing extensive measurements of snow interception and release in a maritime climate mountainous site in Oregon. The model, which was calibrated against one year of weighing lysimeter data and tested at the same site against measurements from the next year was able to reproduce the SWE evolution throughout both winters beneath the canopy as well as the nearby clearing, with correlations ranging from 0.87 to 0.99. Additionally, the model was evaluated using measurements from the BOREAS field campaign in Canada, without any calibration to test the model robustness in a boreal climate given the effects of micro-meteorology on snow interception. Simulated SWE was relatively close to the observations for the forested sites, while simulated snow depth was underestimated during the accumulation period at the forested sites but simulated fairly accurately during ablation.
1. Introduction

Snow is an important part of the hydrologic cycle, especially in high latitude and high elevation river basins. The contrast between snow presence and absence strongly affects the surface energy (e.g. albedo) and water (e.g. storage and streamflow) balances. Where forest cover is present, it alters snow accumulation and ablation processes, mostly by intercepting snowfall and modifying the surface micrometeorology (incoming radiation and wind speed) respectively. Intercepted snow can account for as much as 60% of annual snowfall in both boreal and maritime forests \cite{Storck2002}, while losses to sublimation can reach 30-40% of annual snowfall in coniferous canopies \cite{Pomeroy1993}.

Although the importance of snow interception and sublimation processes has been recognized, their incorporation in hydrologic models as well as their representation in land surface schemes used in numerical weather and climate prediction models has been limited \cite{Pomeroy1998b,Essery1998}. While a number of studies have examined snow interception processes in boreal forests \cite{Claasen1995,Harding1996,Hedstrom1998,Nakai1999,Pomeroy2002,Gusev2003}, few of these studies are applicable to maritime climates. Snow interception can be quite different in maritime and continental climates, mostly because of the dominance of micro-meteorology over canopy morphology in controlling snow interception \cite{Satterlund1970,Schmidt1991}.

Miller \cite{Miller1964} hypothesized that three factors control snow interception: canopy morphology, air temperature and wind speed. Simple interception models were developed for
individual snow storms by Satterlund and Haupt [1967], which demonstrated the different mechanisms controlling rain and snow interception. Schmidt and Gluns [1991] found low interception efficiency for light and heavy snow loadings on the canopy, with an increase in interception efficiency as snowfall increases due to cohesion of snow particles to intercepted snow and increased effective projected area of the canopy. In contrast, Calder [1990] and Harestad and Bunnell [1981] found a decrease in interception efficiency as snowfall increased. The contrast in the magnitude of observed snow interception between maritime and continental climates suggests the importance of micrometeorological conditions [Bunnell et al., 1985]. Schmidt and Gluns [1991] found that snow interception was relatively similar between three branch species at two sites, while Satterlund and Haupt [1970] found that there was no significant differences in the amount of snow intercepted by two tree species which had considerable morphological differences. The effects of air temperature on snow interception were found to be more pronounced as the canopy collection area became narrower with minimum interception at low air temperatures [Ohta et al., 1993].

Intercepted snow can be removed from the canopy by sublimation, mass release, or melt water drip. Sublimation from snowpacks has been studied using tree-weighing techniques [Schmidt, 1991; Lundberg, 1993; Montesi et al., 2004; Molotch et al., 2007]. Lundberg and Halldin [2001] found sublimation rates from intercepted snow reaching 1.3-3.9 mm/day. High rates of sublimation from snow intercepted by forest canopies can be sustained by both net radiation and sensible heat flux, and can be well predicted by a simple energy balance model if the canopy aerodynamic resistance is adjusted for the presence of snow cover. Mass release of intercepted snow occurs due to either mechanical wind effects or
melt. Both of these mechanisms are governed by the adhesion of intercepted snow to the tree branches. As the adhesion becomes stronger (i.e. during snowfall or at temperatures just below freezing) removal of snow due to wind becomes rare. On the other hand, as intercepted snow melts it destroys the bonds between the snow and the canopy facilitating mass release. Meltwater drip was measured by Kittredge [1953] over a period of 5 years in the Sierra Nevada; during 57 snowfalls only 4 produced more than 2 mm of meltwater drip, with an average of 0.8 mm out of an average total snowfall of 40 mm.

Most land surface models do not separate canopy snow from ground snow processes [Pomeroy et al., 1998a]. Verseghy et al. [1993] developed a new snow interception algorithm for the Canadian Land Surface Scheme which controlled the interception efficiency by canopy morphology, with a maximum threshold. Hedstrom and Pomeroy [1998] developed an interception/unloading model by assuming an exponential decay with increasing snowfall. A model of canopy snow interception, sublimation and melt was incorporated into a GCM land scheme by Essery et al. [2003], which they found improved its performance in off-line simulations. A similar model was developed by Niu and Yang [2004] to represent the effects of canopies on the surface snow mass and energy balance, which they found improved the estimation of snow albedo.

In this paper we describe a model of snowpack dynamics in forested environments, based on observations of snow accumulation and ablation at a mountain maritime site [Storck et al., 2002]. The model is evaluated with observations of snow water equivalent and depth from sites in both maritime and boreal climates. These observations are described in Section 2, while the model governing equations and validation results are presented in Sections 3 and 4 respectively.
2. Study sites and measurements

The observations sites are in the Umpqua National Forest, OR where observations were made by the second author during the winters of 1996-1997 and 1997-1998 [Storck et al., 2002], and three sites in the boreal forest of Saskatchewan (first two sites) and Manitoba (second site), respectively. Observations at the boreal forest sites were taken as part of the The Boreal Ecosystem-Atmosphere Study (BOREAS) conducted from 1993-96 [Sellers et al., 1997].

2.1. Umpqua, Oregon

The objective of observations made at the Umpqua National Forest field site was to observe the processes governing snow interception by forest canopies and beneath-canopy snow accumulation and ablation in a mountainous maritime climate, with the ultimate aim of understanding the role of rain-on-snow melt processes on flooding. The field campaign was part of the Demonstration of Ecosystem Management Options (DEMO) experiment [Aubry et al., 1999]. Annual precipitation at the field site is about 2 m, most of it in winter, with an annual maximum snow water equivalent (SWE) of about 350 mm in clearings. Mid-winter melt is common and final melt occurs in late April or early May. Frequent rain-on-snow events occur at this site throughout the winter, while spring melt is radiation-dominated.

Four weighing lysimeters were used to measure ground snowpack accumulation and melt; two of which were beneath a mature canopy (mostly Douglas fir), and the other two in clear-cut and shelterwood (partially harvested) sites, respectively. The sites were all located within 3 km of each other with no significant differences in topography. Differences in SWE between lysimeters beneath the forest canopy and the adjacent clearing were used
to infer snow interception. Additional measurements included precipitation (taken using two tipping bucket gauges), wind speed, incoming shortwave and longwave radiation, air temperature, and relative humidity at 2-m above the soil surface. The field observations are described in detail by Storck et al. [2002].

2.2. BOREAS

The objective of the BOREAS field program was to improve the understanding of the dynamics that govern mass and energy transfer between the boreal forest and the lower atmosphere [Sellers et al., 1995]. The BOREAS study region covered most of Saskatchewan and Manitoba, with individual measurement sites located within the northern and southern study areas (NSA and SSA respectively) [Sellers et al., 1997]. Topographic relief is small in both sites, but land cover is non-uniform within each study area and contains open areas and forests of different canopy types. Meteorological measurements, including below and above canopy air temperature, incoming shortwave and longwave radiation, relative humidity, wind speed, precipitation were taken by Automated Meteorological Stations (AMS) [Osborne et al., 1998] every 15 minutes. Additionally, the AMSs recorded snow depth and canopy temperature at the same temporal resolution, while bi-weekly manual snow depth and water equivalent measurements were taken beneath a range of canopy types during the same period. Two sites in the SSA and two in the NSA were selected for this study; these sites were covered mostly with mature jack pine (SSA-OJP and NSA-OJP), aspen (SSA-OA), and mixed spruce/poplar (NSA-YTH) trees.
3. Model formulation

Model formulation of the canopy interception processes relies heavily on field measurements described by Storck et al. [2002]. The main processes represented in the model are shown schematically in Fig. 1. The snow processes model has been incorporated into the macroscale Variable Infiltration Capacity (VIC) hydrologic model, which essentially solves an energy and mass balance over a gridded domain [Liang et al., 1994]. The spatial resolution for macroscale models usually ranges from 10 to 100 km (grid spacing), which are larger than characteristic scales of the modeled processes. Therefore, subgrid variability in topography, land cover and precipitation is modeled by a mosaic-type representation, wherein each grid cell is partitioned into elevation bands each of which containing a number of land cover tiles. The snow model is then applied to each land cover/elevation tile separately, and the simulated energy and mass fluxes and state variables for each grid cell are calculated as the area-averages of the tiles. Downward energy and moisture fluxes are required to drive the model, these include precipitation, air temperature, wind speed, downward shortwave and longwave radiation, and humidity. Alternatively, the last three terms can be estimated from the maximum and minimum daily air temperature and precipitation according to using algorithms described in Thornton and Running [1999] and Kimball et al. [1997], respectively.

3.1. Ground snowpack accumulation and ablation

The model represents the snowpack as a two-layer medium (a thin surface, and a thick deeper layer), and solves an energy and mass balance for the ground surface snowpack in a manner similar to other cold land processes models [Anderson, 1976; Wigmosta et al., 1994; Tarboton et al., 1995]. Energy exchange between the atmosphere, forest canopy and
snowpack occurs only with the surface layer. The energy balance of the surface layer is

\[ \rho_w c_s \frac{dW T_s}{d t} = Q_r + Q_s + Q_l + Q_p + Q_m \]  \hspace{1cm} (1)

where \( c_s \) is the specific heat of ice, \( \rho_w \) is the density of water, \( W \) is the water equivalent and \( T_s \) is the temperature of the surface layer, \( Q_r \) is the net radiation flux, \( Q_s \) is the sensible heat flux, \( Q_l \) is the latent heat flux, \( Q_p \) is the energy flux advected to the snowpack by rain or snow, and \( Q_m \) is the energy flux given to the pack due to liquid water refreezing or removed from the pack during melt. Eq. 1 is solved via a forward finite difference scheme over the model time step (\( \Delta t \)):

\[ W^{t+\Delta t} T_s^{t+\Delta t} - W^{t} T_s^{t} = \frac{\Delta t}{\rho_w c_s} (Q_r + Q_s + Q_l + Q_p + Q_m) \]  \hspace{1cm} (2)

Net radiation at the snow surface is either measured or calculated given incoming short-wave and longwave radiation as

\[ Q_r = L_i + S_i (1 - \alpha) - \sigma T_s^4 \]  \hspace{1cm} (3)

where \( L_i \) and \( S_i \) are incoming long and shortwave radiation, and \( \alpha \) is the snow surface albedo. The flux of sensible heat to the snowpack is given by

\[ Q_s = \rho c_p \frac{(T_a - T_s)}{r_{a,s}} \]  \hspace{1cm} (4)

where \( \rho \) is the air density, \( c_p \) is the specific heat of air, \( T_a \) is the air temperature, and \( r_{a,s} \) is the aerodynamic resistance between the snow surface and the near-surface reference height, given by

\[ r_{a,s} = \ln \left( \frac{z - d_s}{z_0} \right) \frac{2}{k^2 U_z} \]  \hspace{1cm} (5)

where \( k \) is von Karman’s constant, \( z_0 \) is the snow surface roughness, \( d_s \) is the snow depth, and \( U_z \) is the wind speed at the near-surface reference height \( z \). Similarly, the flux of
latent heat to the snow surface is given by

\[ Q_e = \lambda_i \rho \frac{0.622 e(T_a) - e_s(T_s)}{P_a} r_{a,s} \]  

(6)

where \( \lambda_i \) is the latent heat of vaporization when liquid water is present in the surface layer and the latent heat of sublimation in the absence of it, \( P_a \) is the atmospheric pressure, and \( e \) and \( e_s \) are the vapor and saturation vapor pressure respectively. Advected energy to the snowpack via rain or snow is given by

\[ Q_p = \rho_w c_w T_a \frac{P_r + P_s}{\Delta t} \]  

(7)

where \( c_w \) is the specific heat of water, \( P_r \) is the depth of rainfall, and \( P_s \) is the water equivalent of snowfall. Precipitation is partitioned into snowfall and rainfall based on a temperature threshold

\[
P_s = P \quad T_a \leq T_{min}
\]

\[
P_s = \frac{T_{max} - T_a}{T_{max} - T_{min}} P \quad T_{min} < T_a < T_{max}
\]

\[
P_s = 0 \quad T_a \geq T_{max}
\]

(8)

The total energy available for refreezing liquid water or melting the snowpack over a given time step depends on the net energy exchange at the snow surface

\[ Q_{net} = (Q_r + Q_s + Q_e + Q_p) \Delta t \]  

(9)

If \( Q_{net} \) is negative, then energy is being lost by the pack, and liquid water (if present) is refrozen. If \( Q_{net} \) is sufficiently negative to refreeze all liquid water, then the pack may cool. If \( Q_{net} \) is positive, then the excess energy available after the cold content has been
satisfied, produces snowmelt.

\[
Q_m \Delta t = \min \left( -Q_{net}, \rho_w \lambda_f W_{liq} \right) \quad Q_{net} < 0
\]

\[
Q_m \Delta t = - \left( Q_{net} + c_s W_{ice} T_s^l \right) \quad Q_{net} \geq 0
\]

The mass balance of the surface layer is given by

\[
\Delta W_{liq} = P_r + \frac{Q_l}{\rho_w \lambda_v} - \frac{Q_m}{\rho_w \lambda_f}
\]

\[
\Delta W_{ice} = P_s + \frac{Q_l}{\rho_w \lambda_s} + \frac{Q_m}{\rho_w \lambda_f}
\]

where \(Q_e\) exchanges water with the liquid phase if liquid water is present and \(Q_e\) exchanges water with the ice phase in the absence of liquid water.

If \(W_{ice}\) exceeds the maximum thickness of the surface layer (typically taken as 0.10 m of SWE), then the excess, along with its cold content, is distributed to the deeper (pack) layer. Similarly, if \(W_{liq}\) exceeds the liquid water holding capacity of the surface layer, the excess is drained to the pack layer. If the temperature of the pack layer is below freezing then liquid water transferred from the surface layer can refreeze. Liquid water remaining in the pack above its holding capacity is immediately routed to the soil as snowpack outflow. The dynamics of liquid water routing through the snowpack are not considered in this model because of the relatively coarse temporal and spatial resolutions of the model (typically 1 to 3 hours and 50-100 km\(^2\) respectively) [Lundquist and Dettinger, 2005].

As snow accumulates on the ground it goes through a metamorphism process, which causes the snowpack to compact and increase its density over time (except for depth hoar). In addition to the change in density caused by metamorphism, gravitational settling caused by newly fallen snow also contributes to the densification process. Following a similar approach to Anderson [1976], compaction is calculated as the sum of two frac-
tional compaction rates representing compaction due to metamorphism and overburden respectively

\[
\frac{\Delta \rho_s}{\Delta t} = (CR_m + CR_o) \rho_s
\] (12)

where \( \rho_s \) is the snow density, and \( CR_m, CR_o \) are the compaction rates due to metamorphism and overburden respectively. Metamorphism is important for newer snow, and the following empirical function is used

\[
CR_m = 2.778 \times 10^{-6} c_3 c_4 e^{-0.04(273.15-T_s)}
\] (13)

\[
c_3 = c_4 = 1 \quad \rho_l = 0, \rho_s \leq 150 \text{ km/m}^3
\]

\[
c_3 = e^{-0.046(\rho_s-150)} \quad \rho_s > 150 \text{ kg/m}^3
\]

\[
c_4 = 2 \quad \rho_l > 0
\]

where \( T_s \) is the snowpack temperature, and \( \rho_l \) is the bulk density of the liquid water in the pack. After the initial settling stage, the densification rate is controlled by the overburden snow, and the corresponding compaction rate can be estimated by

\[
CR_o = \frac{P_s}{\eta_0} e^{-c_5(273.15-T_s)} e^{-c_6 \rho_s}
\] (14)

where \( \eta_0 = 3.6 \times 10^6 \text{ N s/m}^2 \) is the viscosity coefficient at 0 °C, \( c_5 = 0.08 \text{ K}^{-1} \), \( c_6 = 0.021 \text{ m}^3/\text{kg} \), and \( P_s \) is the load pressure. Snowpacks are naturally layered media, therefore the load pressure would be different for each layer of the pack corresponding to different compaction rates. The model represents that “internal” compaction as an effective load pressure, i.e.

\[
P_s = \frac{1}{2} g \rho_w (W_{ns} + f W_s)
\] (15)

where \( g \) is the acceleration of gravity, \( W_{ns}, W_s \) is the amount of newly fallen snow and snow on the ground (in water equivalent units) respectively, and \( f \) is the internal com-
paction rate coefficient taken as 0.6 after calibration to measurements from the Cold Land Processes Experiment in Fraser Park, Colorado [Andreadis et al., 2008].

Snow albedo is assumed to decay with age, based on observations from Storck et al. [2002]:

\[
\alpha_a = 0.85 \lambda_a^{0.58} \\
\alpha_m = 0.85 \lambda_m^{0.46}
\]

(16)
(17)

where \( \alpha_a, \alpha_m \) are the albedo during the accumulation and melting seasons, \( t \) is the time since the last snowfall (in days), \( \lambda_a = 0.92 \), and \( \lambda_m = 0.70 \). Accumulation and melt seasons are defined based on the absence and presence of liquid water in the snow surface layer respectively.

3.2. Atmospheric stability

The calculation of turbulent energy exchange (Eqs 4, 5, 6) is complicated by the stability of the atmospheric boundary layer. During snowmelt, the atmosphere immediately above the snow surface is typically warmer. As parcels of cooler air near the snow surface are transported upward by turbulent eddies they tend to sink back toward the surface turbulent exchange is suppressed. In the presence of a snow cover, aerodynamic resistance is typically corrected for atmospheric stability according to the bulk Richardson’s number (\( Ri_b \)). The latter is a dimensionless ratio relating the buoyant and mechanical forces (i.e. turbulent eddies) acting on a parcel of air [Anderson, 1976]

\[
Ri_b = \frac{gz_a(T_a - T_s)}{0.5(T_a + T_s)U(z_a)^2}
\]

(18)
with the correction for stable conditions given as

\[
    r_{as} = \frac{r_{as}}{\left(1 - \frac{Ri_b}{Ri_{cr}}\right)^2} \quad 0 \leq Ri_b < Ri_{cr}
\]

and in unstable conditions as

\[
    r_{as} = \frac{r_{as}}{\left(1 - 16Ri_b\right)^{0.5}} \quad Ri_b < 0
\]

where \(Ri_{cr}\) is the critical value of the Richardson’s number (commonly taken as 0.2).

While the bulk Richardson’s number correction has the advantage of being straightforward to calculate based on observations at only one level above the snow surface, previous investigators have noted that its use results in no turbulent exchange under common melt conditions and leads to an underestimation of the latent and sensible heat fluxes to the snowpack (e.g. Jordan 1991; Tarboton et al. 1995).

An alternative formulation for the stability correction (adopted by Marks et al. 1998) is based on flux-profile relationships in which the vertical near-surface profiles of wind and potential temperature are assumed to be log-linear under stable conditions [Webb, 1970]. In this case, the effect of atmospheric stability is described by the Monin-Obukhov mixing length (\(L\))

\[
    L = \frac{u^3_a \rho}{kg \left(\frac{H}{Ta c_p}\right)}
\]

The friction velocity (\(u_a\)) and the sensible heat flux (\(H\)) are give by

\[
    u_a = \frac{u_a k}{\ln \left(\frac{z_a}{z_0}\right) - \Psi \left(\frac{z_a}{L}\right)}
\]

\[
    H = \frac{(T_a - T_s) k u_a \rho c_p}{\ln \left(\frac{z_a}{z_0}\right) - \Psi \left(\frac{z_a}{L}\right)}
\]
The $\Psi$ functions are given for stable conditions ($z_a/L > 0$) as

$$\Psi \left( \frac{z_a}{L} \right) = -5 \frac{z_a}{L}, \quad 0 \leq \frac{z_a}{L} \leq 1$$
$$\Psi \left( \frac{z_a}{L} \right) = -5, \quad \frac{z_a}{L} \geq 1 \quad (24)$$

The stability correction (Eqs 21-24) does not force the sensible heat flux to zero. Unfortunately solution of these equations requires an iterative procedure which is computationally too burdensome for a large-scale, spatially distributed model.

Therefore, a $Ri_b$ formulation that does not entirely suppress turbulent exchange under stable conditions was developed. Similar to the limit on $\Psi$ imposed by Eq. 24, an upper limit can be placed on $Ri_b$ at which $z/L$ is equal to unity. Combining Eqs 21 to 23 into one expression for $L$ yields the following expression for the bulk Richardson’s number when $z/L$ is equal to 1

$$Ri_u = \frac{1}{\ln \left( \frac{z_a}{z_0} \right) + 5} \quad (25)$$

where $Ri_u$ is the upper limit on $Ri_b$. Consequently, the stability correction with the bulk Richardson’s number becomes

$$r_{as} = \frac{1}{\left( 1 - \frac{Ri_b}{Ri_{cr}} \right)^2 r_a} \quad 0 \leq Ri_b \leq Ri_u \quad (26)$$
$$r_{as} = \frac{1}{\left( 1 - \frac{Ri_u}{Ri_{cr}} \right)^2 r_a} \quad Ri_b > Ri_u \quad (27)$$

### 3.3. Snow interception

While many models characterize the effect of the forest canopy on the micro-meteorology above the forest snowpack, few attempt to model explicitly the combined canopy processes that govern snow interception, sublimation, mass release, and melt from the forest canopy.
A simple snow interception algorithm is described here that represents canopy interception, snowmelt, and mass release at the spatial scales of distributed hydrology models.

During each time step, snowfall is intercepted by the overstory up to the maximum interception storage capacity according to

\[ I = fP_s \]  

where \( I \) is the water equivalent of snow intercepted during a time step, \( P_s \) is the snowfall over the time step, and \( f \) is the efficiency of snow interception (taken as 0.6) [Storck et al., 2002]. The maximum interception capacity is given by

\[ B = L_r m(LAI) \]  

where \( LAI \) is the single-sided leaf area index of the canopy and \( m \) is determined based on observations of maximum snow interception capacity. The leaf area ratio \( L_r \) is a step function of temperature

\[ L_r = \begin{cases} 0.004 & T_a > -5^\circ C \\ 0.001 & T_a \leq -5^\circ C \end{cases} \]  

which is based on observations from previous studies of intercepted snow as well as data collected during the field campaign described Storck et al. [2002]. Kobayashi [1987] observed that maximum snow interception loads on narrow surfaces decreased rapidly as air temperature decreases below 3\(^\circ\) C. Results from Storck et al. [2002] suggest that intercepted load on wide surfaces of conifer canopies is unaffected by decreases in air temperature to -5\(^\circ\) C.

Newly intercepted rain is calculated with respect to the water holding capacity of the intercepted snow \( (W_c) \), which is given by the sum of capacity of the snow and the bare
where \( h \) is the water holding capacity of snow (taken approximately as 3.5\%) and \( LAI_2 \) is the all sided leaf area index of the canopy. Excess rain becomes throughfall.

The intercepted snowpack can contain both ice and liquid water. The mass balance for each phase is

\[
\Delta W_{\text{ice}} = I - M + \left( \frac{Q_e}{\rho_w \lambda_v} + \frac{Q_m}{\rho_w \lambda_f} \right) \Delta t
\]

(32)

\[
\Delta W_{\text{liq}} = P_r + \left( \frac{Q_e}{\rho_w \lambda_v} - \frac{Q_m}{\rho_w \lambda_f} \right) \Delta t
\]

(33)

where \( M \) is the snow mass release from the canopy, and \( \lambda_s, \lambda_v, \lambda_f \) are the latent heat of sublimation, vaporization, and fusion respectively. Snowmelt is calculated directly from a modified energy balance, similar to that applied for the ground snowpack, with canopy temperature being computed by iteratively solving the intercepted snow energy balance (Eq. 1). Given the intercepted snow temperature and air temperature, snowmelt is calculated directly from Eqs 9 and 10. The individual terms of the energy balance are as described for the ground snowpack model. However, the aerodynamic resistance is calculated with respect to the sum of the displacement and roughness heights of the canopy. Incoming shortwave and longwave radiation are taken as the values at the canopy reference height. The same formulation is used for the albedo of the ground snowpack and the snow on the canopy (Eqs 16-17), the difference being in the transmitted radiative fluxes.

Snowmelt in excess of the liquid water holding capacity of the snow results in meltwater drip \((D)\). Mass release of snow from the canopy occurs if sufficient snow is available and
is related linearly to the production of meltwater drip

\[ M = 0 \quad C \leq n \]
\[ M = 0.4D \quad C > n \]  \hspace{1cm} (34)

where \( n \) is the residual intercepted snow that can only be melted (or sublimated) off the canopy (taken as 5 mm based on observations of residual intercepted load). The ratio of 0.4 in Eq. 34 is derived from observations of the ratio of mass release to meltwater drip \cite{Storcketal, 2002}.

4. Model calibration and evaluation

The measurements that were used for model calibration and testing spanned two winters at both the Oregon (1997 and 1998) and the BOREAS (1995 and 1996) sites. Calibration was performed utilizing only the 1997 Oregon data set, while the remainder of the observations were used for model evaluation. This allowed testing of the robustness of the model given the difference between the climates at the two study areas (maritime versus boreal). Canopy characteristics were derived from the North American Land Data Assimilation System (N-LDAS) data set \cite{GutmanandIgnatov, 1998; Hansenetal, 2000} based on the dominating forest characteristics at each observation site.

4.1. Umpqua, Oregon

4.1.1. Meteorological data

Hourly micro-meteorological observations from the shelterwood site were used to force the model during the calibration and testing periods. These included observations of precipitation, air temperature, incoming shortwave and longwave radiation, wind speed, relative humidity, and were taken to be representative of above-canopy conditions.
comparison of observed and predicted below-canopy meteorological data is shown in Fig. 2 for a rain-on-snow and radiation-dominated melt event. Wind speed at the 2-m height was corrected for snow accumulation beneath the anemometer and then scaled to the 80-m canopy height assuming a logarithmic profile and a surface roughness of 1 cm. Below-canopy shortwave radiation was taken as 16% of the shelterwood value based on observations (Fig. 2d). In addition, below-canopy wind speed was adjusted to match observations, i.e. 50% of shelterwood value (Fig. 2e). Below-canopy longwave radiation was calculated via

\[ L_c = (1 - F)L_0 + (F)^*\sigma T_a^4 \]

(35)

where \( L_0 \) is above canopy longwave. The effective fractional canopy coverage \( F \) was determined as 80% by calibrating Eq. 35 to observed longwave radiation beneath the forest canopy (Fig. 2a). Air temperature and humidity were assumed identical at the shelterwood and beneath-canopy sites.

**4.1.2. Calibration (1996/97)**

The model was calibrated using the shelterwood and beneath-canopy weighing lysimeter data for the 1996/97 winter season. During calibration, the only parameters used to adjust the predicted SWE were snow roughness length, and the value of \( m \) (Eq. 29), which controls the maximum snow interception capacity as a function of LAI. Other calibration parameters could have included the air temperature thresholds when rain or snow can fall (minimum and maximum respectively), which were set to \(-0.5^\circ C\) and \(0.5^\circ C\) respectively as default parameters prior to calibration. Fig. 3a (grey lines) shows the measured SWE from the two lysimeters at the beneath-canopy site compared to the model predicted SWE after calibration which resulted in values of 1 cm for the snow roughness length, and 0.01
for the LAI multiplier $m$. Although the model underestimates snow accumulation beneath the forest canopy during the initial snowfall event (1 Dec 1996 to 1 Jan 1997), it follows the variability of the observations closely, and predicts the complete melt of the snowpack very accurately. The correlation between the observed and modeled SWE (number of observations is 1770 with a 2-hr time step) was 0.94, while the relative mean squared error (RMSE) after calibration was 16.3 mm (9.6%). Fig. 3b shows the model predicted snow interception (in mm of SWE), with a maximum of about 70 mm. According to Storck et al. [2002] snow interception was observed to reach a maximum of about 40 mm, which might explain the discrepancy between the modeled and observed accumulated SWE during December 1996 to January 1997.

Fig. 3(a) (black lines) shows the same comparisons, as above, for the shelterwood site (no canopy). Model predictions show very good agreement with the observed SWE, after calibrating snow roughness length to 0.01 cm. The model slightly underestimates SWE late in December 1996, and early in April 1997, but is able to predict the complete melt of the snowpack. Correlation for this site was 0.99 (same number of observations as above), while the RMSE was 16.8 mm (4.9%).

Snow roughness length was used as a calibration parameter for the two sites. The difference in the snow roughness length between the shelterwood (clearing) and the site beneath the canopy can be explained by the usually shallower snowpack under the canopy (a deeper snowpack corresponds to a shorter roughness length), as well as the presence of forest litter which can increase the effective roughness length of snow. Lee and Mahrt [2004] used a roughness length of 0.1 cm for a snowmelt simulation over short vegetation in North Park, Colorado, while Essery et al. [1999] compared four different snow models
at a non-forested site in the French Alps using roughness lengths ranging from 0.05 to 0.1 cm. These roughness lengths are similar to the one used during the calibration at the shelterwood site, but simulating snow accumulation and ablation beneath the canopy with such a roughness length leads to an RMSE of 30.1 mm, almost twice the RMSE of the 1 cm length simulation.

4.1.3. Evaluation (1997/98)

The model was evaluated using observed SWE for both beneath-canopy and shelterwood sites, by using identical parameters (snow roughness length and LAI multiplier) as were used for 1996-97. Fig. 4a shows that the model underestimates snow accumulation during December 1997, but predicts SWE reasonably well until the beginning of March 1998. The model again underestimates snow accumulation during an early spring snowfall event, possibly due to overestimating intercepted SWE (Fig. 4b). Observations show that the snowpack melts almost completely by early April 1998, which the model captures fairly well. The RMSE for the testing period was 9.5 mm (18.7%), while the correlation coefficient was 0.87. It is interesting to note the difference in snow accumulation between the calibration and testing periods for the beneath-canopy site (about 150 and less than 50 mm respectively), while observed intercepted snow loads exceeded 30 mm during the calibration period but remained below 20 mm during the testing period [Storck et al., 2002]. Fig. 4a also shows comparisons between observed and model-predicted SWE at the shelterwood site; the model tracks the observed SWE variability quite well, although it does underestimate snow accumulation during late December 1997 with the error propagating through the initial melting event in mid-February 1998. After that, the model
predicts SWE very accurately, with RMSE and the correlation coefficient being 14.5 mm (6.4%) and 0.99 respectively.

4.2. BOREAS

4.2.1. Meteorological data

Simulations were performed for the period 1 September 1994 to 30 June 1996 at an hourly time step, with meteorological data (precipitation, air temperature, wind speed, incoming shortwave and longwave radiation, and relative humidity) aggregated to the same temporal resolution. Whenever observations were missing, data from the BOREAS Derived Surface Meteorological Dataset [Knapp and Newcomer, 1999] which contains interpolated data from several surface observation sites over the NSA and SSA were used. Comparisons of the simulated snow depth and water equivalent were made beneath the canopy using the snow course measurements, and AMS-measured snow depth from small open areas within different canopy types was also used to evaluate the simulated snow depths.

4.2.2. Evaluation (1994-96)

Fig. 5 show the continuous snow depth measurements from the AMS towers compared with the model-simulated snow depth at four different sites, OJP (a) and OA (b) at the SSA, and OJP (c) and YTH (d) at the NSA. The tower measurements have been aggregated to hourly to be directly comparable with the model simulations. The model snow depth is very close to the observations at SSA-OJP (Fig. 5a), with a correlation coefficient of 0.90 and an RMSE of 6.2 cm (8.8%), although it predicts melt too early by 2 days at the end of the 1995/96 season. The model underestimates snow depth throughout the winter of 1995/96 at the SSA-OA site (Fig. 5b), but performs very well
during 1994/95 with an overall correlation of 0.62 (n=16,056) and an RMSE of 15.9 cm (19.0%). Despite the differences during the second winter, the model appears to follow the accumulation and melt events very closely, which suggests that it might be overestimating snow density leading to an underestimation of snow depth. Results from the NSA-OJP site (Fig. 5c) show a relatively good agreement between the model predictions and the observations, although the model underestimates snow depth until late winter 1995 and starts melt with a delay for both seasons, nevertheless capturing the date of complete melt fairly accurately. The correlation for this site is 0.77 (n=16,056) and the RMSE is 13.8 cm (12.2%). Fig. 5d shows the a similar comparison for the NSA-YTH site, where the correlation coefficient was 0.83 (n=16,056) and the RMSE 12.5 cm (10.8%). The model predicts the dates of melt onset and complete melt relatively well (with differences of less than 1 week), but displays a slower melt rate and overestimates snow depth during spring 1995.

In addition to the automatic snow depth measurements, manual snow depth and water equivalent measurements were taken about every 15 days beneath the dominant canopy type close to each study site, with the exception of NSA-YTH. Figs 6 and 7 show comparisons between the model simulations and snow course measurements for SWE and snow depth respectively at selected dates for three sites. At the SSA-OJP site (jack pine), the model predicts the snow accumulation in terms of SWE very well during 1994-95 but underestimates depth suggesting an overestimation of density. However, both snow depth and water equivalent are captured relatively accurately during the ablation period, including a late accumulation event in April 1996. During the 1995/96 season the model underestimates SWE and depth during the accumulation, but predicts them well during
snowmelt (Figs 6a and 7a) with an overall RMSE of 14.2 mm for SWE and 10.1 cm for depth. At the SSA-OA site (aspen), model predictions of SWE follow the observations very closely (Fig. 6b) in both seasons (RMSE 13.9 mm), whereas simulated snow depth is consistently smaller than the observed during the 1994-95 season and very close to the observed depth during the 1995-96 water year (Fig. 7b) with the RMSE of 8.1 cm. At the NSA-OJP study site (jack pine), model performance was relatively poorer, with an underestimation of both SWE and depth during the winter of 1994/95 and late prediction of the onset of melt in both 1994/95 and 1995/95. This is evident in the RMSE which is 43.5 mm for SWE and 18.2 cm for snow depth, although the model does follow the observations closely during the accumulation period of 1995-96 (Figs 6c and 7c).

Observations of solar radiation were made beneath the canopies at three sites within the SSA, with each set of measurements lasting for about 4 days. The radiative flux of shortwave energy reaching the snow surface beneath a forest canopy is very important for the ground snowpack energy balance. Fig. 8 shows the comparison of the model-predicted and observed shortwave radiation reaching the ground surface at the SSA-OA site (aspen) during March 4, 1996 20:00 GMT and March 8, 1996 17:00 GMT. Measurements were taken with a 1-minute frequency, and were aggregated to hourly for purposes of this comparison. Fig. 8 shows the downward shortwave radiation measured at the top of the canopy for reference. The model is able to reproduce the time series of observed solar radiation quite well, with an RMSE of 25.6 W/m² (5.4%) and a correlation of 0.97 with observations. Although the duration of the measurements is relatively short, the comparison indicates that the models simplified approach for predicting shortwave
radiation transfer through forest canopies works well, which is also reflected in the SWE comparisons for this study site (Fig. 6b).

5. Conclusions

A mass and energy balance model for snow accumulation and ablation processes in forested environments was developed utilizing extensive measurements of snow interception and release in a maritime climate mountainous site in Oregon. A computationally efficient atmospheric stability correction algorithm was also developed, and the model was calibrated using one year (1996/97) of weighing lysimeter data and tested at the same site against measurements from the next year (1997/98). The model was able to reproduce the SWE evolution throughout both winters beneath the canopy as well as the nearby clearing, with correlations ranging from 0.87 to 0.99. The model was also evaluated using observations from the BOREAS field campaign in Canada, a much different continental climate with thinner snowpacks than the Oregon site, without any calibration. Simulations of snow depth for two seasons (1994-1996) were relatively close to the observations, with exceptions during the accumulation periods at certain forested sites when an underestimation of snow depth was evident that can be attributed mostly to an overestimation of snow density. That is corroborated by the model SWE validation, which has a relatively smaller error during the accumulation period. During snowmelt periods, the model was able to predict the dates complete melt fairly accurately, although there was a discrepancy for the onset of melt at the NSA-OJP forested site which could be alleviated by modifying the solar attenuation coefficient in the LDAS canopy parameters.

The model is intended primarily for large-scale applications. It has been incorporated as the standard snow scheme within the Variable Infiltration Capacity model, which rep-
resents sub-grid spatial variability by simulating state and fluxes in land cover/elevation tiles, and also contains modeling components for wind redistribution of snow [Bowling et al., 2004]. Within the VIC model, it is used in a parameterization of partial snow cover and frozen soil [Cherkauer and Lettenmaier, 2003]. It is also used in a real-time hydrologic forecast system for the western U.S. (Wood and Lettenmaier 2006), and has been used in numerous analyses, diagnoses, and predictions of climate variability and change (e.g. Su et al. 2006; Christensen and Lettenmaier 2007).

References


from Oak Ridge National Laboratory Distributed Active Archive Center, Oak Ridge, TN, USA.


Figure 1. Schematic of snow accumulation and ablation processes modeled by VIC.
Figure 2. Observed beneath-canopy and shelterwood meteorology at Umpqua for a major rain-on-snow event (26 Dec 1996 to 5 Jan 1997) and a spring melt event (16 Mar to 26 Mar, 1996) a) observed longwave radiation at both sites versus predicted beneath-canopy, b) observed air temperature and relative humidity at both sites, c) observed decay of snow albedo with age during two periods with no snowfall, d) hourly observed shelterwood and beneath-canopy shortwave radiation (same period as a) and b)), and e) hourly observed shelterwood and beneath-canopy wind speed (1997/98 season).
Figure 3.  a) Comparisons between observed and model-predicted SWE during the 1996/97 calibration period, b) model-predicted intercepted snow water equivalent during the same period.
Figure 4. a) Comparisons between observed and model-predicted SWE during the 1997/98 testing period at the shelterwood and beneath-canopy sites, b) model-predicted intercepted snow water equivalent during the same period.
Figure 5. Measured (gray) and VIC-simulated (black) snow depths at the BOREAS SSA-OJP (a), SSA-OA (b), NSA-OJP (c) and NSA-YTH (d) sites during September 1, 1994 to June 30, 1996.
Figure 6. Measured snow water equivalent (gray circles) from snow courses compared with VIC-simulated SWE (black line) at the BOREAS SSA-OJP (a), SSA-OA (b) and NSA-OJP (c) sites for the period September 1, 1994 to June 30, 1996.
Figure 7. Measured snow depth (gray circles) from snow courses compared with VIC-simulated depth (black line) at the BOREAS SSA-OJP (a), SSA-OA (b) and NSA-OJP (c) sites for the period September 1, 1994 to June 30, 1996.
Figure 8. Comparison of simulated (black solid line) and observed (gray line) shortwave radiation beneath the canopy, along with the incoming radiation (black dashed line) at the SSA-OA BOREAS site during March 4, 1996 20:00 GMT and March 8, 1996 17:00 GMT.