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Multiple-scale modelling of forest snow sublimation: initial findings

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Abstract:

Physically based equations describing snow interception and sublimation processes were applied to canopy-intercepted snow using a fractal scaling technique to provide a snow-covered forest boundary condition for a one-dimensional land surface scheme. Modification of the land surface scheme's calculation of turbulent transfer and within-canopy ambient humidity was required to accommodate this nested control volume approach. Tests in late winter in a southern boreal forest mature jack pine stand against measured sublimation found that the coupled model provides good approximations of sublimation losses on half-hourly and event bases. Daily sublimation averaged 0·5 kg m^{-2} daily, with minimum and maximum daily losses of 0·16 and 0·72 kg m^{-2}. Cumulative errors in estimating canopy temperature, humidity and intercepted snow load over 7 days of simulation were $-0{\cdot}7$ K, $-4{\cdot}15\%$ of the average observed vapour pressure, and 0·103 kg m^{-2}, respectively. At a nearby regenerating jack pine site, measured peak latent heat ranged from $-14{\cdot}6$ W m^{-2} to -40·9 W m^{-2}. Testing of the model at this site yielded reasonable estimates of latent and sensible heat fluxes during an overnight event, but did not estimate latent heat flux as well during events involving larger snow loads and incoming solar radiation, possibly as a result of errors introduced by solving for within-canopy humidity and neglect of subcanopy snow energetics. Further work to improve heat storage terms, and the inclusion of subcanopy snow energetics could help improve the coupled model performance. Copyright © 2000 John Wiley & Sons, Ltd.

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INTRODUCTION

Conifer canopies have high winter leaf areas and therefore can intercept a large proportion of annual snowfall. This intercepted snow is well exposed to the atmosphere and hence subject to higher sublimation rates than ground snow surfaces. Field results from boreal forests in western Canada show that 30% to 45% of annual snowfall sublimates as a result of its exposure as intercepted snow (Pomeroy and Gray, 1995; Pomeroy *et al*., 1998a). The sublimation of snow in high-latitude and high-altitude forests is an important hydrological process, which remains incompletely understood because of the complexity of the mass and energy exchanges, although recent progress has been made in process studies (Schmidt *et al*., 1988; Schmidt, 1991; Pomeroy and Schmidt, 1993; Lundberg and Halldin, 1994; Nakai *et al*., 1994; Claasen and Downey, 1995; Pomeroy and Gray, 1995; Harding and Pomeroy, 1996; Lundberg *et al*., 1998; Pomeroy *et al*., 1998a; Nakai *et al*., 1999).

The Canadian Land Surface Scheme, CLASS (Verseghy, 1991; Verseghy *et al*., 1993), is a physically based energy and water mass balance model for soil and vegetation that was developed for the Canadian General Circulation Model to improve simulation of soil thermal and hydrological regimes and to better represent

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the influence of vegetation on surface temperature and fluxes of heat and water vapour. CLASS has a relatively comprehensive treatment of snow processes, and in particular includes calculations for sublimation and mass balance of intercepted snow. However, a recent evaluation of snow accumulation and ablation processes by Pomeroy *et al*. (1998b) suggested that CLASS still has shortcomings in its representation of snow interception and ablation processes and recommended that studies of snow energetics be undertaken to provide improvements for land surface models such as CLASS. Recent development of process-based snow interception and sublimation algorithms (Pomeroy *et al*., 1998a) has provided a physically based technique for calculating interception gains and sublimation losses. However, application of the technique requires information on forest-stand-scale energetics and the canopy boundary layer, such as air temperature and wind speed within the canopy, that are not normally available from field measurements or climate model outputs. It is the objective of this paper to couple the 'small-scale' snow interception and sublimation algorithms used in Pomeroy *et al*. (1998a) with the 'large-scale' turbulent transfer and energy balance algorithms of CLASS and to evaluate the performance of the resulting coupled model by comparing the predicted within-canopy air temperature, humidity and intercepted snow load to field measurements.

METHODOLOGICAL APPROACH

Experimental sites

The experiment was conducted in two jack pine (*Pinus banksiana*) stands, one mature and one a regenerating 14-year-old clearcut, in the southern boreal forest of central Canada (in and near Prince Albert National Park, Saskatchewan). The climate is cool and subhumid with six months of snow cover and a cold dry winter. At the mature jack pine site, the winter leaf area index (LAI) is 2·2 $m^2 m^{-2}$ and canopy coverage is 82% as measured with a LICOR LAI-2000 (Gower and Norman, 1991), with a canopy height averaging 20 m. The regenerating clearcut site has a winter LAI of 2·5 $m^2 m^{-2}$, a canopy coverage of 86%, and an average canopy height of 5 m. Meteorological measurements taken at these sites include above-canopy measurements of net, incoming and outgoing solar radiation, radiation, as well as wind speed, air temperature, and relative humidity measured above and within the canopy. At each site, intercepted snow weight was measured by freely suspending a single mature tree, cut and sealed at the base with tar, from an in-line force transducer (Hedstrom and Pomeroy, 1998). Additionally, an eddy correlation system monitored by a Campbell Scientific 21X datalogger and consisting of a Kaijo Denki sonic anemometer, a Krypton hygrometer, and a fine wire thermocouple, was installed at the regenerating clearcut site during the 1997–98 winter season and provided measurements of latent and sensible heat fluxes. These measurements provided inputs and evaluation data for the model.

Model description

The coupled model considers the energy balance at three scales, that of the single ice particle, that of canopy-intercepted snow and that of the canopy as a whole (Fig. 1). Energy and mass exchange with the subcanopy snow surface and melt of intercepted snow are neglected for this analysis.

At all scales the energy balance is defined as

$$Q_n + Q_h + Q_e = \Delta U/\Delta t \tag{1}$$

where Q_n is net radiation, Q_h is sensible heat, Q_e is latent heat, U is internal energy of the control volume under consideration and t is time.

Net radiation for the ice sphere is calculated as

$$Q_n = \pi r^2 [S\downarrow + S\uparrow)(1 - \alpha_i) + L\downarrow + L\uparrow - 2\sigma T_i^4] \tag{2}$$

where r is the ice sphere radius (m), $S\downarrow$ is downward short-wave radiation (W m^{-2}) transmitted through to the canopy mid-point height (z, m), $S\uparrow$ is upward short-wave radiation (W m^{-2}) transmitted through to the canopy mid-point, α_i is the ice sphere albedo, set at 0·8, $L\downarrow$ is downward long-wave radiation (W m^{-2}) at the canopy mid-point, $L\uparrow$ is upward long-wave radiation (W m^{-2}) at the canopy mid-point, and $2\sigma T_i^4$ represents long-wave radiation emitted by the ice sphere, where σ is the Stephan–Boltzmann constant (W m^{-2} K^{-4}) and T_i is the ice sphere temperature (K). Solar radiation transmission is modelled using the results of Pomeroy and Dion (1996), who calculated boreal pine canopy transmissivity, τ, as

$$\tau = e^{(-\mu l)} \tag{3}$$

where μ is the extinction coefficient and l is the path length through the canopy. The path length for direct sun above a forest canopy is a function of solar angle above the horizon, θ, and d, the vertical depth of canopy scattering elements, where $l = d/\sin\theta$. The extinction coefficient is a function of the dimensionless extinction efficiency, Q_{ext}, having the form

$$\mu = Q_{ext}\frac{LAI}{d} \tag{4}$$

Pomeroy and Dion (1996) modelled Q_{ext} with the equation

$$Q_{ext} = 0.7810\theta\cos(\theta) + 0.0591 \tag{5}$$

for a boreal pine canopy. For solar radiation reflected from the subcanopy snow (albedo = 0·8), and transmitted back up through the canopy mid-point height, the same equations are used and the solar angle is set at 90°. Outgoing long-wave radiation, $\uparrow$ with respect to the canopy mid-point height, is modelled as the weighted sum of long-wave emission from the ground snow surface and the canopy. The ground snow long-wave emission component is evaluated at its measured temperature T_{gs} through canopy gaps (σT_{gs}^4

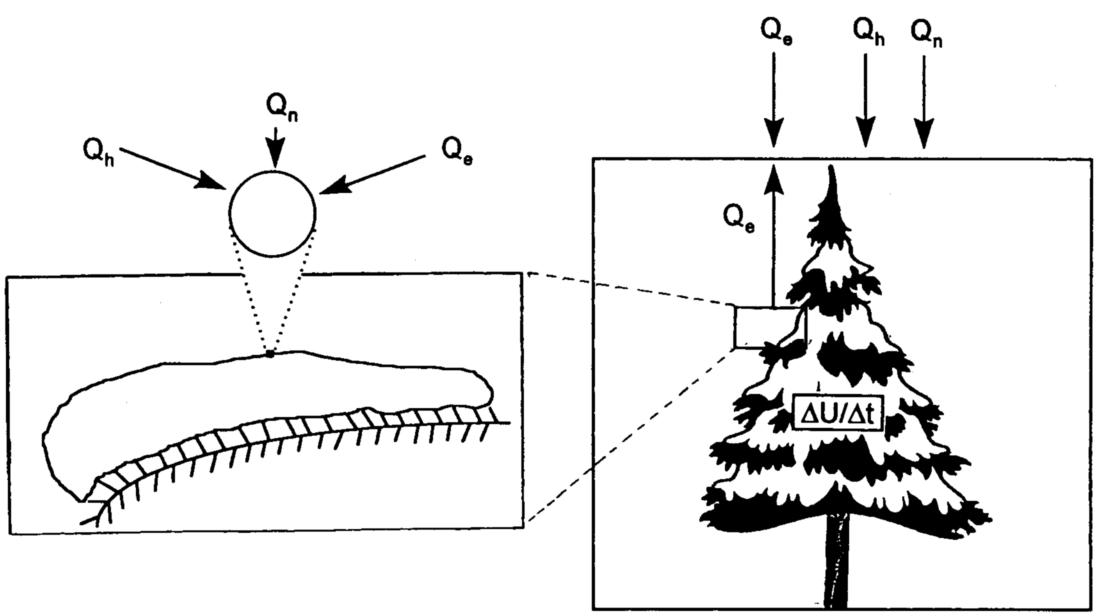

Figure 1. Illustration of the three control volumes under consideration in the development of the coupled model: the ice sphere, intercepted snow in the canopy, and the entire canopy

(1 − canopy closure)), whereas the canopy long-wave emission component is evaluated as its modelled temperature, T_c (σT_c^4 canopy closure). Incoming long-wave radiation ↓ with respect to the canopy mid-point height is modelled as the sum of incoming long-wave radiation above the canopy (Q_{1w}, as solved through a net radiation balance in CLASS) through larger gaps in the upper canopy (Q_{1w} (1 − canopy closure/2)) and long-wave emission from the canopy at its modelled temperature T_c (σT_c^4 canopy closure/2). The canopy closure reduction of 50% represents an assumed reduction of leaf area index by the same amount at this height.

Sensible heat transfer from a surface of uniform ice sphere Q_h is determined by

$$Q_h = C\lambda_T Nu(T_i - T_c) \tag{6}$$

where C is a surface-scale coefficient, equal to $2r$ for a sphere, λ_T is the thermal conductivity of air (J m^{-1} s^{-1} K^{-1}), Nu is the dimensionless Nusselt number for turbulent transfer of heat, and T_i and T_c are within-canopy air temperature and ice sphere surface temperature (K), respectively. Similarly, latent heat transfer is found as

$$Q_e = Ch_s D\rho_a Sh(q_c - q_i) \tag{7}$$

where h_s is the latent heat of sublimation (2·838 MJ kg^{-1}), D is the diffusivity of water vapour in air (m^2 s^{-1}), ρ_a is the density of the canopy air (kg m^{-3}), Sh is the dimensionless Sherwood number for turbulent transfer of water vapour, and q is specific humidity (kg kg^{-1}). Diffusivity, D, is calculated by the equation

$$D = 2.06\mathrm{E}^{-5}(T_c/273.15)^{-1.75} \tag{8}$$

where T_c is the canopy temperature (Thorpe and Mason, 1966). Lee (1975) confirmed that for Reynolds numbers between 0·7 and 10, the Nusselt and Sherwood numbers for a spherical particle are equivalent and are calculated by the equation

$$Sh = Nu = 1.79 + 0.606 N_R^{0.5} \tag{9}$$

where N_R is the particle Reynolds number, found as

$$N_R = \frac{2ru_z}{\nu} \tag{10}$$

where r is the particle radius, u_z is the within-canopy wind speed (m s^{-1}) and ν is the kinematic viscosity of air, which may be obtained from meteorological tables.

The within-canopy wind speed is calculated using a canopy wind flow model developed by Cionco (1965) that calculates wind speed at height z within the canopy (u_z), as a function of wind speed at the top of the canopy (u_H, m s^{-1}) and a canopy flow index, a, in the form

$$u_z = u_H e^{a\left(\frac{z}{H}-1\right)} \tag{11}$$

The canopy-top wind speed may be calculated from the measured reference height wind speed assuming a logarithmic wind profile above the canopy and given values of zero plane displacement height (d_0) and roughness length (z_0) for the canopy provided by CLASS, which calculates these parameters based on canopy type and height. The reference measurement height for the mature site is 27 m, and 8 m for the regenerating site. The canopy flow index (a) was obtained by rearranging Equation (11) and solved using measured values. These calculations showed that a was not constant, but rather increased exponentially as u_H decreased, and approached published values for similar canopy types (Cionco, 1978) at higher wind speeds.

Canopy flow index, a, was thus modelled as an exponential decay function of u_H

$$a = n + m\,\mathrm{e}^{-(u_H)} \tag{12}$$

(where n and m are constants) to better represent these observations in the model. Values of n and m were determined to have values of 2·43 and 3·46 at the mature site and 2·97 and 3·2 at the regenerating site. The fit of a to Equation (12) had low r^2 values (0·35 at the mature site and 0·16 for the regenerating site) owing to the sensitivity in the calculation of a to scatter in measured wind speeds. Figure 2 shows the within-canopy wind speed measured and modelled using Equation (11) at both sites. Within-canopy wind speed mean error (measured − modelled) is 0·007 m s^{-1}, with a standard deviation of 0·85 m s^{-1}, at the mature site and − 0·0036 m s^{-1}, with a standard deviation of 0·074 m s^{-1}, at the regenerating site.

Equations (1), (2), (6) and (7) are balanced by presuming that a steady-stake thermodynamic equilibrium rapidly develops and that the air at the ice surface is saturated at the 'ice bulb' temperature associated with T_c and q_c, and that $\Delta U/\Delta t$ is negligible (Schmidt, 1991).

To scale Equations (6) and (7) from the ice sphere to snow in the canopy, C is modified, to account for incomplete exposure of ice spheres located in clumps of intercepted snow (Pomeroy and Schmidt, 1993) and for the mass of snow in the canopy (Pomeroy and Gray, 1995), to become

$$C = \frac{3I}{2r^2\rho_i} k \left(\frac{I}{I^*}\right)^{-F} \tag{13}$$

where I is intercepted snow load per unit area of ground (kg m^{-2}), ρ_i is density of ice (900 kg m^{-3}), k is a dimensionless coefficient indexing the shape of snow, I^* is the maximum intercepted snow load (kg m^{-2}), and F is 1·0 minus the fractal dimension of intercepted snow. The dimensions of C change from metres when used at the ice sphere scale, and the units for energy are in watts, to m^{-1} when used at the canopy snow scale, and the units for energy are in W m^{-2}.

Analyses by Pomeroy *et al.* (1998a) derived values of 0·0114 for k and 0·4 for F for the mature site. Mean errors associated with these coefficients (measured − modelled) are 3·8E^{-6} with a standard deviation of

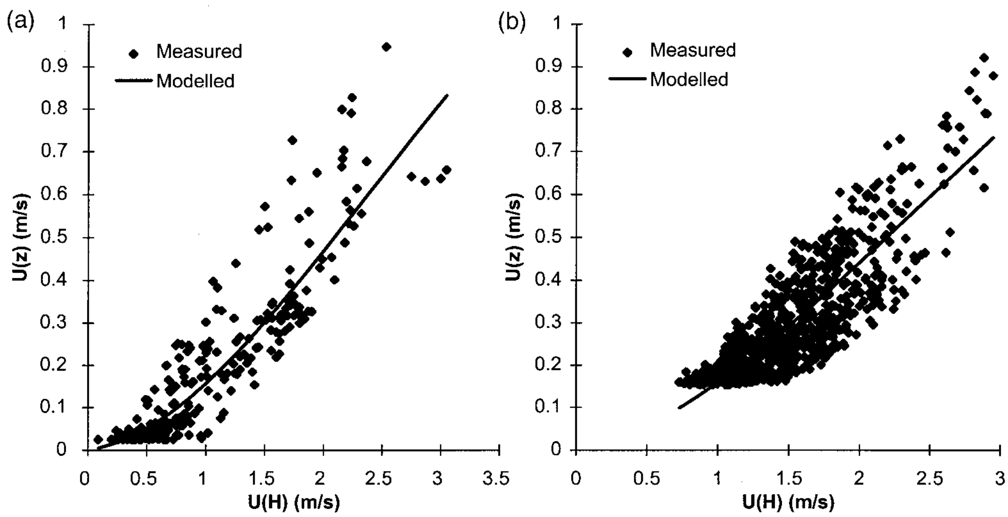

Figure 2. Measured and modelled within-canopy wind speed (u_z) versus measured wind at the canopy top (u_H) for (a) the mature site and (b) the regenerating site

0·0064 for k and $-0{\cdot}002$ with a standard deviation of 0·027 for F. Using the procedures described in Pomeroy *et al.* (1998a), values for k were calculated by using field measurements at the regenerating site and rearranging Equations (6), (7) and (13), and F was derived through the analysis of digital photographs of intercepted snow at the regenerating site. An average value of 0·0105 was found for k (with mean error of 0·0064 and standard deviation of 0·016). The same average value of 0·4 used at the mature site was found for F (with mean error of $-0{\cdot}003$ and standard deviation of 0·03). The value of I is calculated from intercepted snow mass on the weighed tree by applying a conversion factor, N (Hedstrom and Pomeroy, 1998). The value of N was determined by examining tree mass increases and differences between open area and subcanopy snow accumulation in Nipher snow gauges during periods when little sublimation or unloading had occurred. The value of I^* is a function of LAI, fresh snow density, ρ_s (kg m^{-3}), and a tree-species-specific branch loading coefficient, S_p (kg m^{-2}), following Hedstrom and Pomeroy (1998)

$$I^* = S_p(0.27 + 46/\rho_s)LAI \tag{14}$$

where S_p for pine is 6·6 (Schmidt and Gluns, 1991). It should be noted that the maximum canopy snow load from Equation (14) is more than 10 times greater than that specified originally by CLASS for mature pine canopies.

To scale to the canopy level, CLASS algorithms are used to calculate the canopy heat flux, Q_h (W m^{-2}), as

$$Q_h = \rho_a c_p (T_a - T_c)/r_a \tag{15}$$

where ρ_a is the density of air above the canopy (kg m^{-3}), c_p is the specific heat of air (1013 J kg^{-1} K^{-1}), T_a and T_c are the air temperatures at the reference height and within-canopy, respectively, and r_a is the atmospheric resistance (s m^{-1}), calculated using the reference height wind speed measurement, canopy zero-plane displacement height, roughness length and a stability correction. The stability correction used here is modified from that found in CLASS as upper level atmospheric information was not available, and is similarly derived from the Richardson number (Oke, 1987).

The calculation of latent heat flux, Q_e, is of the same form as Equation (15), substituting specific humidity for temperature and the latent heat of sublimation for the heat capacity of air. The calculation of r_a for vapour transfer in the coupled model was modified from its original form in CLASS, to represent resistance in the upper model layer that covers the top of the canopy through to the reference height. This change was made to accommodate model coupling, where the sublimation model calculates vapour flux from the surface to the canopy air (the canopy model layer), and CLASS calculates vapour flux through the upper model layer from the top of the canopy model layer.

Change in energy storage ($\Delta U/\Delta t$) is calculated as

$$\Delta U/\Delta t = \Delta T_c(c_i I + c_c M_c)/\Delta t \tag{16}$$

where ΔT_c is the change in average canopy temperature, c_i and c_c are the specific heats of ice (2090 J kg^{-1} K^{-1}) and canopy vegetation (2700 J kg^{-1} K^{-1}), I is the intercepted snow load, M_c is the standing canopy biomass (30 and 15 kg m^{-2} for mature and regenerating sites, respectively), and Δt is the time interval over which the change takes place. The average canopy temperature is calculated by an iterative solution of the energy balance, Equation (1), using measured net radiation above the canopy.

By coupling these models, CLASS provides an average canopy air temperature for the sublimation model, and the sublimation model provides a process-based vapour flux to the canopy model layer for removal through the upper model layer by CLASS. This required an iterative solution for canopy humidity as calculated by Equation (7) and by the CLASS algorithms, to force equality between the vapour flux values in the sublimation models and CLASS, while retaining the energy balance at various scales.

RESULTS

Dates and meteorological conditions for the sublimation events at the two sties are summarized in Table I.

Mature site

Two sublimation events ('A' and 'B') totalling eight days between Julian Days 56 and 79 (late February and March of 1995) were selected for analysis. The selection criteria were as follows: (i) data from periods when frequent site visits were performed, ensuring clean radiometers and working instrumentation, (ii) no snowfall, (iii) no apparent unloading or wind redistribution of snow, (iv) no evidence of intercepted snow melt, (v) no rainfall and (vi) unequivocal sublimation sequences with a continuous loss of intercepted snow mass. No event examined completely satisfied all points, however, the two events selected were those that best met the selection criteria.

Figure 3 shows the progression of measured and modelled ablation of intercepted snow load for the two events. Measured sublimation loss during these two events totalled 3·44 kg m^{-2} and averaged 0·5 kg m^{-2} daily, with minimum and maximum daily losses of 0·16 and 0·72 kg m^{-2}. The sublimation model coupled with CLASS, referred to as the coupled model, captures the timing and magnitude of sublimation reasonably well. The sharp diurnal cycling of sublimation also appears to be represented better by the coupled model than by the sublimation model running independent of CLASS, referred to as the uncoupled model, which used measured values of within-canopy air temperature and humidity. The coupled model, however, still overestimates sublimation at night and underestimates sublimation during the day, suggesting a problem with overestimation of heat storage. The mean error for the two events (measured – modelled) is 0·103 kg m^{-2} with a standard deviation of 0·0711 kg m^{-2}, suggesting reasonable model performance when compared with the total and average daily sublimation amounts.

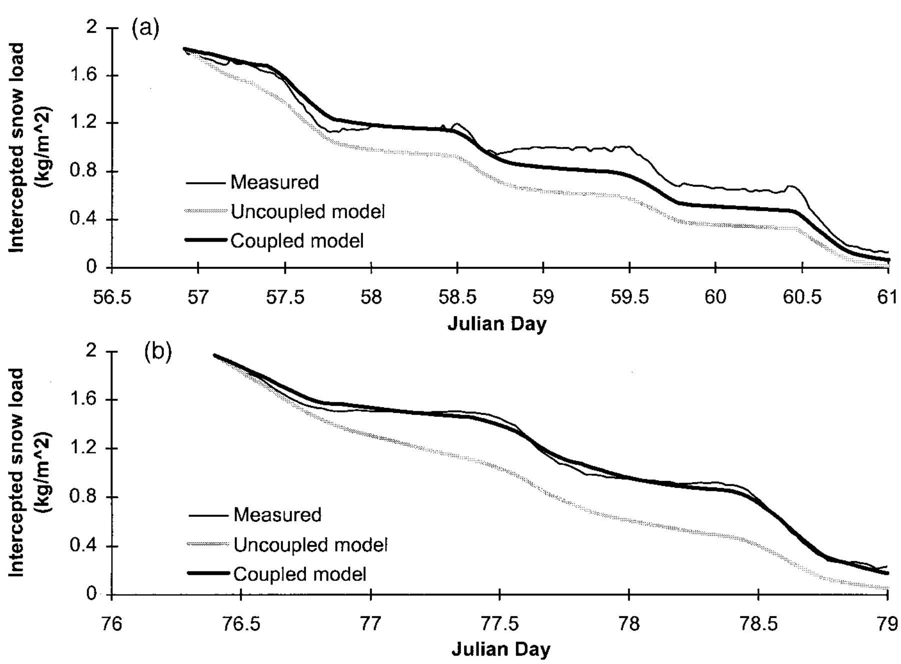

Figure 3. Cumulative sublimation measured using loss of snow mass on a weighed pine tree and modelled using the uncoupled and coupled sublimation models with a measured initial snow load for two sublimation events

Table I. Meteorological conditions during sublimation events measured and modelled in 1995 and 1997–98

Event	Site	Julian day/year	Mean snow load[a]	Total sublimation[a]	Mean T_a	Mean *RH*	Mean *VP*	Mean Q^*	Mean *U*
A	Mature	57–65/95	0·98	1·70	− 21	66	68	37	2·3
B	Mature	76–81/95	1·14	1·74	− 4·3	84	358	34	1·8
C	Regenerating	362–3/97	0·29	0·17	− 4·1	95	411	− 4	2·4
D	Regenerating	365/97	3·03	0·09	− 1·6	99	530	20	2·1
E	Regenerating	33/98	4·36	0·26	− 11·3	85	197	165	3·2

Mean snow load and total sublimation are in millimetres of snow water equivalent (mmSWE). T_a is ambient air temperature in the canopy (C). *RH* is relative humidity in the canopy (%). *VP* is ambient water vapour pressure in the canopy (Pa). Q^* is net radiation above the canopy (W m^{-2} s^{-1}). *U* is wind speed above the canopy (m s^{-1}), measured at the reference height.

To evaluate the energetics of the model, the predicted average canopy air temperature and ambient water vapour pressure were compared with measured values. The mean temperature error (measured − modelled) was found to be −0·71 K with a standard deviation of 1·79 K. A large part of this error occurred during nights when measured canopy air temperatures decreased substantially, approximately 15 to 20 K. During these periods modelled canopy temperatures showed less cooling (9 to 15 K) and averaged 3·4 K higher than measured values. CLASS does not consider the thermal conductivity of biomass in calculating U, assuming that all biomass is at the same temperature. When such rapid air temperature changes occurred, the diurnal period may not have been sufficient for the core of the standing canopy biomass to cool or warm by a similar amount. If true, the associated thermal capacity of the canopy is being overestimated by CLASS when air temperatures change rapidly, thereby overestimating the change in heat storage term and reducing the amount that the canopy temperature will change when solved for by closure of the energy balance. In contrast, the modelled vapour pressure tracked measured values very closely. The mean error (measured − modelled) was −4·15% of the average measured vapour pressure of 0·18 kPa, with a standard deviation of 5·14%, suggesting that the forcing of balanced vapour fluxes between the sublimation model and CLASS does not result in a misrepresentation of canopy humidity conditions at this site.

Regenerating site

The 1997–98 winter season was characterized by frequent snowfalls, cold calm periods, hoar frost accumulations, and an early onset of melt. As a result, few sublimation events occurred that met the criteria used at the mature jack pine site. Six events, ranging in length from 2·5 to 12 h, that appeared to best fit these criteria were used to derive sublimation model coefficients and in model runs. Three of these periods will be discussed.

Figure 4 shows the progression of measured and modelled sublimation and a comparison of measured and modelled sensible and latent heat fluxes during an overnight event (event ‘C’). Measured sublimation over this event was 0·17 kg m^{-2}, and measured latent heat flux averaged −8 W m^{-2} and peaked at −14·6 W m^{-2}. There is an underestimation of snow load ablation, suggesting errors in snow load estimation resulting from scaling from the hanging tree, a point measurement, to a canopy areal value. Despite this underestimation, latent and sensible heat fluxes are reasonably approximated. For this event, average error in heat fluxes are small; the latent heat error (measured − modelled) is 1·1 W m^{-2}, and that for sensible heat flux (measured − modelled) is −2·85 W m^{-2}. Mean errors in canopy temperature and vapour pressure (measured − modelled) were −0·21 K and −0·97% of the average measured vapour pressure of 0·41 kPa.

Figure 5 shows the progression of measured and modelled sublimation and a comparison of measured and modelled sensible and latent heat fluxes during a mid-day event with diffuse sunshine (event ‘D’). Measured sublimation over this event was 0·09 kg m^{-2}, and measured latent heat flux averaged −27·8 W m^{-2} and peaked at −39·5 W m^{-2}. Latent flux is underestimated despite a reasonable estimation of snow load ablation, suggesting an unaccounted source of water vapour such as an undetected snowfall near the beginning of the period or a difference between the suspended tree and the eddy flux footprint. However, sensible heat is well simulated. Harding and Pomeroy (1996) also observed imbalances in latent fluxes, but in their study measured latent heat exceeded available energy. For this event, the average error in latent heat flux (measured − modelled) is −12·8 W m^{-2}, and the average error in sensible heat flux (measured − modelled) is 0·38 W m^{-2}. Mean errors in canopy temperature and vapour pressure (measured − modelled) were −0·18 K and −0·48% of the average measured vapour pressure of 0·53 kPa.

Figure 6 shows the progression of measured and modelled sublimation and a comparison of measured and modelled sensible and latent heat fluxes during a mid-day event with bright sunshine (event ‘E’), generally representative of the other three events not discussed. Measured sublimation over this event was 0·26 kg m^{-2}, and measured latent heat flux averaged −28·2 W m^{-2} and peaked at −40·9 W m^{-2}. Sensible heat flux is reasonably simulated, whereas latent heat flux is overestimated, as is snow load ablation. Possible reasons for this will be discussed below. For this event, the average error in latent heat flux (measured − modelled) is 33·68 W m^{-2}, and the average error in sensible heat flux (measured − modelled) is 2·53 W m^{-2}. Mean errors in canopy temperature and vapour pressure (measured − modelled) were −0·94 K and 9·74% of the average measured vapour pressure of 0·2 kPa. For this event alone, these errors appear to be compounding because of the overestimation of temperature and the underestimation of vapour pressure. At the beginning of this event, the reference height wind speed averaged 1·9 m s^{-1} and the ambient air temperature rose rapidly (9 K in 3 h), a result of radiative warming of the canopy. Under such conditions, the assumption of equality between the air and canopy surface temperatures may have been violated, thus making the model unrepresentative of sublimation conditions in both parts of the model, and quite prone to error.

These results do not concur with those from the mature jack pine site, as events at the mature site showed underestimation of snow ablation during periods of bright sunshine, where the opposite is apparent at the regenerating site. The neglect of ground snowcover energetics, such as heat storage and sublimation, may account for this overestimation, as latent heat from the canopy may have been increased in the coupled model runs to compensate for the absence of these terms at this site. This perhaps is reflected by the large humidity error seen during this event. This is less of a concern at the mature site, as the canopy is much higher above the snowpack than at the regenerating site and therefore it is less likely that the ground snow will have a strong interaction with canopy energetics during periods when there is intercepted snow in the

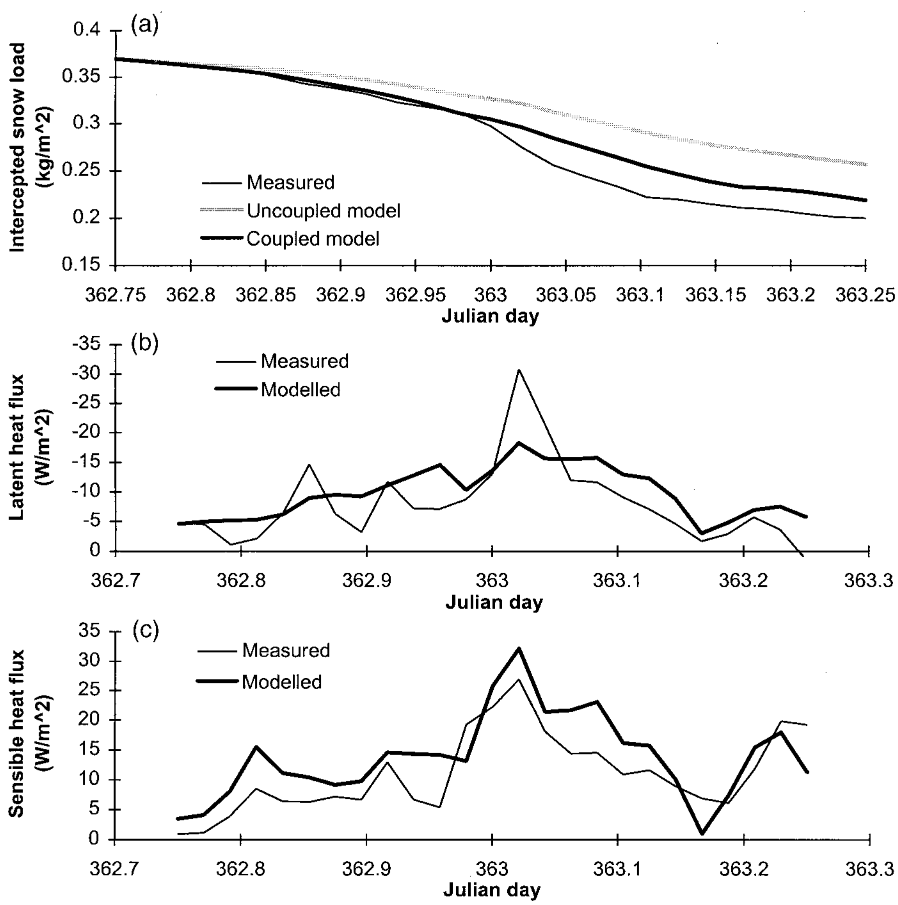

Figure 4. Cumulative sublimation measured using loss of snow mass on a pine tree and modelled using the uncoupled and coupled sublimation models, and comparison of measured latent and sensible heat fluxes with those calculated using the coupled sublimation model for an overnight sublimation event driven by turbulent transfer

canopy. Completion of the coupled model by including subcanopy energetics at this site may solve these problems.

CONCLUSIONS

Tests of a new multiscale snow sublimation boundary-layer model in a mature boreal forest pine stand found that the model can provide reasonable approximations of sublimation losses and within-canopy energetics on half-hourly and event bases. The new model permits stand-alone calculations of intercepted snow sublimation from standard measurements of above-canopy air temperature, humidity, wind speed and net radiation. At a mature jack pine site, measured daily sublimation averaged 0·5 kg m^{-2} daily, with minimum and maximum daily losses of 0·16 and 0·72 kg m^{-2}. Cumulative errors in estimating canopy temperature, humidity and intercepted snow load over 7 days of simulation were −0·7 K, −4·15% of the average measured vapour pressure, and 0·103 kg m^{-2}, respectively. The model underestimates diurnal canopy temperature changes, possibly due to an overestimation of canopy heat storage.

At a regenerating pine site, measured peak latent heat ranged from −14·6 W m^{-2} to −40·9 W m^{-2}. Testing of the model at this site yielded reasonable estimates of latent and sensible heat fluxes during an

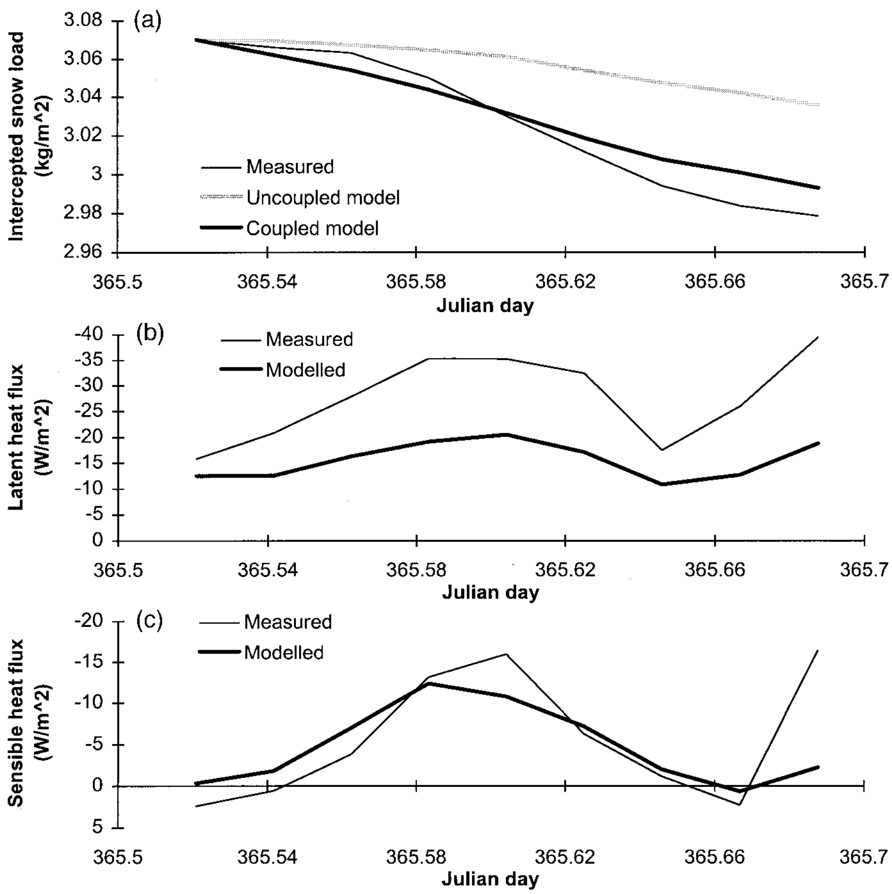

Figure 5. Cumulative sublimation measured using loss of snow mass on a suspended pine tree and modelled using the uncoupled and coupled sublimation models, and comparison of measured latent and sensible heat fluxes with those calculated using the coupled sublimation model for a mid-day sublimation event under diffuse sunshine

overnight event driven by turbulent transfer, but did not perform as well with respect to latent heat flux estimation during other events involving larger snow loads and high inputs of solar radiation. This may be related to errors introduced by the neglect of subcanopy energetics at this site. Further evaluation of the model throughout a season and in other environments is warranted once this issue has been addressed, as is an examination of improvements to CLASS performance with this coupling.

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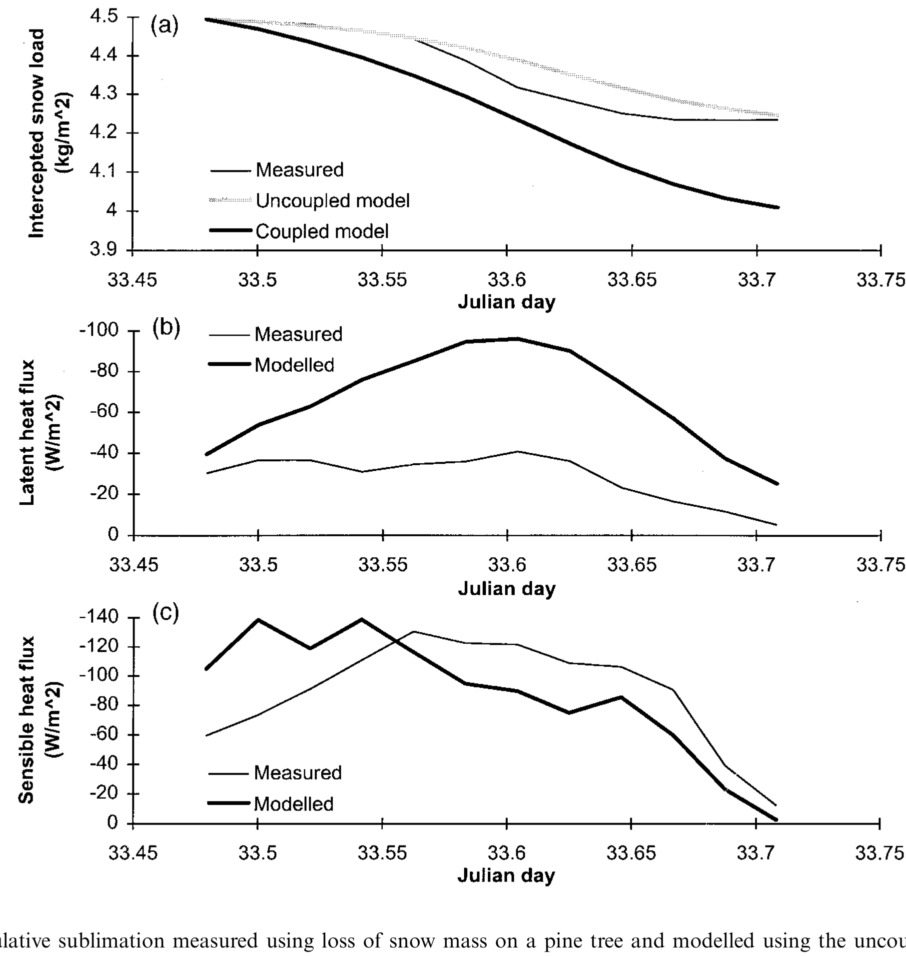

Figure 6. Cumulative sublimation measured using loss of snow mass on a pine tree and modelled using the uncoupled and coupled sublimation models, and comparison of measured latent and sensible heat fluxes with those calculated using the coupled sublimation model for a mid-day sublimation event under bright sunshine

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