Sublimation of Snow from Coniferous Forests in a Climate Model

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ABSTRACT

Improved representations of snow interception by coniferous forest canopies and sublimation of intercepted snow are implemented in a land surface model. Driven with meteorological observations from forested sites in Canada, the United States, and Sweden, the modified model is found to give reduced sublimation, better simulations of snow loads on and below canopies, and improved predictions of snowmelt runoff. When coupled to an atmospheric model in a GCM, however, drying and warming of the air because of the reduced sublimation provides a feedback that limits the impact of the new canopy snow model on the predicted sublimation. There is little impact on the average annual snowmelt runoff in the GCM, but runoff is delayed and peak runoff increased by the introduction of the canopy snow model.

1. Introduction

Boreal forests cover a significant fraction of the Northern Hemisphere land surface at midlatitudes and high latitudes. This major biome has important interactions with the climate and the carbon cycle (Apps et al. 1993; Betts 2000; Chapin et al. 2000). It is thus important that the general circulation models (GCMs) used in simulating climate change should have accurate representations of processes exchanging heat, moisture, and CO_2 between forest canopies and the atmosphere.

The presence of snowcover for much of the year has a major influence on the surface energy balance and hydrology of boreal forests. Intercepted snow on a forest canopy has a large exposed surface area, and a large fraction of the annual snowfall over boreal forests in dry continental climates sublimates from the canopy without ever reaching the ground (Schmidt and Troendle 1992; Pomeroy and Gray 1995; Lundberg and Halldin 2001). Snow on the ground below the canopy, however, is sheltered from wind and solar radiation, although it may be subject to large longwave radiation fluxes if the canopy is warm and snow free. Winter measurements of latent heat fluxes above coniferous canopies have shown that the sublimation is much greater when the canopy is snow covered than when it is not and snow on the ground is the only source of moisture (Harding and Pomeroy 1996; Nakai et al. 1999). Using meteorological measurements made above a pine canopy (Harding and Pomeroy 1996) to drive a surface model, Essery (1998) found that the assumed partitioning of snow between the canopy and the ground had a large influence on simulated heat and moisture fluxes, but coupling the surface model to a single-column atmospheric model reduced this sensitivity.

Several models of canopy and subcanopy snow processes have been developed recently for hydrological applications (Yamazaki and Kondo 1992; Hardy et al. 1997; Lundberg et al. 1998; Pomeroy et al. 1998b; Par-

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viainen and Pomeroy 2000; Storck 2000; Gryning et al. 2001; Gusev and Nasonova 2001). In this paper, we use simplified versions of parameterizations in the Pomeroy et al. (1998b) and Storck (2000) models to investigate interactions between canopy snow and the atmosphere in climate simulations with the Met Office GCM (Pope et al. 2000). Fluxes of heat, moisture and momentum between the surface and the atmosphere are calculated using the second version of the Met Office Surface Exchange Scheme (MOSES2), which includes a tiled representation of surface heterogeneity (Essery et al. 2003). Each land grid box, except those classified as ice, can contain a mixture of eight surface types: broad-leaf trees, needle-leaf trees, temperate C_3 grass, tropical C_4 grass, shrubs, urban development, inland water, and bare soil. Separate fluxes are calculated for each surface type and an area-weighted average is passed to the atmosphere. An advantage of surface tiling is that models of processes specific to certain surface types can be implemented in a more direct fashion than through effective parameters representative of the grid box as a whole. Here, the treatment of snowcover on the ''needleleaf tree'' tile is modified to use improved representations of interception and sublimation of snow on coniferous canopies.

MOSES2 has an optional canopy model that was not used by Essery et al. (2003) but is used here, so this is described in section 2 before modifications to represent canopy snow processes are introduced in section 3. The original version of MOSES2, including the canopy option, and the version modified to include canopy snow processes are referred to as MOSES2o and MOSES2c, respectively. Both versions are assessed in comparison with observations from sites in Saskatchewan, Canada; Oregon; and Sweden in section 4, and results from global climate simulations are presented in section 5.

2. The MOSES2 canopy model (MOSES2o)

With the optional canopy model, the surface skin temperature and conductive ground heat flux used by Essery et al. (2003) is replaced by a canopy layer temperature and radiative coupling between the canopy and the ground. At each time step, the GCM provides gridbox-mean values of downward shortwave radiation $\mathrm{SW}_\downarrow$ and longwave radiation $\mathrm{LW}_\downarrow$ at the surface, and temperature T_1, humidity q_1, and wind speed U_1 on the lowest atmospheric model level at height z_1 (typically around 20 m). For dense vegetation with negligible penetration of shortwave radiation to the ground, the net radiation absorbed by the canopy is

$$R_c = (1 - \alpha)\mathrm{SW}_\downarrow + \mathrm{LW}_\downarrow + \sigma T_0^4 - 2\sigma T_c^4, \quad (1)$$

where σ is the Stefan–Boltzman constant, α is the canopy albedo, T_c is the canopy temperature, and T_0 is the temperature of the ground or the snow surface below the canopy (a Beer's Law formulation is used for sparse canopies). A snow-free albedo α_o and a deep-snow albedo α_s are calculated for each surface type and weighted to give an effective albedo,

$$\alpha = (1 - f_s)\alpha_o + f_s\alpha_s, \quad (2)$$

where

$$f_s = \frac{d}{d + 10z_o} \quad (3)$$

for a tile with snow depth d and surface roughness length z_o. A forest tile with large z_o thus retains a low albedo even when covered with snow (Robinson and Kukla 1985; Pomeroy and Dion 1996; Betts and Ball 1997). The roughness length is reduced as a linear function of increasing snow depth, but this has little influence on rough forest tiles.

Expressions for surface fluxes of sensible heat and moisture over each tile are derived from the bulk aerodynamic formulas

$$H = c_p\frac{\rho}{r_a}\left(T_c - T_1 - \frac{gz_1}{c_p}\right) \quad \text{and} \quad (4)$$

$$E = \frac{\rho}{r_a + r_c}[q_{\mathrm{sat}}(T_c, p_*) - q_1], \quad (5)$$

where ρ and c_p are the density and specific heat capacity of dry air, and $q_{\mathrm{sat}}(T_c, p_*)$ is the saturation humidity at temperature T_c and surface pressure p_*. The aerodynamic resistance r_a depends on surface roughness, wind speed, and atmospheric stability. The moisture flux is additionally limited by canopy resistance r_c, calculated by a photosynthesis model (Cox et al. 1999) for dry canopies. When the canopy is snow covered, r_c is set to 0.

An areal canopy heat capacity, C_c, is calculated assuming specific heat capacities (in kilojoules per Kelvin per kilogram of carbon) of 570 for leaves and 110 for wood, estimated from figures given by Jones (1983) and Moore and Fisch (1986), and 2.1 kJ K^{-1} kg^{-1} for snow. The masses of carbon in leaves and stems per unit area of canopy with leaf area index L are parameterized as $\sigma_l L$ and $a_{wl}L^{5/3}$, with $\sigma_l = 0.1$ and $a_{wl} = 0.65$ for needleleaf trees (Cox 2001). The energy balance of a snow-covered canopy is expressed as

$$C_c\frac{dT_c}{dt} = R_c - H - L_sE - L_fS_M, \quad (6)$$

where L_s is the latent heat of sublimation, L_f is the latent heat of fusion, and S_M is the rate of snowmelt. The time derivative in Eq. (6) is discretized as

$$\frac{dT_c}{dt} \approx \frac{T_c - T_c(0)}{\delta t}, \quad (7)$$

where $T_c(0)$ is the value of T_c at the end of the previous time step and δt is the time step length (1800 s for the GCM). Linearizing q_{sat} about T_1 and R_c about T_0 to allow the elimination of T_c gives

$$q_{\mathrm{sat}}(T_c, p_*) \approx q_{\mathrm{sat}}(T_1, p_*) + D(T_c - T_1) \quad \text{and} \tag{8}$$

$$R_c \approx R_0 + 8\sigma T_0^3(T_0 - T_c), \tag{9}$$

where

$$D = \left.\frac{dq_{\mathrm{sat}}}{dT}\right|_{T=T_1} \quad \text{and} \tag{10}$$

$$R_0 = (1 - \alpha)\mathrm{SW}_\downarrow + \mathrm{LW}_\downarrow - \sigma T_0^4. \tag{11}$$

Eliminating T_c from Eqs. (1), (4), (5), and (6), the surface heat and moisture fluxes are then given by

$$H = c_p\frac{\rho}{r_a}\left[\frac{\tilde{R} - L_f S_M - L_s\psi(\rho/r_a)\Delta q_1}{(c_p + L_s D\psi)\rho/r_a + A_c}\right] \quad \text{and} \tag{12}$$

$$E = \psi\frac{\rho}{r_a}\left[\frac{D(\tilde{R} - L_f S_M) + (c_p\rho/r_a + A_c)\Delta q_1}{(c_p + L_s D\psi)\rho/r_a + A_c}\right], \tag{13}$$

where

$$\psi = \frac{r_a}{r_a + r_c}, \tag{14}$$

$$\Delta q_1 = q_{\mathrm{sat}}(T_1, p_*) - q_1 + D\frac{gz_1}{c_p}, \tag{15}$$

$$A_c = \frac{C_c}{\delta t} + 8\sigma T_0^3, \quad \text{and} \tag{16}$$

$$\tilde{R} = R_0 + A_c\left(T_0 - T_1 - \frac{gz_1}{c_p}\right) + \frac{C_c}{\delta t}[T_c(0) - T_0]. \tag{17}$$

Equation (13) is the familiar Penman–Monteith equation (Monteith 1981), modified by canopy heat storage, longwave radiation beneath the canopy, and snowmelt. Equations (12) and (13) for the surface fluxes take exactly the same forms with or without the canopy model, but the A_c and $\tilde{R}$ terms are modified (Essery et al. 2003).

Inverting Eq. (6), a first estimate of the canopy temperature is diagnosed as

$$T_c = T_0 + \frac{1}{A_c}\left\{R_0 - H - L_s E + \frac{C_c}{\delta t}[T_c(0) - T_0]\right\} \tag{18}$$

with H and E given by Eqs. (12) and (13) for $S_M = 0$. If this gives $T_c > T_m = 273.15$ K for a snow-covered canopy, the melt rate is calculated as that required to set $T_c = T_m$, provided there is sufficient snow to melt, and the surface fluxes are recalculated accordingly. Since $T_c \leq T_m$ and $r_c = 0$ whenever there is snow cover, MOSES2o effectively assumes that all snow is held on the canopy. An improved canopy snow model is described in the next section.

3. Canopy snow processes (MOSES2c)

a. Interception and unloading

From measured loads on individual branches, Schmidt and Gluns (1991) suggested an expression,

$$S = \overline{S}\left(0.27 + \frac{46}{\rho_s}\right), \tag{19}$$

for the maximum snow load (kg) that can be held per unit branch area, where ρ_s is the density of fresh snow and $\overline{S}$ was given as 6.6 for pine or 5.9 for spruce. Hedstrom and Pomeroy (1998) scaled this to give the maximum intercepted load (in kg m^{-2}) as $I_{\max} = SL$ for a canopy with leaf area index L. Typical values of ρ_s and $\overline{S}$ are used here to estimate the snow interception capacity as $I_{\max} = 4.4L$, much greater than the capacities of 0.1–0.2 L often used by land surface models (e.g., Verseghy et al. 1993; Sellers et al. 1996) for the interception of both rain and snow. From the model of Pomeroy et al. (1998a), the change in canopy load during a time step with snowfall amount S_f on a canopy with initial load I_0 is

$$\Delta I = 0.7(I_{\max} - I_0)(1 - e^{-S_f/I_{\max}}). \tag{20}$$

For light snow loads and moderate snowfall, this gives $\Delta I \sim 0.7S_f$, close to the value of $0.6S_f$ observed for snowfall on Douglas fir by Storck et al. (2002).

Unloading of snow from a canopy accelerates with increasing temperature because of weakening of the snow structure and decreased branch stiffness (Schmidt and Pomeroy 1990). Based on measurements by Storck et al. (2002), we set the unloading rate equal to 40% of the diagnosed canopy snowmelt rate during a time step.

b. Sublimation

According to Thorpe and Mason (1966), an ice sphere of radius r, density ρ_i, and mass $m = 4/3\pi\rho_i r^3$ in air at temperature T_c and humidity q_c sublimates at rate

$$\frac{dm}{dt} = 2\pi r D_w \rho \mathrm{Sh}[q_c - q_{\mathrm{sat}}(T_c, p_*)], \tag{21}$$

where

$$D_w = 2.06 \times 10^{-5}\left(\frac{T_c}{T_m}\right)^{-1.75} \tag{22}$$

is the diffusivity (m^2 s^{-1}) of water vapor in air, and Sh is the Sherwood number, given by Lee (1975) as

$$\mathrm{Sh} = 1.79 + 0.606\left(\frac{2ru}{\nu}\right)^{1/2}, \tag{23}$$

for air viscosity ν (m^2 s^{-1}) and ventilation velocity u, equal to the wind speed within the canopy in this case. Canopy wind speeds are often modeled using an exponential profile (Thom 1971; Cionco 1978),

$$u(z) = U_h e^{-n(1-z/h)}, \tag{24}$$

where U_h is the wind speed at the top of the canopy and h is the canopy height. The commonly used value of $n = 2.5$ gives a reasonable agreement with wind speeds measured in a pine canopy by Parviainen and

Pomeroy (2000), although a better match can be obtained by making n a function of U_h. Assuming spherical grains of typical radius 500 μm and taking a nominal height $z = 0.6h$ gives $\mathrm{Sh} \approx 1.79 + 3U_h^{1/2}$.

The exposed surface area of intercepted snow is less than the total surface area of the constituent grains, so the sublimation rate is assumed to be scaled by the ratio of these areas. From measurements of intercepted snow clumps in trees and sublimation from an artificial conifer, Pomeroy and Schmidt (1993) derived an exposure coefficient,

$$C_e = k_1\left(\frac{I}{I_{\max}}\right)^{-0.4}, \tag{25}$$

to give the sublimation per unit ground area as

$$E = -C_e\frac{I}{m}\frac{dm}{dt}. \tag{26}$$

As the snow grain size and exposure are difficult to model accurately, they have to be calibrated for practical applications. Calibrating against measured sublimation and relative humidities with respect to water, Pomeroy et al. (1998b) chose a value of $k_1 = 0.01$; recalibration for use with specific humidities gives $k_1 = 0.02$. Combining Eqs. (21) and (26) gives

$$E = \frac{\rho}{r_i}[q_{\mathrm{sat}}(T_c, p_*) - q_c], \tag{27}$$

where a resistance

$$r_i = \frac{2\rho_i r^2}{3C_e I D_w \mathrm{Sh}} \tag{28}$$

for transport of moisture from the intercepted snow to the canopy air space has been defined. Neglecting sublimation from snow on the ground and storage in the canopy space, q_c can be eliminated to give

$$E = \frac{\rho}{r_a + r_i}[q_{\mathrm{sat}}(T_c, p_*) - q_1], \tag{29}$$

which has the same form as Eq. (5) with r_i in place of r_c. For low wind speeds or stable conditions, $r_a > r_i$ and sublimation is controlled by moisture transport out of the canopy air space; for high wind speeds and convective conditions, r_i dominates.

c. Ground snowmelt

The canopy model option in MOSES2o was originally developed to improve the forecasting of temperatures over grass, and it only has radiative coupling between the canopy and the ground (Best 1998; Best and Hopwood 2001). Initial tests of MOSES2c showed that the melting of snow on the ground beneath a forest canopy was unrealistically delayed unless a turbulent component was included in the energy balance of the snowpack. Pomeroy and Granger (1997) found the observed net radiation to be insufficient to account for the snowmelt rate beneath a pine canopy and attributed the difference to turbulent fluxes.

Blyth et al. (1999) developed a "dual-source" version of MOSES including heat and moisture fluxes between the canopy air space and the ground for sparse vegetation, but this greatly complicates the model and has only been implemented in an off-line version. Here, we adopt a simpler approach by neglecting moisture transport beneath the canopy and assuming that the canopy air temperature is equal to T_c. The sensible heat flux beneath the canopy is parameterized as

$$H_c = c_p\frac{\rho}{r_{ah}}(T_0 - T_c), \tag{30}$$

where r_{ah} is an in-canopy resistance. Including H_c in the canopy energy balance, the forms of Eqs. (12), (13), and (18) are retained, but Eq. (16) is modified to

$$A_c = \frac{C_c}{\delta t} + 8\sigma T_0^3 + c_p\frac{\rho}{r_{ah}}. \tag{31}$$

Integrating the eddy diffusivity through the canopy using the wind profile given by Eq. (24), Huntingford et al. (1995) gave an expression for r_{ah} as

$$r_{ah} = \frac{e^n h}{kn(h-d)}\left(\frac{r_a}{U_h}\right)^{1/2}[e^{-nz_{og}/h} - e^{-n(z_o+d)/h}], \tag{32}$$

where k is the von Kármán constant, z_{og} is the roughness length of the surface beneath the canopy, and z_o and d are the roughness length and displacement height for the canopy, respectively. Assuming that $z_{og} \ll h$ and following Huntingford et al. (1995) in setting $z_o = 0.1h$ and $d = 0.75h$ gives

$$r_{ah} = \frac{4e^n}{kn}\left(\frac{r_a}{U_h}\right)^{1/2}(1 - e^{-0.85n}) \approx 43\left(\frac{r_a}{U_h}\right)^{1/2}. \tag{33}$$

Storck (2000) used a resistance that included a correction for atmospheric stability beneath the canopy and found this to improve the simulation of snowmelt.

4. Offline tests

During development of the canopy snow model for use in the GCM, MOSES2o and MOSES2c were tested using data from sites in Canada, the United States, and Sweden, representing midlatitude continental, midlatitude maritime, and low Arctic environments. In these offline tests, the models were driven with meteorological observations.

a. Saskatchewan, Canada

A jack pine stand in Prince Albert National Park, Saskatchewan, was among the sites used by Pomeroy and colleagues in developing the canopy interception and sublimation models (Hedstrom and Pomeroy

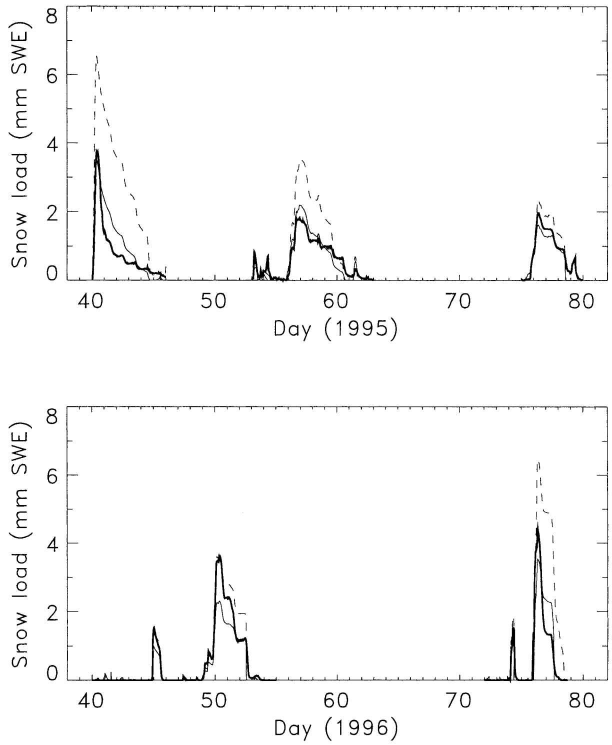

FIG. 1. Snow loads on a pine canopy in Saskatchewan from measurements (heavy solid lines), and simulations with the canopy snow model (thin solid lines) and without (dashed lines).

1998; Pomeroy et al. 1998b; Parviainen and Pomeroy 2000). The trees were 16–22 m tall, with a winter leaf area index of 2.2 and 82% canopy coverage. Snow interception was measured by weighing a suspended tree.

Observed and simulated canopy snow loads during February and March of 1995 and 1996 are compared in Fig. 1. Because all snow is held in the canopy, MOSES2o generally overestimates the peak load after snowfall, but rapid sublimation of the intercepted snow gives a reasonable match to the observed periods of canopy snowcover. Two of the events in 1996 gave snow loads close to the measured snowfall, so MOSES2c underestimates the load in these cases, but inclusion of the canopy snow model generally improves the simulations of peak snow loads and decay rates.

Although the two versions of MOSES2 give similar predictions of how long snow persists on the canopy, the fate of the snow differs. Over the periods shown, there was 18 mm of snowfall in 1995 and 16 mm in 1996. In the simulations with MOSES2o, 96% of this snow sublimates from the canopy in 1995 and 69% in 1996. For MOSES2c, these numbers are 52% and 31%, the rest of the snow remaining on the ground below the canopy at the end of the simulations. Over several winters, Pomeroy and Gray (1995) and Pomeroy et al. (1998b) recorded sublimation of between 29 and

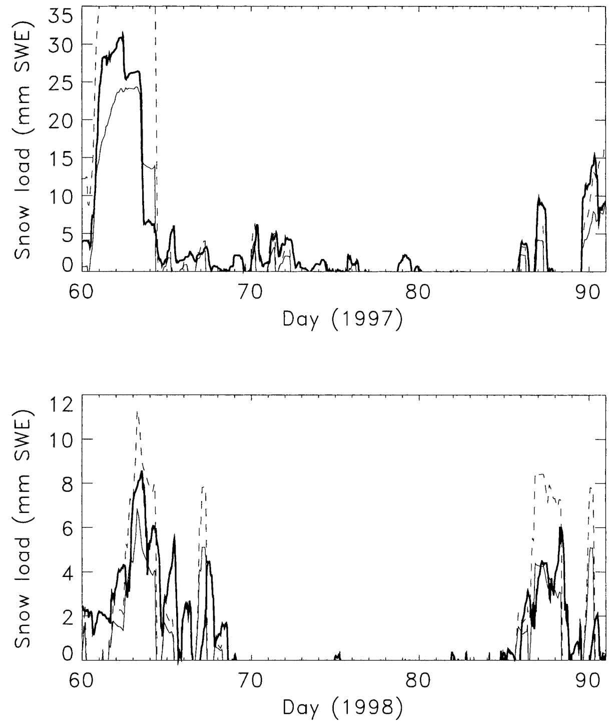

FIG. 2. As in Fig. 1 but for a Douglas fir canopy in Oregon.

39 mm yr^{-1} at this site, amounting to between 30% and 32% of the total snowfall.

b. Oregon

Storck et al. (2002) gathered meteorological and hydrological data in the Umpqua National Forest, Oregon. Modeled canopy snow loads are compared with measured loads on cut Douglas fir trees during March 1997 and 1998 in Fig. 2. MOSES2o greatly overestimates snow loads after heavy snowfall events; MOSES2c often underestimates the load but generally gives an improved simulation.

Frequent midwinter melts and high humidities limit the sublimation. From observations, Storck et al. (2002) estimated that 10% of the snow that falls, or about 100 mm yr^{-1}, sublimates from the canopy; although the fractional sublimation is lower than for the colder and drier climate of Saskatchewan, the absolute amount is greater. Both versions of the model sublimate about 8% of the annual snowfall, close to the observed fraction, but the remaining snow is removed from the canopy by direct melt in MOSES2o and by a combination of melt and unloading in MOSES2c.

Figure 3 shows ground snowpack loads measured by lysimeters under a dense Douglas fir canopy and in a clearing for the winters of 1996/97 and 1997/98; the influence of the canopy in decreasing the snow depth on the ground is clear. MOSES2o does not model snow beneath the canopy, but a reasonable simulation is given

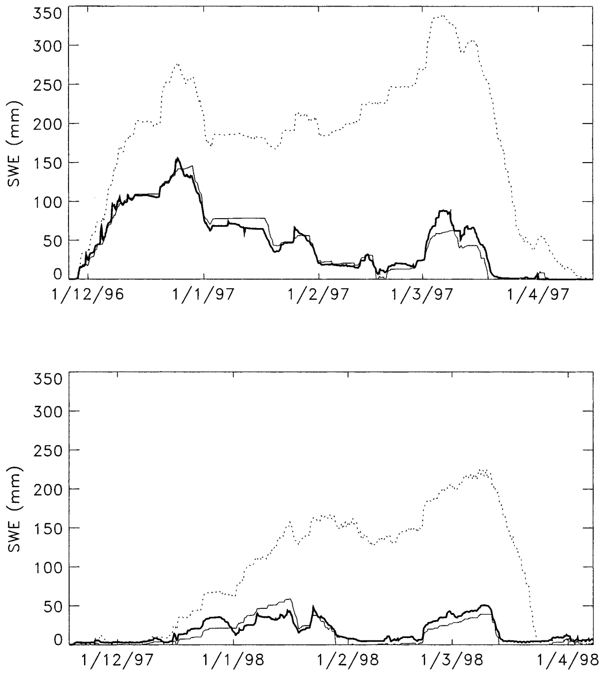

FIG. 3. Snow loads on a lysimeter beneath a Douglas fir canopy (heavy solid lines) and simulated with the canopy snow model (thin solid lines). Dotted lines show snow loads measured in a clearing.

by MOSES2c. Sensible heat provides a large fraction of the modeled energy input to the snow on the ground; this is surprising, considering the low wind speed environment found beneath the canopy in dense forests. Further experimental and theoretical work is required on radiative and turbulent transfers through forest canopies for application in surface models.

c. Sweden

In the Project for Intercomparison of Land Surface Parameterization Schemes (PILPS) 2e Arctic model intercomparison (Bowling et al. 2003), 21 land surface schemes were used to simulate snowmelt and runoff for the Torne–Kalix river system in northern Scandinavia over the period 1989–98. Part of the project involved comparisons between simulated and measured discharge from Ovre Lansjarv, a forested 1341 km^2 subbasin of the Kalix. The peak snowmelt runoff simulated by MOSES2o was underestimated and too early. Compared with models that gave better simulations of the peak flow, MOSES2o had high winter sublimation. Incorporating the canopy snow model in MOSES2c limits sublimation from the canopy, making more snow available for melt, and improves the timing of the peak runoff by delaying the melt of snow below the canopy; these results are shown by Essery and Clark (2003). The canopy snow model reduces the average sublimation from 23% of the annual snowfall to 12%, varying between 7% and 19% over the nine complete winters in the simulation.

5. GCM results

The HadAM3 version of the Met Office GCM (Pope et al. 2000) was run for 15 years with both MOSES2o and MOSES2c. Sea surface temperatures and sea-ice extents were prescribed from climatology, and fractions of vegetation types within grid boxes were derived from the University of Maryland 1-km land cover classification (Hansen et al. 2000). Comparisons between results from similar simulations and climatology are discussed by Essery et al. (2003). To investigate the role of atmospheric feedbacks, the GCM run using MOSES2c included parallel off-line calculations using MOSES2o for needle-leaf tree tiles of zero area that thus respond to the meteorology of the modified GCM but do not influence it.

Figure 4a shows the average annual snowfall in the simulation with MOSES2c; deposition is greatest in maritime and high elevation regions. Sublimation from the forested fractions of model grid boxes is shown in Fig. 4b and presented as a fraction of the annual snowfall in Fig. 4c. A histogram of the sublimation fractions for forested grid boxes (Fig. 4d) is peaked between 20% and 30% of the annual snowfall, similar to measured values for sublimation from coniferous canopies in continental environments (Pomeroy and Gray 1995).

Sublimation from forested and open fractions of grid boxes and gridbox-mean sublimation in the modified (MOSES2c) and control (MOSES2o) simulations are compared in Fig. 5; the solid lines are linear least squares fits with slopes given by the numbers printed on the plots. In the offline MOSES2o simulation with meteorological forcing from the modified GCM, the average forest sublimation is 83 mm, almost twice the 43 mm obtained using MOSES2c (Fig. 5a). When MOSES2o is fully coupled to the atmosphere in the GCM, however, moistening and cooling of the air because of increased sublimation gives a negative feedback, limiting the forest sublimation to 70 mm (Fig. 5b). Atmospheric feedbacks do not just influence the forest sublimation; the average sublimation from the open fractions of forested grid boxes is *increased* from 8 mm in the original model to 12 mm in the modified model (Fig. 5c). The overall result, shown in Fig. 5d, is that the canopy modifications give only a relatively small decrease from 30 to 25 mm yr^{-1} in the gridbox-mean sublimation. Similarly, Samuelsson et al. (2003) found a large decrease in sublimation on reducing roughness lengths in offline simulations for the PILPS 2e domain but only a small decrease in coupled simulations.

With little impact on the recycling of precipitation by sublimation, there is little difference in the annual snowfall over forests in the two versions of the model (Fig. 6a); implementing the canopy snow model reduces the

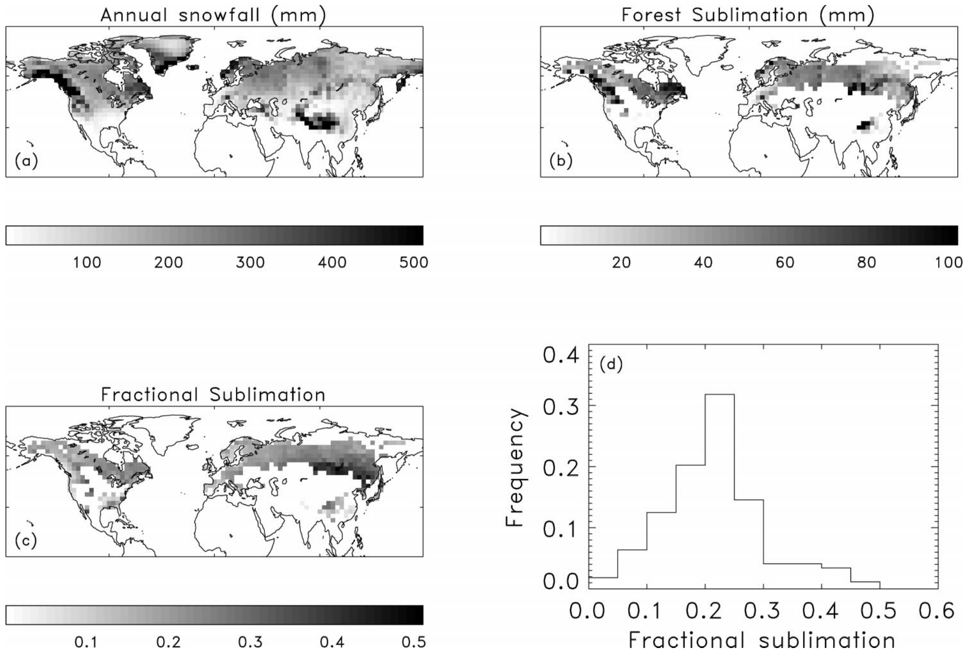

FIG. 4. (a) Simulated average annual snowfall, (b) average annual sublimation from the forested fractions of model grid boxes, (c) forest sublimation as a fraction of annual snowfall, and (d) histogram of forest sublimation fractions.

average simulated snowfall over forested grid boxes from 209 to 205 mm yr^{-1}. Because this decrease in snowfall almost balances the decrease in sublimation, the canopy snow model only increases the average amount of snow available for melt (Fig. 6b) from 179 to 180 mm.

Although there is little change in the average annual runoff, MOSES2c changes the timing of runoff. Figure 7 shows average monthly runoff from snowmelt for a grid box centered on 55°N, 105°W (Saskatchewan). Both models give an average snowmelt runoff of 191 mm yr^{-1} for this grid box, but MOSES2c (solid line) gives a later onset for melt and a higher peak melt than MOSES2o (dashed line), similar to the influence seen in off-line simulations (Essery and Clark 2003). A river routing scheme is being developed for the GCM so that land surface process representations affecting runoff can be assessed in comparison with observed discharge from large basins.

Despite the negative feedbacks noted here, changes in forest sublimation could induce positive feedbacks involving subgrid processes not currently represented in the GCM. Although the surface energy and moisture budgets are tiled, MOSES2 lacks a model of horizontal moisture transfers within grid boxes; soil moisture is assumed to be homogeneous, so differences in moisture availability between surfaces in a grid box that might influence the simulation cannot develop in response to differences in sublimation and melt. The increase in sublimation from the open fractions of forested grid boxes requires further attention. MOSES2, in common with all current GCM land surface schemes, does not represent the transport and sublimation of wind-blown snow, which can remove a large fraction of the snowfall from open environments. Increasing surface roughness in MOSES2 increases sublimation, but studies with blowing snow models have shown that stubble or shrubs can decrease sublimation by trapping snow (Pomeroy and Gray 1995; Liston et al. 2002).

6. Conclusions

A model of snow processes in coniferous forest canopies has been implemented in the MOSES2 land surface scheme. Falling snow is partitioned into interception by the canopy and throughfall to the ground. Canopy snow may be removed by sublimation, unloading, and melt. Snow on the ground beneath the canopy is melted by radiative and turbulent heat fluxes.

The canopy snow model greatly reduces sublimation from forests and improves the performance of MOSES2 in offline simulations. When coupled to the GCM, negative feedbacks through the atmosphere limit the grid-box-mean differences between the original and modified versions of MOSES2. Cooling and moistening of the air because of greater sublimation from forests in the original version limits the sublimation relative to the same model driven offline with the meteorology of the modified GCM. Lower sublimation from the open fractions of forested grid boxes in the original model further

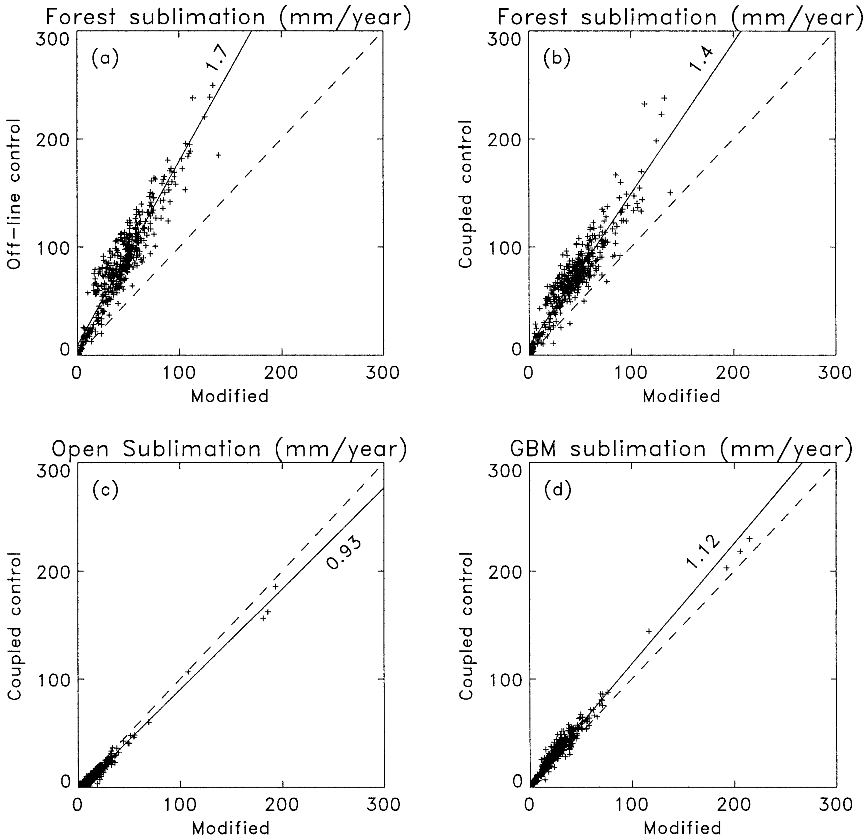

FIG. 5. Average annual sublimation from (a) forested fractions of grid boxes in modified and off-line control simulations, (b) forested fractions of grid boxes in modified and coupled control simulations, (c) open fractions of forested grid boxes in modified and coupled control simulations, and (d) forested grid boxes in modified and coupled control simulations (gridbox means). Solid lines are least squares fits with the given slopes.

offsets differences in gridbox-mean sublimation between the model versions. Changes in sublimation and snowfall balance to give little change in the snow available for melt, but the canopy snow model modifies the timing of melt.

Vegetation distributions were fixed in the simulations presented here, but MOSES2 was developed to complement the Top-down Representation of Interactive Foliage and Flora Including Dynamics (TRIFFID) vegetation dynamics model (Cox 2001), which is being used

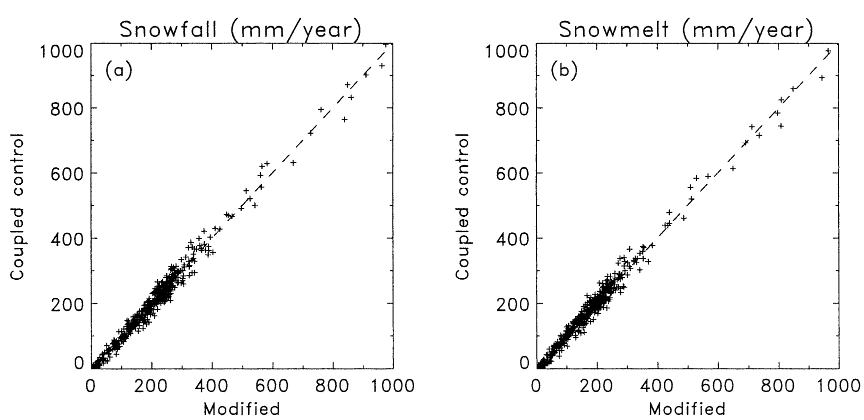

FIG. 6. Average annual (a) snowfall and (b) snowmelt for forested grid boxes in modified and coupled control simulations.

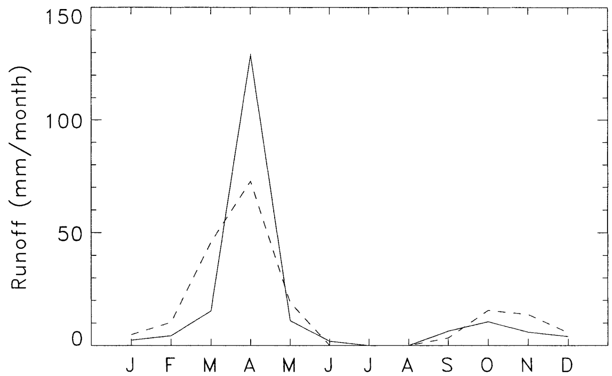

FIG. 7. Average monthly snowmelt runoff for a grid box centered on 55°N, 105°W, in simulations using the original (dashed line) and modified (solid line) versions of MOSES2.

to model interactions between the global carbon cycle and climate change (Cox et al. 2000). In simulations where vegetation distributions can respond to local climate and moisture stresses, modifying the surface model will modify the responses. With increasing interest in the role of the biosphere in climate and climate change, it is likely that improved representations of interactions between snow and vegetation will become increasingly important.

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