ADVANCES IN ATMOSPHERIC SCIENCES, VOL. 25, NO. 3, 2008, 348–360

Simulation of Snow Processes Beneath a Boreal Scots Pine Canopy

LI Weiping*[1,2] (李伟平), LUO Yong[1,2] (罗 勇), XIA Kun[1,2] (夏 坤), and LIU Xin[3] (刘 新)

[1] *Laboratory for Climate Studies, China Meteorological Administration, Beijing* 100081

[2] *National Climate Center, China Meteorological Administration, Beijing* 100081

[3] *Institute of Tibetan Plateau Research, Chinese Academy of Sciences, Beijing* 100085

(Received 5 July 2007; revised 30 September 2007)

ABSTRACT

A physically-based multi-layer snow model Snow-Atmosphere-Soil-Transfer scheme (SAST) and a land surface model Biosphere-Atmosphere Transfer Scheme (BATS) were employed to investigate how boreal forests influence snow accumulation and ablation under the canopy. Mass balance and energetics of snow beneath a Scots pine canopy in Finland at different stages of the 2003–2004 and 2004–2005 snow seasons are analyzed. For the fairly dense Scots pine forest, drop-off of the canopy-intercepted snow contributes, in some cases, twice as much to the underlying snowpack as the direct throughfall of snow. During early winter snow melting, downward turbulent sensible and condensation heat fluxes play a dominant role together with downward net longwave radiation. In the final stage of snow ablation in middle spring, downward net all-wave radiation dominates the snow melting. Although the downward sensible heat flux is comparable to the net solar radiation during this period, evaporative cooling of the melting snow surface makes the turbulent heat flux weaker than net radiation. Sensitivities of snow processes to leaf area index (LAI) indicate that a denser canopy speeds up early winter snowmelt, but also suppresses melting later in the snow season. Higher LAI increases the interception of snowfall, therefore reduces snow accumulation under the canopy during the snow season; this effect and the enhancement of downward longwave radiation by denser foliage outweighs the increased attenuation of solar radiation, resulting in earlier snow ablation under a denser canopy. The difference in sensitivities to LAI in two snow seasons implies that the impact of canopy density on the underlying snowpack is modulated by interannual variations of climate regimes.

Key words: snow process, boreal forest, radiation, turbulent flux, sensitivity

DOI: 10.1007/s00376-008-0348-2

1. Introduction

Forests are well recognized to alter both accumulation and ablation of snow on the ground. Forest canopies may intercept a portion of snowfall, and the intercepted snow may later fall to the ground, melt, or sublimate depending on the climate conditions (Golding and Swanson, 1986; Lundberg and Halldin, 1994). Canopies strongly influence the radiation balance at the snow surface by absorbing and reflecting incoming solar radiation and altering the emission of thermal radiation. The forest canopy shelters the snow surface from wind and alters temperature and humidity gradients, greatly reducing the efficiency of turbulent energy transfer (Marks et al., 1998). Niu and Yang (2004) addressed the effects of physical processes due to the canopy on snow mass and energy balances in boreal ecosystems. They suggested that explicitly including a canopy heat storage term in the canopy energy balance equation decreases the spuriously large amplitude of the diurnal canopy temperature variation and reduces the excessive daytime sensible heat flux from the canopy downward to the snow surface for the usually stable conditions with snowpack under relatively warm canopy.

In general, snowmelt decreases as the canopy density increases. However, with conditions of high atmospheric temperature, calm winds, and high snow albedo, greater snowmelts can occur under a dense canopy due to enhanced infrared thermal radiation from the canopy elements than under a sparse canopy (Yamazaki and Kondo, 1992). As to the net all-

*Corresponding author: LI Weiping, liwp@cma.gov.cn

wave radiation, an important snowmelt energy source, FitzGibbon and Dunne (1983) examined the relationship between forest cover, net radiation, and snowmelt, and found that the net all-wave radiation increases with forest density as the transmission of solar radiation through the canopy decreases. But Suzuki et al. (1999) found that the net radiation does not vary with forest density during the snowmelt season because of the net effects of decreased below-canopy snow albedo which offsets the attenuation of solar radiation and increased downward longwave radiation as forest density increases.

The relative contribution of the radiative components to the energy balance of snowpack depends on latitude, those atmospheric conditions (especially cloud amount) that influence the potential incident solar radiation, the characteristics of the canopy, and the snow surface albedo, all of which influence the surface absorbed solar radiation. When snow albedo is low, a steady decrease in snowmelt energy is found with increasing canopy cover, while snowmelt rates for high albedos are either insensitive or increase with increasing canopy cover (Sicart et al., 2004). Nakabayashi et al. (1999) obtained similar results concerning the seasonal progression of snowpack on surface albedo and solar radiation. At the beginning of the snowmelt season when the albedo remained high, the forest effect became null because the increase in thermal radiation from the canopy offset the decrease in solar radiation. As the albedo gradually lowered with the advance of the snowmelt season, the decrease in solar radiation due to the presence of forest cover exceeded the increase in thermal radiation and the tendency for the forest to effectively decrease net all-wave radiation reaching the ground became evident.

As for the relative importance of vegetation cover and climate regimes on the under-canopy snow process, Link and Marks (1999) proposed that seasonal snow cover in the boreal environment may be more sensitive to land use transitions (e.g., forest versus bare soil), rather than climate shifts, due to the strong forcing exerted by vegetation canopies on radiation transfer processes. Gelfan et al. (2004) documented only a moderate sensitivity of snow accumulation to forest leaf area, while a substantial variation was observed from season to season with changing weather patterns, which suggested that the ensemble of snow processes is more sensitive to variations in atmospheric processes than in specific characteristics of vegetation cover.

Despite many previous studies concerning vegetation impact on snowpack, it is not clear which is the dominant energy term contributing to snowmelt beneath the forest, partly due to the lack of available observations for establishing under canopy energy budgets, and the highly variable weather conditions in the seasonal evolution of snowpack. Besides, uncertainties involved in using empirical formulations to estimate the under-canopy solar and thermal radiation budgets in previous studies are also responsible for some controversies in modeling under-canopy snow processes. In this study, the snow mass balance and energy budget of the snowpack under a high latitude boreal forest is investigated by using a physically-based, multi-layer snow model Snow-Atmosphere-Soil-Transfer scheme (SAST) (Sun et al., 1999) and a physically-based land surface process model Biosphere-Atmosphere Transfer Scheme (BATS) (Dickinson et al., 1986). Thereby, attempts are made to address snow melting energetics at different stages of the snow season. The model and data used in this study are briefly introduced in section two. The third section concerns the modeling results. A brief discussion of the results is given in the fourth section which is followed by concluding remarks in the fifth section.

2. Model and data

2.1 *Model*

In the BATS version 1e (Dickinson et al., 1993) the following four aspects to be employed in this study were modified: (1) the original one layer snow scheme was substituted by the multi-layer snow scheme SAST; (2) canopy albedo and canopy-absorbed solar radiation was modified according to Dai and Zeng (1997) and Yang et al. (1999, see their Eq. 6) on the basis of the two-stream concept where the canopy-absorbed solar radiation increases with leaf area index (LAI); (3) the emissivity of thermal radiation from the canopy is parameterized to increase with LAI (Oleson et al., 2004), but in the original BATS solar and longwave radiation transmitted through vegetation canopy was neglected; (4) in addition to the original stability correction in the calculation of drag coefficients in BATS, the transfer coefficient between the canopy air and underlying soil is simply assumed to be half of the original value (0.004) used in BATS (see Eq. 65 of Dickinson et al., 1993) to account for the usually stable atmospheric condition under boreal forest canopy during snow season.

SAST was developed from up-to-date comprehensive and complex snow schemes, but with substantial simplification and improvement for GCM applications, validations against observations at different sites indicate that SAST is a reliable snow model in climate studies (Sun et al., 1999; Sun and Xue, 2001; Jin et al., 1999; Xue et al., 2003). The scheme includes many important physical processes such as snow compaction, heat conduction, snow grain growth, and snow melting. There are three prognostic variables in the model: specific enthalpy, snow water equivalent, and snow depth. The vapor effects on snow processes have been simply parameterized in the energy balance calculation. A unique feature of SAST is using enthalpy rather than temperature in the energy balance equation, which greatly simplifies the computation procedure for the phase change calculation in the snow process.

SAST consists of at most three layers. In order to have reasonable simulations of diurnal changes in surface temperature, the surface layer thickness should be thinner than the thermal damping depth of snow and is no more than 3 cm. The second layer thickness is restricted to less than 20 cm because the diurnal variation of snowpack properties is more pronounced in the top 15–20 cm of the snowpack.

2.2 *Data*

A Scots pine forest site with sandy glacial till at Hyytiälä (61°51′N, 24°17′E) in Finland (Vesala et al., 1998) was chosen for this study and two consecutive snow seasons, 2003–2004 and 2004–2005, were investigated. The average height of the Scots pine forest is 15 m, the all-sided LAI is 6 (Vesala et al., 2005), thus one-sided LAI is 3. According to the definition of Chen et al. (1997), the effective LAI is 2.4 by assuming the ratio of woody to total vegetation area is 0.25 and the clumping index is 0.6. Meteorological data at 30 minute time-intervals for complete winters was obtained. Daily precipitation amounts were measured with a shielded gauge in a clearing 1 km from the forest site; no correction was applied. Cloud amounts were estimated from temperature and humidity as described by Liston and Elder (2006), and daily precipitation was divided equally between half-hour periods with cloud cover exceeding 70%. Total precipitation is partitioned into snow and rain using a linear function between 0°C (100% snow) and 2°C (100% rain). Weekly snow depth observations were recorded at seven snow pits, the values from five of which represent snow depth under closed forest canopy and the other two below relatively sparse canopy. In this study, we treat the average of the seven pits' snow depth data as the overall observations for the forest site.

Each numerical experiment was integrated from 1 October to the end of April of the following year. The simulations of snow surface energy and snow mass balance from November to the end of the snow season are analyzed to investigate the snowmelt energetics under a high latitude boreal forest canopy at different stages of the snow season. In this paper, snow refers to those accumulated on the ground under the canopy.

3. Analysis of results

Simulations of the seasonal evolution of snow depth under the canopy in both snow seasons captured the observed accumulation, melting, and final ablation of snow quite well, except for the fact that the peak of snow depth was underestimated in 2004 and overestimated in 2005 (Fig. 1). In the 2003–2004 snow season, accumulation of snow began in late November, 2003 and continued gradually until late January, 2004 (Fig. 1a), followed by an abrupt increase from the end of January to early February and some decrease afterwards. The under canopy snow depth peaked around the beginning of March, a little earlier than observed and was followed by decreasing snow depth, partly due

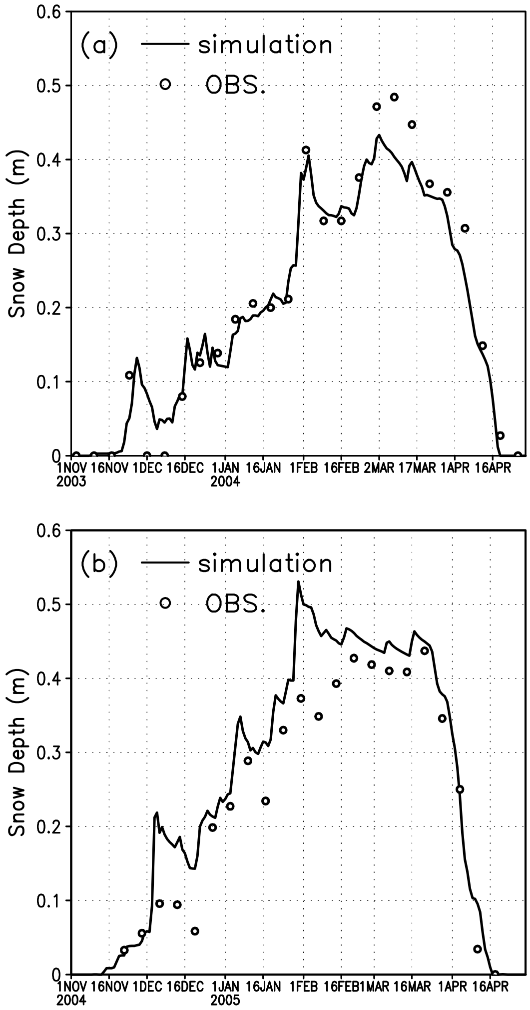

Fig. 1. Temporal evolution of simulated daily mean snow depth (m) beneath the vegetation canopy in comparison with weekly observations in (a) 2003–2004 and (b) 2004–2005 snow seasons.

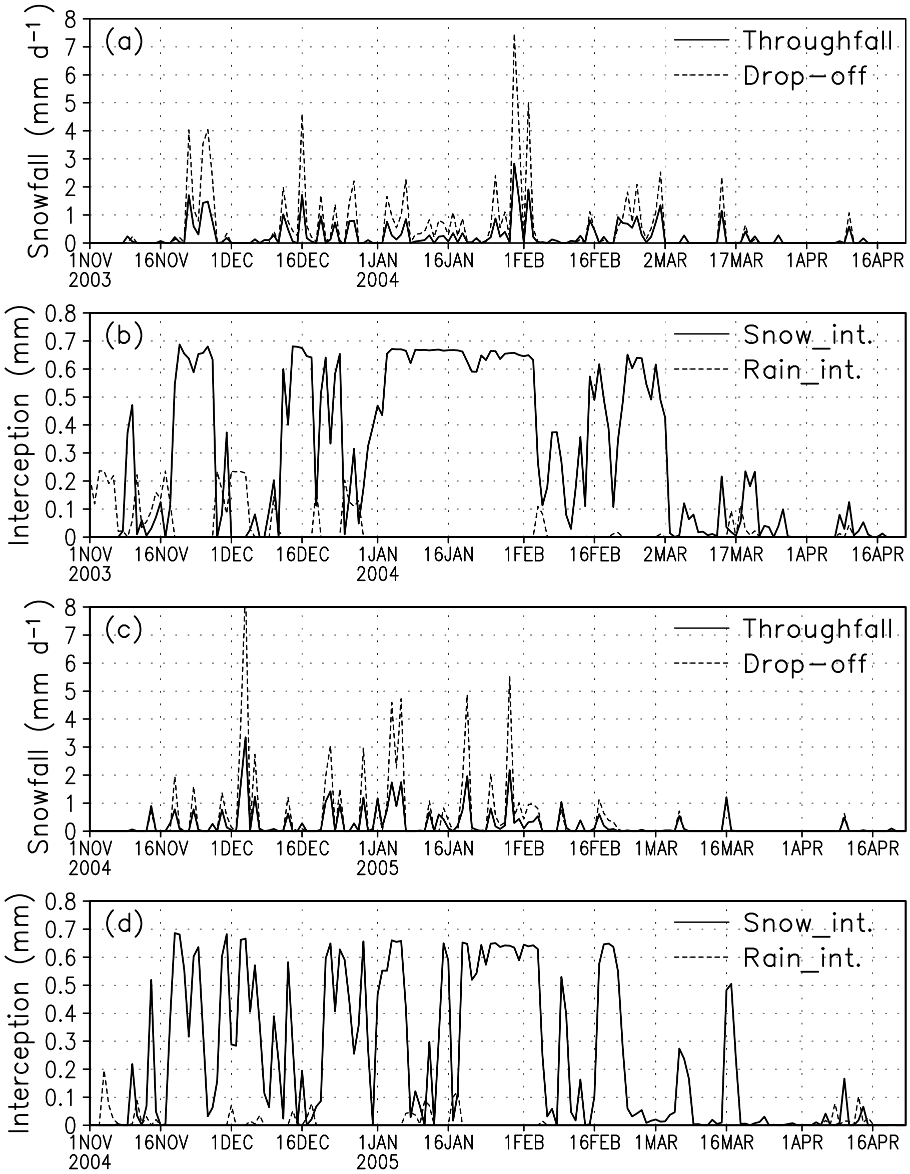

Fig. 2. Seasonal evolution of daily mean (a, c) direct throughfall of snowfall and drop-off from the canopy-intercepted snow (mm d^{-1}) and (b, d) the interception of snowfall and rainfall by the canopy (mm) in snow seasons of (a, b) 2003–2004 and (c, d) 2004–2005.

to compaction but mostly attributable to melting since early April. The final snow ablation occurred on 19 April, a little earlier than observed. Compared with the modeled snow water equivalent (not shown), several melting processes occurred during the snow season, one is from the end of November to early December, one in late December, the last one from early to mid-April. The decrease of snow depth in early February (from 0.4 m to 0.32 m) is due to snow compaction rather than melting. The overall characteristics of snow accumulation in 2004–2005 are similar to that in 2003–2004, except that the under-canopy snowpack before February is deeper than its counterpart in the 2003–2004 snow season (Fig. 1b).

Components contributing to snow accumulation under the canopy and precipitation intercepted by foliage are compared in Fig. 2. Since we assume 100% vegetation coverage for the simulation at the Hyytiälä site, direct throughfall of snowfall is quite small for the relatively dense canopy, since interception by the canopy is significant. Snow accumulation on the canopy that exceeds the maximum holding capacity of the canopy drops off towards the ground, which is at least twice the direct throughfall and is

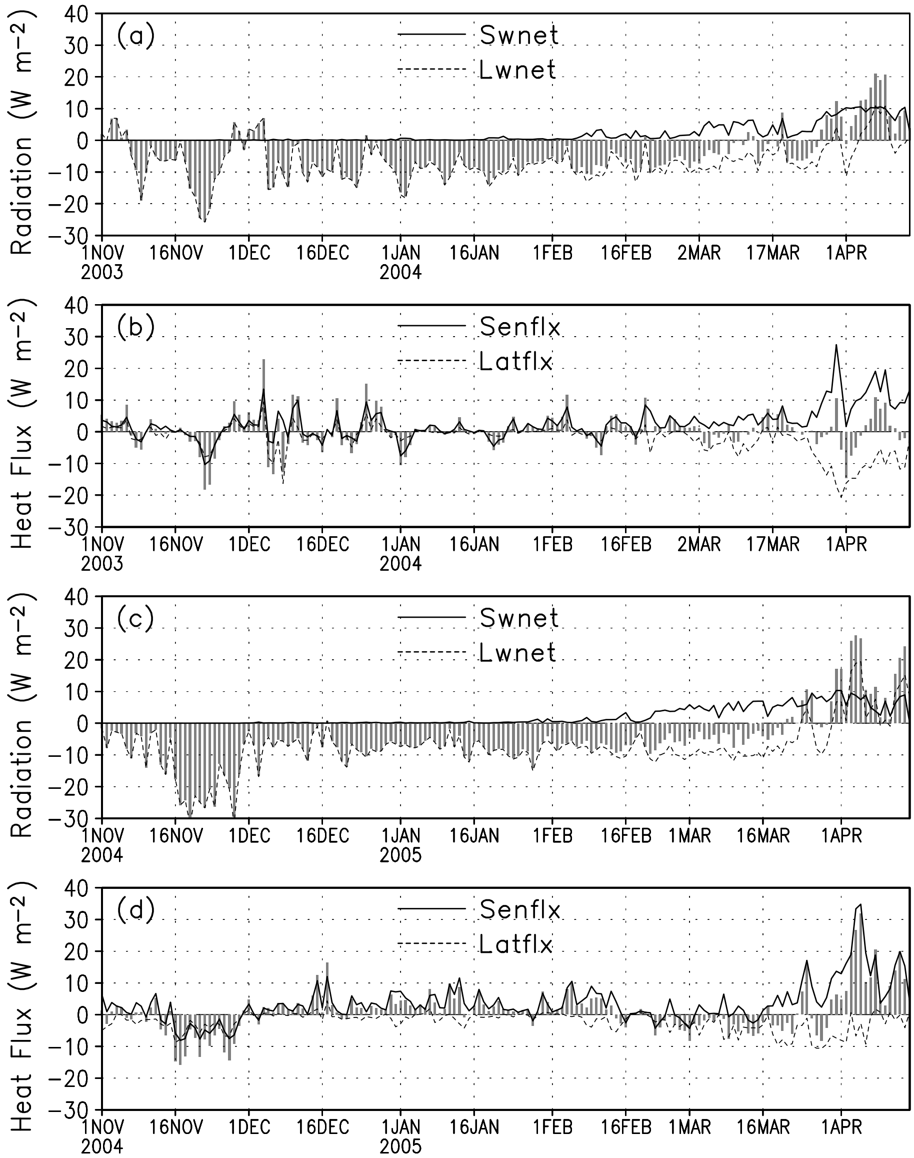

Fig. 3. Temporal evolution of simulated daily mean snow surface energy budget (W m^{-2}) in (a, b) 2003–2004 and (c, d) 2004–2005 snow seasons. (a), (c) radiation and (b), (d) turbulent heat fluxes. Grey bars in each panel represent the sum of the two components. Positive and negative values represent energy flux towards and away from the snow surface, respectively. The same for Figs. 4, 5, 6 concerning the energy flux direction.

the largest contribution of snow falling to the ground beneath canopy (Figs. 2a, 2c). Therefore, the maximum water (snow) holding capacity is a very important parameter for the simulation of the under-canopy snow process. BATS assumes the same maximum holding capacity (0.1 mm) for both liquid water and snow, which considerably underestimates the interception of snow by the canopy (Niu and Yang, 2004). In this study we assume the same amount when the atmospheric temperature is above 1.0°C but take 0.3 mm when the atmospheric temperature is below 1.0°C. Of every month of the two snow seasons that experienced snowfall, most was recorded between early December 2004 and January 2005 (Fig. 2c). The contrast of rainfall and snowfall interception is obvious because of the different maximum holding capacities of the canopy. Fluctuations in the canopy-intercepted snow indicate varying weather conditions (Figs. 2b, 2d).

Daily evolution of the surface snow layer energy budget is shown in Fig. 3, with positive values indicating energy fluxes downward to the snow surface involved in heating the snowpack. In the snow season of

2003–2004, net solar radiation is negligible until early February, then gradually intensifies, with sporadic low radiations in late March associated with cloudy conditions. It should be pointed out that net solar radiation in this study is referred to as the solar radiation absorbed by the top layer in the SAST scheme which excludes the part of the visible solar radiation transmitted through the top snow layer, it is therefore available for heating the top snow layer. Net longwave radiation, in general, cools down the snow surface, except for several heating periods that contribute to snow melts which are discussed later. The total net radiation is dominated by net longwave radiation before mid-March, 2004. However, net solar radiation is comparable to or more than the net longwave radiation from mid-March onward (Fig. 3a). The turbulent sensible and latent heat flux exchanges between the canopy and the snowpack fluctuate with variations in temperature and vapor pressure differences between the canopy and the snow surface except for the weak turbulent heat exchanges in January, 2004. Sensible and latent heat fluxes act in a downward direction to heat the snow surface intermittently before mid-February, 2004. Thereafter, downward sensible heat flux warms the snow surface while evaporation cools it off. Condensation of atmospheric water vapor and its associated heating are conducive to the snow melting process between late November and early December, 2003. The sensible heat flux from the warmer canopy to the snow competes with snow surface evaporation from late February onwards and dominates the overall turbulent heat flux intermittently (Fig. 3b).

The temporal evolutions of net radiations and turbulent heat fluxes in the 2004–2005 snow season are similar to the discussion above, except that net longwave cooling is dominant until late March and net longwave heating dominates after April 2005 (Fig. 3c). Sensible heating dominates over latent cooling during most part of the 2004–2005 snow season, excluding late November, 2004 when turbulent cooling occurs, and the period from late February to mid-March when evaporative cooling is slightly stronger (Fig. 3d).

In the following paragraphs, we will select some specific periods, including early winter snow melting and fast snow ablation in mid-spring, to demonstrate the energy balance characteristics at different stages of the snow season.

In the early stage of the 2003–2004 snow season, net solar radiation is negligible, and the total net radiation is dominated by net longwave radiation (Fig. 4). Longwave radiation from the relatively warmer canopy to the colder snow surface heats the snowpack, except for two days in November (29th and 30th) when cooling occurred through net longwave radiation and strong advection of cold air led to more than six degrees cooling within 24 hours (Fig. 4c). Turbulent sensible heating and condensation heating contribute more than the total radiation to snow melting on 28 November, around noon of 29 November, and during the daytime on 1 December (Fig. 4b). For the 2004–2005 snow season, net longwave radiation contributes a small part to snow melting between 15 and 17 December (Fig. 4d) but total turbulent heat fluxes contribute the largest part during the middle of December 2004 (Fig. 4e). In short, the turbulent heat flux under the canopy is the dominant energy source for snow melting during early winter. A relatively large amount of turbulent heat flux is associated with the combination of high air temperature, high humidity, and strong wind (Figs. 4c, 4f). More sensible heat in the 2004–2005 snow season is mostly attributed to stronger winds than its counterpart in 2003–2004. These results are in line with Marks et al. (1998) who emphasized the importance of the turbulent sensible and latent transfers during winter rain-on-snow events. This along with the combination of warm air temperatures, strong winds and high humidities can cause significant condensation on the snow surface, providing adequate energy for rapid melting of the snow cover. Essery et al. (2003) also found the large fraction of downward sensible heat flux in the under-canopy snow surface energetics, although the atmospheric conditions under the boreal forest canopy are often stable during snow melt season.

During the melting period in spring, solar radiation is rather stable and net longwave heating during daytime is fairly strong (Fig. 5). The downward sensible heat flux is comparable to or stronger than solar radiation, and daytime cooling due to snow evaporation is also significant (Figs. 5b, 5e). By comparing with the aforementioned winter snow melting period, more downward net longwave radiation and sensible heat can be attributed to the warmer atmosphere and warmer canopy. This results in a larger temperature difference between the air within the canopy and the underlying snow surface, since wind speed in spring is comparable to or weaker than its wintertime counterpart (Figs. 5c, 5f). The total net radiative heating along with the downward sensible heat flux during the daytime causes the quick melting of the snowpack. Net longwave radiation contributes slightly more than net solar radiation, while downward sensible heating dominates the turbulent heat flux of the snow surface, especially during the daytime.

Blyth et al. (1994) provided evidence of the mechanism for maintaining negative sensible heat flux over a forest which is capable of sustaining wet canopy evaporation when the radiation is low. The main condition

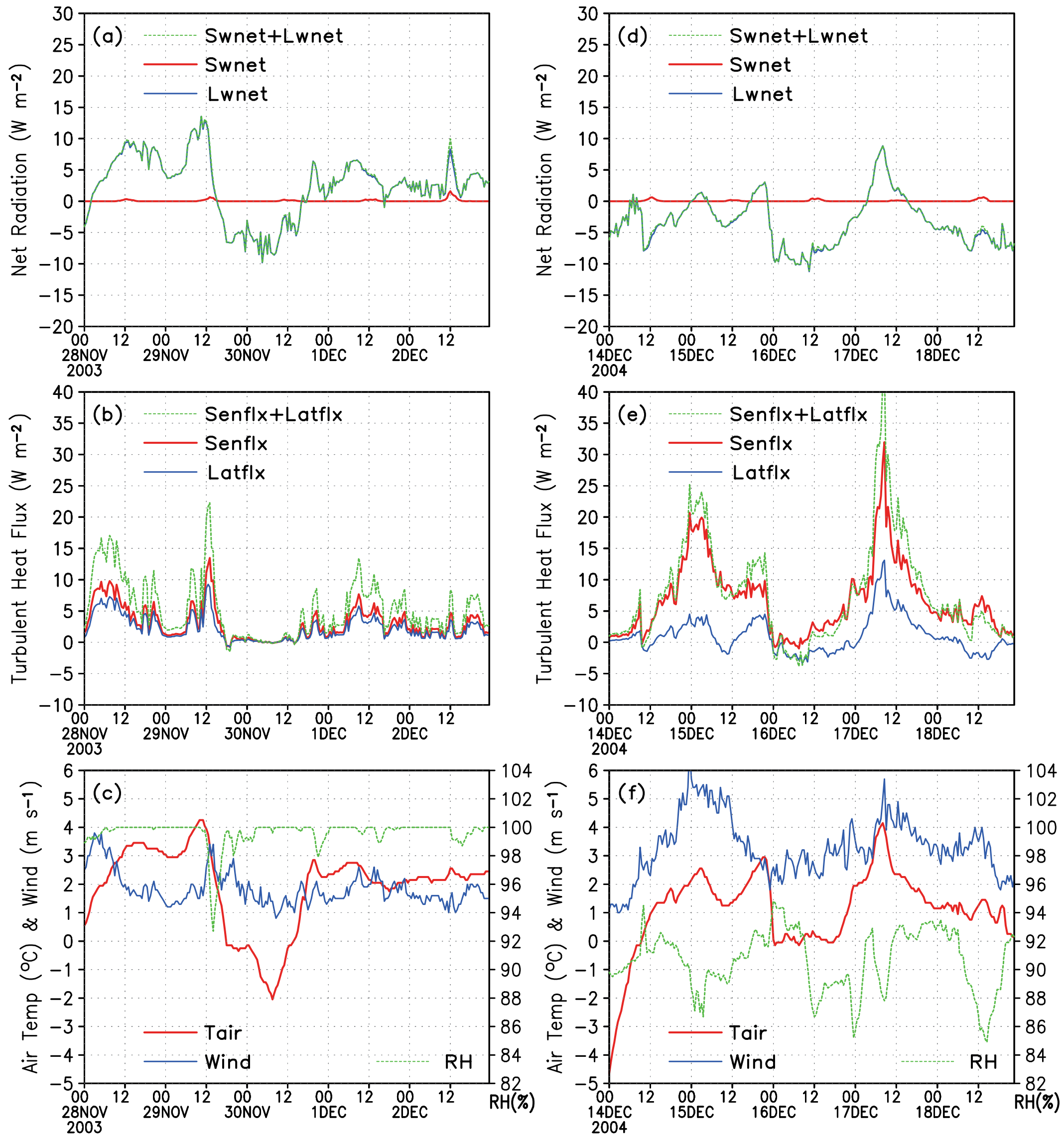

Fig. 4. Temporal evolution of snow surface energy budget components (W m^{-2}) in each 30-minute simulation period and the concurrent atmospheric forcing of temperature (°C), wind speed (m s^{-1}), and relative humidity (percent) during the early winter melting events in snow seasons of (a, b, c) 2003–2004 and (d, e, f) 2004–2005. (a), (d) net radiation, (b), (e) turbulent heat fluxes, and (c), (f) atmospheric forcing.

needed for this mechanism to occur is a strong wind shear within a stable, stratified boundary layer. Such conditions are likely to occur when relatively strong winds in the boundary layer are associated with a wet surface with a high aerodynamic roughness, such as forests. This mechanism for downward transport of heat can partly explain the aforementioned downward sensible heart fluxes to the under-canopy snowpack when the canopy is much warmer than the snow surface.

In order to study the influence of canopy density on snow processes, we conducted sensitivity studies by increasing or decreasing LAI by 1.0 for the 2004–2005 snow season. We found that the impact of net solar radiation increases with lowering LAI, especially after spring when the incident solar radiation increases with solar angle (Fig. 6). The effect of net longwave radiation decreases more than the increase in net solar radiation during both winter and early spring due to the lower canopy temperature resulting from less solar radiation being absorbed by the sparser canopy. For this reason, the total net radiation decreases when LAI becomes smaller and vice versa (Figs. 6a–c). Towards the end of the snow ablation season (after mid-April), net solar radiation exceeds net longwave radiation emanating from the sparser canopy, resulting in more net

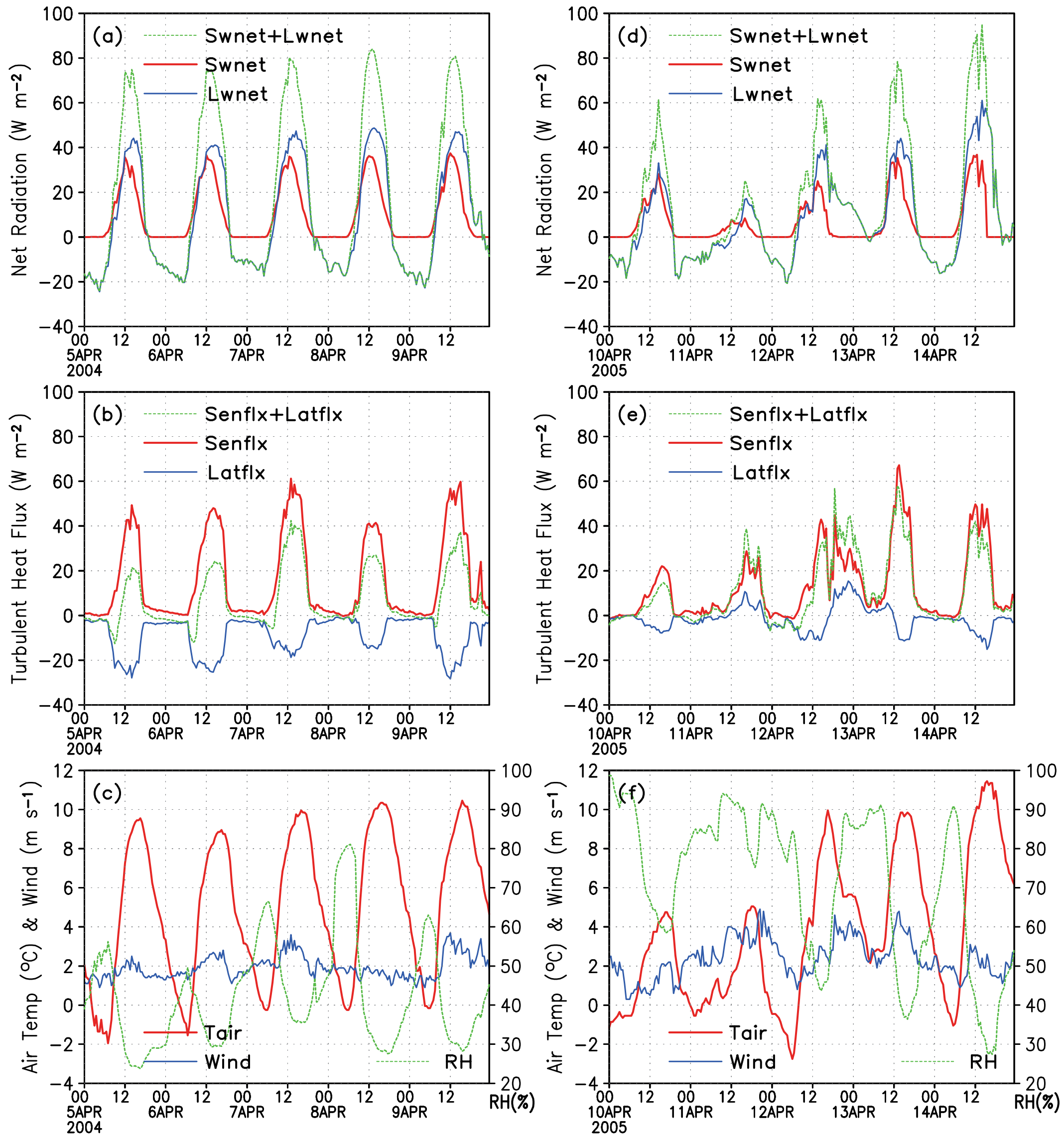

Fig. 5. As in Fig. 4, but for the quick ablation periods in mid-spring.

all-wave radiation from that point on. The strong fluctuation around mid-April is caused by the sharp contrast between the numerical experiments where snow exists in one experiment but disappears in the other. Since the atmospheric temperature/humidity near the snow surface decreases more than does the canopy air temperature/humidity as LAI decreases (not shown), stronger downward sensible and latent heating (or weaker cooling) results in more downward turbulent heat fluxes (Figs. 6d–f). The decrease in net all-wave radiation surpasses the increase in downward turbulent heat flux associated with lower LAI for most of the time, but when the solar angle increases and incident solar radiation intensifies by mid-April, more solar radiation transferring through the sparser canopy dominates the snow surface energetics. Both downward net all-wave radiation and total turbulent heat flux increase with lower LAI around the final snow ablation period in middle April (Figs. 6c, 6f). One aspect that needs to be pointed out is the asymmetry in the sensitivities of energy components to changes in LAI. Less LAI shows larger sensitivity, which is in line with the results of Yang et al. (1999). The reason for this is both the attenuation of solar radiation and the canopy emissivity of longwave radiation increase exponentially rather than linearly with LAI.

Other influences of the canopy concern interception and throughfall of snow to the ground. It is reasonable that the canopy-intercepted snow and subsequent drop off increases with LAI (Fig. 7), but this effect is outweighed by less direct throughfall of snowfall through the denser foliage (Fig. 7c). Therefore the overall ef-

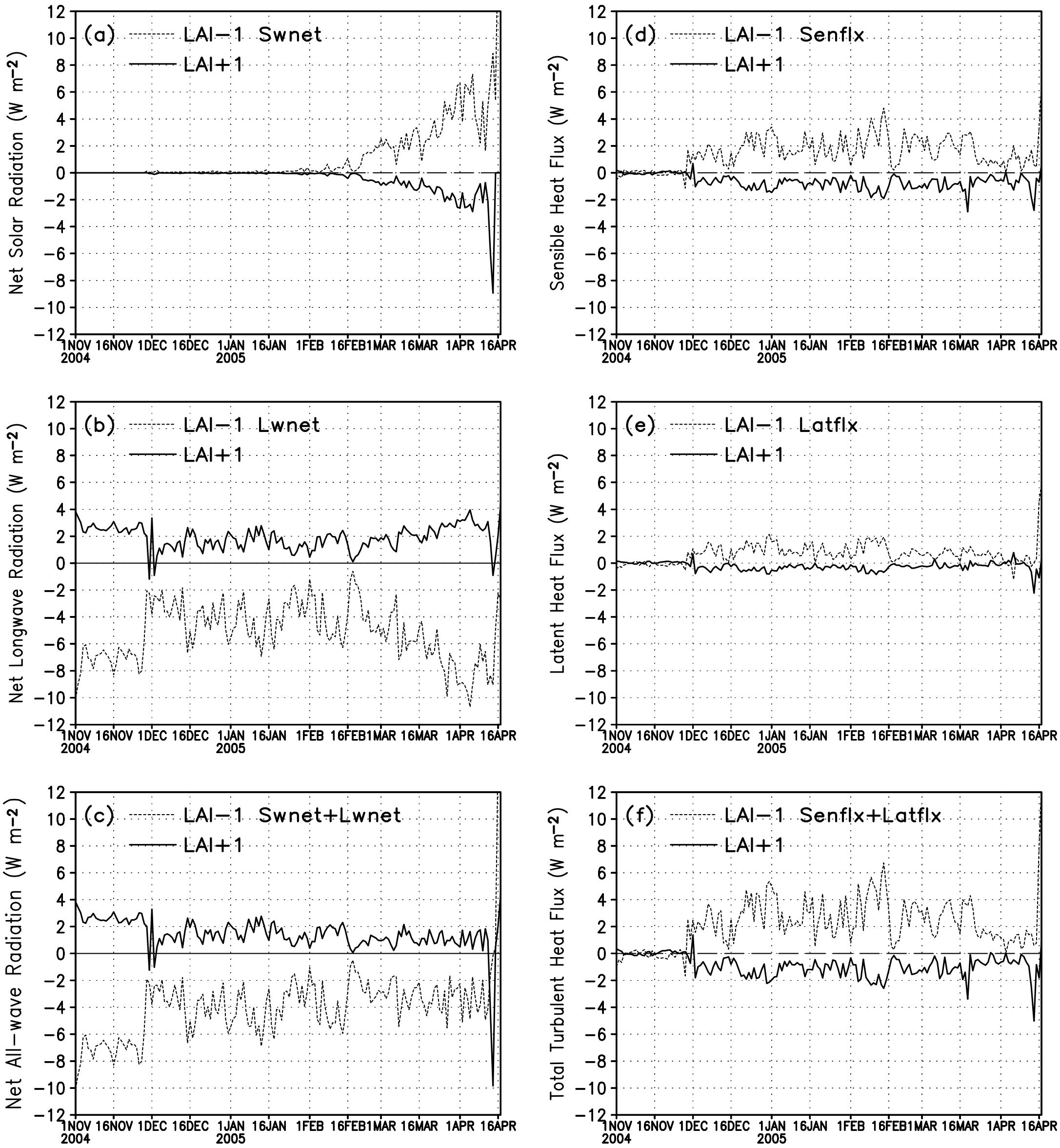

Fig. 6. Sensitivities of daily mean snow surface energy budget components (W m^{-2}) to canopy density in the 2004–2005 snow season. (a) net solar radiation, (b) net longwave radiation, (c) net all-wave radiation, (d) sensible heat flux, (e) latent heat flux, and (f) total turbulent heat flux.

fect of foliage density results in less snow falling to the ground as LAI increases, and vice versa (Fig. 7d). The snowpack melting under the canopy generally increases in winter but decreases in spring when LAI increases, and vice versa (Fig. 7e), which corresponds to changes in available energy for melting (Figs. 6c, 6f). This is similar to the results of Tribbeck et al. (2004), whereby they proposed that an increase in the forest density increased snow melt early and in the middle of the season but decreased the melt rate late in the season. Under-canopy snow accumulation represented by snow water equivalent (SWE) increases (decreases) as LAI decreases (increases) (Fig. 7f). In the sensitivity experiment when LAI decreases by 1.0, 29 mm more SWE is simulated peaking around 12 April, in which accumulated snow falling to the ground increases by about 11.6 mm due to more direct throughfall, and integrated snowmelt decreases by about 17.4 mm mostly due to less thermal radiation from the sparser canopy (Fig. 6b). In the sensitivity experiment as LAI increases by 1.0, accumulated snow falling to the ground decreases by nearly 8.5 mm, but integrated snowmelt increases by about 7 mm, therefore the simulated SWE is 15.5 mm less than that in the control experiment.

Taking into account the above discussions of energy budget what materializes is the slower ablation of

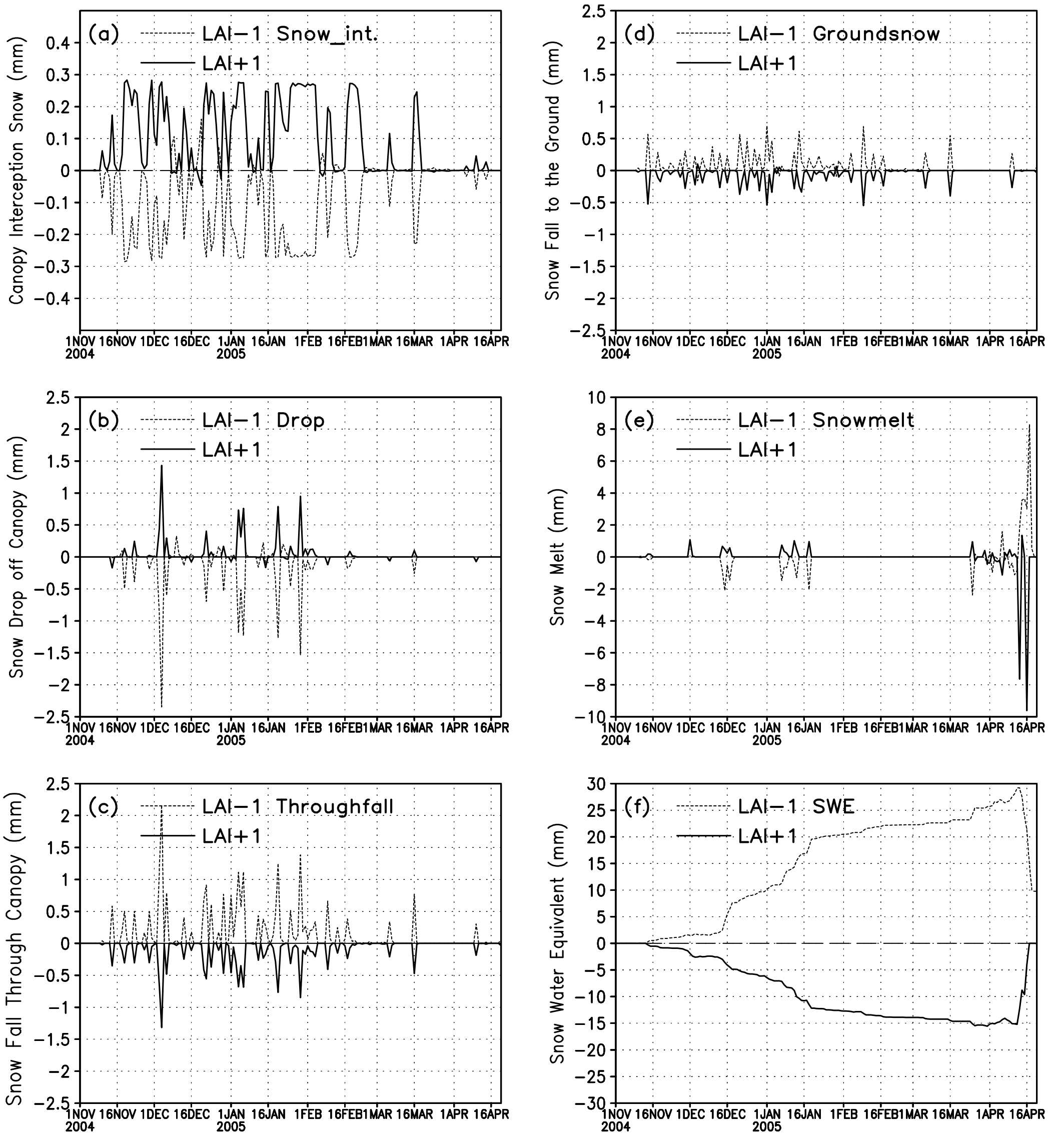

Fig. 7. Daily mean sensitivities to leaf area index of (a) interception of snow by canopy, (b) drop-off from the canopy-intercepted snow, (c) snow falling through the canopy, (d) total amount of snow falling to the ground, (e) loss of snow due to melting, and (f) ground snow water equivalent. Units in mm.

ground level snow beneath the sparser canopy (Fig. 8), since less energy is available for snow melting, while at the same time more snow is falling to the ground due to lower LAI. Snow depth can be 0.07 m higher when LAI decreases by 1.0, and the final ablation of snow is on 23 April, six days later than the control experiment. The sensitivity of snow process to denser foliage is less significant, the maximum difference in simulated snow depth between the sensitivity experiment with LAI increasing by 1.0 and the control experiment is less than 0.05 m. The final ablation of snow is two days earlier than that in the control experiment, 15 April versus 17 April. The parallel LAI sensitivity experiments in the 2003–2004 snow season are much less significant. As LAI decreases by 1.0, the maximum difference in simulated SWE is about 0.05 m, whereas the final ablation of snow is only one day later than in the control experiment; as LAI increases by 1.0, the melting away of snow occurs at almost the same time as in the control experiment (not shown). This indicates the effect of modulation of interannual variations of dominant climate conditions on the impact of foliage density on the under-canopy snowpack.

4. Discussion

Short wave radiation that is absorbed by branches and foliage is emitted as longwave radiation partly downward towards the snow cover. At high latitudes,

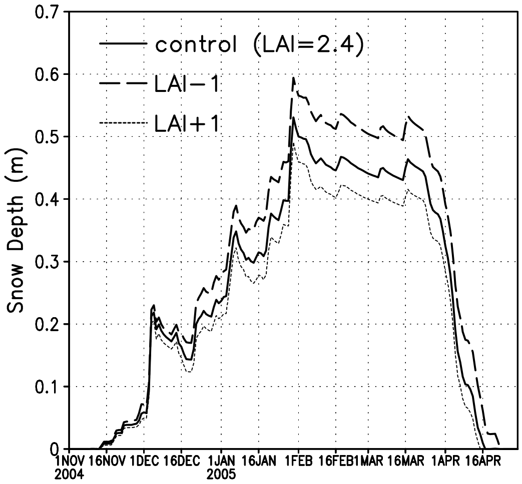

Fig. 8. Sensitivities of under canopy snow depth (m) to leaf area index.

the wintertime solar angle is low, and the incoming shortwave radiation is also low especially under cloudy conditions. In addition, when the vegetation density is high, solar radiation transmitted through the canopy is small, and the albedo of the under-canopy snow is high. Under all the above conditions, the net solar radiation absorbed by the under-canopy snow surface is negligible (Yang et al., 1997). The daily mean net radiation at the snow cover surface becomes positive when the daily mean solar angle exceeds 22°N in late March for a Jack pine forest in Canada (Pomeroy and Dion, 1996). It is reasonable that solar radiation contributes little to the snowmelt energy in this study. Since Hyytiala is at such high latitude, the daytime is short and solar radiation is very weak in winter. However, net radiation fluxes dominate the snowmelt energetics in spring when solar radiation intensifies with the season evolution. In previous studies, the estimation of the total solar radiation downward to the under-canopy snowpack using Beer's law couldn't distinguish the visible part which can transmit through the snowpack from the near infrared part which was almost totally absorbed by the top layer. This usually underestimated solar radiation available for snow melting at the top layer because of the underestimation of the incident near infrared solar radiation. Considering the strong interannual variations of climate regimes along with solar angle in determining incident solar radiation, more cases from different years and at other boreal forest sites at different latitudes are needed to draw general conclusions.

Observations on radiation budgets under the canopy are necessary to validate the calculation of radiation transmission through the vegetation canopy. Turbulent heat flux observations are also needed to evaluate the parameterizations in the model. Although a warmer canopy overlying the melting snow forms a stable condition beneath the canopy, the great temperature and moisture gradients under the canopy can maintain the downward under-canopy turbulences that are comparable to or greater than the net radiation. Different schemes for calculating both radiation transmission through the canopy and turbulent heat fluxes under the canopy and comparisons with corresponding observations are further objectives of this study.

Yang et al. (1999) pointed out that since the averaged canopy temperature is usually higher than the canopy bottom temperature, the one layer canopy in the BATS model may overestimate the canopy temperature and, in turn, overestimate the downward sensible and latent heat fluxes to the snow surface. They further suggested a stronger sensitivity to changes in LAI when LAI is small, and a less sensitivity when LAI is large. This corresponds with the smaller significance of the simulations conducted in this study when LAI becomes higher.

Advection of heat from the surrounding snow-free terrain and vegetation can account for 60% of the energy available for snowmelt (Zuzel and Cox, 1979). The model in this study does not explicitly account for the heat advection from surrounding snow-free areas, but it implicitly accounts for this in the sensible heat, which uses the measured air temperature and wind speed. Errors in the estimation of snowfall amount may also contribute to the uncertainty in snow depth simulation.

It should be pointed out that the simulations are dependent on the performance of the model; although SAST is a reliable snow model, comparative studies using different models may help us understand the vegetation influence on snow accumulation and ablation with more confidence.

5. Concluding remarks

This study employed a physically-based multi-layer snow scheme SAST and a land surface model BATS to investigate how the boreal forest influences snow accumulation and ablation under a high-latitude Scots pine canopy in Finland. Snowpack mass balance and energetics of the snow surface beneath the canopy at different stages of two snow seasons (2003–2004, 2004–2005) are analyzed. For the fairly dense Scots pine forest studied, drop-off of canopy-intercepted snow contributes twice as much or more than direct throughfall of snow to the snow accumulation on the ground.

At different stages of the snow season, the energy available for snow melting beneath the canopy is different. During early winter snow melting, downward turbulent sensible and condensation heat fluxes play the dominant role, with downward net longwave radiation also contributing partly to the snow melting. All energy components are associated with sporadic rainfall events, warm air temperature and high humidities, indicating the importance of weather conditions in the snowmelt of early winter.

In the final stage of snow melting, net solar heating is stable and fairly strong, but downward net thermal radiation of the canopy is slightly stronger than solar radiation, partly due to extinction of solar radiation by the foliage and warming up the canopy. Although the sensible heat flux is comparable to or slightly larger than solar radiation, evaporative cooling of the melting snow surface associated with daytime high atmospheric temperature and low relative humidity makes the overall turbulent heat flux weaker than net radiation. It is the net radiation that dominates the energetics of the snow surface during the final ablation of snowpack.

Snowmelt during the early part of the season may speed up under denser canopies, but snow melting later in the snow season is suppressed. Higher LAI increases the interception of snowfall, therefore reduces snow accumulation under canopy during the snow season; this impact along with the enhancement of thermal radiation outweighs the greater attenuation of solar radiation by denser foliage, resulting in earlier snow ablation under denser canopy. In contrast, more direct throughfall of snow to the ground and less thermal radiation with a sparser canopy leads to more snow accumulation and later ablation of snowpack in a high latitude boreal forest environment. The interannual variations of climate regimes can modulate the impact of canopy density on underlying snow processes, making the sensitivities of the snowpack to canopy density fluctuate with climate conditions in different snow seasons.

Acknowledgements. The authors are grateful to Profs. Sun Shufen and Ji Jinjun of LASG/IAP, Chinese Academy of Sciences and Prof. Yongkang Xue of University of California at Los Angeles for their valuable comments and suggestions throughout this study. Discussions with Jiming Jin of Lawrence Berkley National Laboratory and Zong-Liang Yang and Guo-Yue Niu of Unversity of Texas at Austin have been very helpful. Thanks go to Timo Vesala and Jukka Pumpanen of University of Helsinki, Finland and Richard Essery of University of Wales, UK who provided the meteorological and snow depth data at the Hyytiälä forest site. Editing helps from Peter Meyers at New Jersey Department of Environmental Protection greatly improved the English readability of the manuscript. Funding was provided by the Ministry of Science and Technology of China under Grant Nos. 2006CB403600 and 2005CB422000, and the climate system model development project of the China Meteorological Administration.

REFERENCES

Blyth, E. M., A. J. Dolman, and J. Noilhan, 1994: The effect of forest on mesoscale rainfall: An example from HAPEX-MOBILHY. *J. Appl. Meteor.*, **33**, 445–454.

Chen, J.-M., P. M. Rich, S. T. Gower, J. M. Norman, and S. Plummer, 1997: Leaf area index of boreal forest: Theory, techniques, and measurements. *J. Geophys. Res.*, **102**, 29429–29443.

Dai, Y-J., and Q.-C. Zeng, 1997: A land surface model (IAP94) for climate studies, I: Formulation and validation in off-line experiments. *Adv. Atmos. Sci.*, **14**, 433–460.

Dickinson, R. E., A. Henderson-Sellers, P. J. Kennedy, and M. F. Wilson, 1986: Biosphere Atmosphere Transfer scheme (BATS) for the NCAR Community Climate Model. NCAR Technical Note, NCAR/TN275+STR, 69pp.

Dickinson, R. E., A. Henderson-Sellers, and P. J. Kennedy, 1993: Biosphere Atmosphere Transfer scheme (BATS) version 1e as coupled for Community Climate Model. NCAR Technical Note, NCAR/TN-378+STR, 72pp.

Essery, R., J. Pomeroy, J. Parviainen, and P. Storck, 2003: Sublimation of snow from conifer forests in a climate model. *J. Climate*, **16**, 1855–1864.

FitzGibbon, J. E., and T. Dunne, 1983: Influence of subarctic vegetation cover on snowmelt. *Physical Geography*, **4**, 61–70.

Gelfan, A. N., J. W. Pomeroy, and L. S. Kuchment, 2004: Modeling forest cover influences on snow accumulation, sublimation, and melt. *Journal of Hydrometeorology*, **5**, 785–803.

Golding, D. L., and R. H. Swanson, 1986: Snow distribution patterns in clearings and adjacent forest. *Water Resour. Res.*, **22**, 1931–1940.

Jin, J., X. Gao, Z-L. Yang, R. C. Bales, S. Sorooshian, R. E. Dickinson, S.-F. Sun, and G-X. Wu, 1999: Comparative Analyses of Physically Based Snowmelt Models for Climate Simulations. *J. Climate*, **12**, 2643–2657.

Link, T. E., and D. Marks, 1999: Point simulation of seasonal snow cover dynamics beneath boreal forest canopies. *J. Geophys. Res.*, **104**(D22), 27841–27857.

Liston, G., and K. Elder, 2006: A meteorological distribution system for high-resolution terrestrial modeling (MicroMet). *Journal of Hydrometeorology*, **7**, 217–234.

Lundberg, A., and S. Halldin, 1994: Evaporation of intercepted snow: Analysis of governing factors. *Water Resour. Res.*, **30**, 2587–2598.

Marks, D., J. Kimball, D. Tingey, and T. Link, 1998: The sensitivity of snowmelt processes to climate conditions and forest cover during rain-on-snow: A case study of the 1996 Pacific Northwest flood. *Hydrological Processes*, **12**, 1569–1587.

Nakabayashi, H., N., Ishikawa, and Y. Kodama, 1999: Radiative characteristics in a Japanese forested drainage basin during snowmelt. *Hydrological Processes*, **13**, 157–167.

Niu, G.-Y., and Z.-L. Yang, 2004: Effects of vegetation canopy processes on snow surface energy and mass balances. *J. Geophys. Res.*, **109**, D23111, doi:10.1029/2004JD004884.

Oleson, K. W., and Coauthors, 2004: Technical description of the Community Land Model. NCAR Tech. Note NCAR/TN-461+STR, 174pp.

Pomeroy, J. W., and K. Dion, 1996: Winter radiation extinction and reflection in a boreal pine canopy: Measurements and modeling. *Hydrological Processes*, **10**, 1591–1608.

Sicart, J. E., J. W. Pomeroy, R. L. H. Essery, J. Hardy, T. Link, and D. Marks, 2004: A sensitivity study of daytime net radiation during snowmelt to forest canopy and atmospheric conditions. *Journal of Hydrometeorology*, **5**, 774–784.

Sun, S.-F., J.-M. Jin, and Y.-K. Xue, 1999: A simple snow-atmosphere-soil transfer model. *J. Geophys. Res.*, **104**, 19587–19597.

Sun, S.-F., and Y.-K. Xue, 2001: Implementing a new snow scheme in simplified simple biosphere model. *Adv. Atmos. Sci.*, **18**(3), 335–354.

Suzuki, K., T. Ohta, A. Kojima, and T. Hashimoto, 1999: Variations in snowmelt energy and heat balance characteristics with larch forest density on Mt. Iwate, Japan: Obervations and heat balance analyses. *Hydrological Processes*, **13**, 2675–2688.

Tribbeck, M. J., R. J. Gurney, E. M. Morris, and D. W. C. Pearson, 2004: A new Snow-SVAT to simulate the accumulation and ablation of seasonal snow cover beneath a forest canopy. *Journal of Glaciology*, **50**, 171–182.

Vesala, T., and Coauthors, 1998: Long-term field measurements of atmosphere-surface interactions in boreal forest combining forest ecology, micrometeorology, aerosol physics and atmospheric chemistry. *Trends in Heat, Mass and Momentum Transfer*, **4**, 17–35.

Vesala, T., and Coauthors, 2005: Effect of thinning on surface fluxes in a boreal forest. *Global Biogeochemical Cycles*, **19**, GB2001, doi:10.1029/2004GB002316.

Xue, Y-K., S-F. Sun, D. S. Kahan, and Y-J. Jiao, 2003: Impact of parameterizations in snow physics and interface processes on the simulation of snow cover and runoff at several cold region sites. *J. Geophys. Res.*, **108**(D22), 8859, doi:10.1029/2002JD003174.

Yamazaki, T., and J. Kondo, 1992: The snowmelt and heat balance in snow-covered forested areas. *J. Appl. Meteor.*, **13**, 1322–1327.

Yang, Z-L., R. E. Dickinson, A. Robock, and K. Y. Vinikov, 1997: Validation of the snow sub-model of the biosphere-atmosphere transfer scheme with Russian snow cover and meteorological observation data. *J. Climate*, **10**, 353–373.

Yang, Z-L., Y. Dai, R. E. Dickinson, and W. J. Shuttleworth, 1999: Sensitivity of ground heat flux to vegetation cover fraction and leaf area index. *J. Geophys. Res.*, **104**(D16), 19505–19514.

Zuzel, J. F., and L. M. Cox, 1979: Ablation of an isolated snowdrift. *Proc. Modeling of Snow Cover Runoff*, Hanover, NH, U. S. Army Cold Regions Research and Engineering Laboratory, 135–143.