The snowmelt and heat balance in snow-covered forested areas

The Snowmelt and Heat Balance in Snow-covered Forested Areas

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ABSTRACT

The snowmelt and heat balance in snow-covered forested areas have been studied with the use of a canopy model. It was found that, in general, as the canopy density increased the snowmelt decreased. However, with conditions of high air temperature, weak winds, and large snow albedo, a greater degree of snowmelt occurred under a dense canopy due to infrared radiation from the canopy elements than under a sparse canopy. Although the snow temperature was never higher than $0°C$ while the air temperature was greater than $0°C$, an upward sensible heat flux was supplied from the forest canopy, resulting in the atmospheric heating.

1. Introduction

Snowmelt in forested areas differs from that in an open snow cover with no forest. The energy exchange between the atmosphere and a forested ground surface with snow cover depends on the density of the forest. However, there are few models that take into account the influence of forested areas on the snowmelt. Otterman et al. (1988) investigated the dependencies of albedo, roughness length, and solar-radiation absorption at the snow cover on the structure of the forest. Koike et al. (1985) estimated the amount of snowmelt for cases when the shielding ratio of solar radiation by a forest is known. Recently, Ohta et al. (1990) showed that solar radiation and wind speed are the main meteorological components changed by forest and controlling the snowmelt in a forested area.

In this paper, the dependencies of snowmelt and the heat balance above and within a forest on the meteorological conditions, snow albedo, and forest density are studied by making use of a canopy model (Yamazaki et al. 1992) combined with a snowmelt model (Kondo and Yamazaki 1990).

2. Model

The heat balance of canopy layers and snow cover can be solved given the input parameters of air temperature T, wind speed U_a, specific humidity q, solar radiation S_0, and downward atmospheric radiation L_a. Here, the canopy and snowmelt model used in the present paper will be briefly described.

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a. The canopy model (two-layer model)

The details of this model and its performance have been described in a previous paper (Yamazaki et al. 1992). Figure 1 shows a schematic representation of the model. The canopy is divided into a "crown space" and "trunk space" (without leaves), while the crown space is subdivided into a lower layer ($C1$) and upper layer ($C2$). The heat balance is solved with respect to the radiative, sensible, and latent heat fluxes between the atmosphere and the two crown layers.

The heat-balance equations for each layer are

$$S_{n1} + L_{n1} = H_{C1} + lE_{C1} \qquad (1)$$

and

$$S_{n2} + L_{n2} = H_{C2} + lE_{C2}. \qquad (2)$$

Here, S_{ni} and L_{ni} ($i = 1, 2$) are the net absorption of solar and infrared radiation while H_{Ci} and lE_{Ci} the sensible and latent heat fluxes, respectively, from the canopy elements to the surrounding air in the ith layer.

The flux H_{C1} equals the difference between the upward sensible heat flux (H_{12}) at the top of the $C1$ layer and that (H_S) at the bottom of the $C1$ layer. That is,

$$H_{C1} = H_{12} - H_S. \qquad (3)$$

Note that H_S denotes the sensible heat flux from the snow surface and can be obtained by the snowmelt model as will be shown below. In the same manner as H_{C1},

$$H_{C2} = H - H_{12}, \qquad (4)$$

where H is the sensible heat flux from the top of the canopy layer to the atmosphere. For the latent heat fluxes,
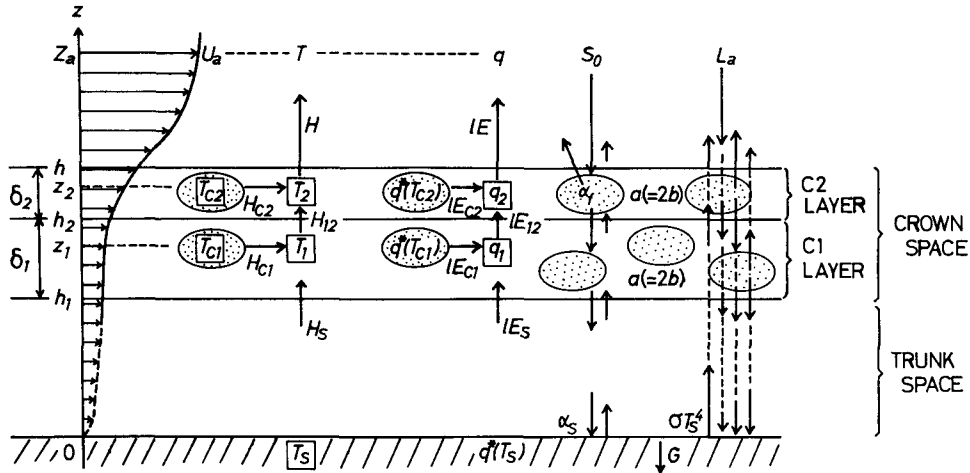
FIG. 1. Schematic diagram of the two-layer model (2LM).

$$lE_{C1} = lE_{12} - lE_S \qquad (5)$$

and

$$lE_{C2} = lE - lE_{12}. \qquad (6)$$

Each term in Eqs. (1)–(6) can be described with canopy parameters (canopy height, leaf-area density, etc.), external conditions (meteorological conditions above the canopy), and unknown variables such as leaf surface temperatures, T_{C1} and T_{C2}, air temperature and specific humidity in each canopy layer T_1, T_2, q_1, and q_2.

The individual terms associated with solar radiation are written as

$$S_{n1} = S^{\downarrow}(h_2) - S^{\downarrow}(h_1) + S^{\uparrow}(h_1) - S^{\uparrow}(h_2), \qquad (7)$$

and

$$S_{n2} = (1 - \alpha'_f)S_0 - S^{\downarrow}(h_2) + S^{\uparrow}(h_2) - S^{\uparrow}(h), \qquad (8)$$

where $S^{\downarrow}(z)$ and $S^{\uparrow}(z)$ are the downward and upward solar-radiation fluxes through a horizontal plane at the height z, respectively, and can be written as follows:

$$S^{\downarrow}(h_2) = (1 - \alpha'_f)X_{m2}S_0,$$

$$S^{\downarrow}(h_1) = (1 - \alpha'_f)X_{m1}X_{m2}S_0,$$

$$S^{\uparrow}(h_1) = (1 - \alpha'_f)\alpha_S X_{m1}X_{m2}S_0,$$

$$S^{\uparrow}(h_2) = (1 - \alpha'_f)\alpha_S X_{m1}X_{m2}X_1 S_0,$$

$$S^{\uparrow}(h) = (1 - \alpha'_f)\alpha_S X_{m1}X_{m2}X_1 X_2 S_0,$$

with

$$\alpha'_f \equiv (1 - X_{m1}X_{m2})\alpha_f.$$

Here, S_0 is the incoming solar radiation at the top of the canopy, α_f the reflectance of individual leaves (assumed to be 0.1), and α_S the snow albedo; X_{mi} and X_i ($i = 1, 2$) are the transmittance of the Ci layer for the

direct solar beam and the reflected radiation, respectively, which are given by

$$X_{mi} = \exp(-m\delta_i b), \quad (m = \sec\theta) \qquad (9)$$

and

$$X_i = \exp(-1.66\delta_i b), \quad (i = 1, 2). \qquad (10)$$

Here, δ_i is the thickness of the Ci layer, θ the solar zenith angle, and b the attenuation coefficient for radiation (assumed to be $a/2$, a: the leaf-area density).

In the same manner as solar radiation, the infrared-radiation terms can be written as

$$L_{n1} = L^{\downarrow}(h_2) - L^{\downarrow}(h_1) + L^{\uparrow}(h_1) - L^{\uparrow}(h_2), \qquad (11)$$

and

$$L_{n2} = L_a - L^{\downarrow}(h_2) + L^{\uparrow}(h_2) - L^{\uparrow}(h). \qquad (12)$$

Here,

$$L^{\downarrow}(h_2) = X_2 L_a + (1 - X_2)\sigma T_{C2}^2,$$

$$L^{\downarrow}(h_1) = X_1 X_2 L_a + X_1(1 - X_2)\sigma T_{C2}^4$$
$$+ (1 - X_1)\sigma T_{C1}^4,$$

$$L^{\uparrow}(h_1) = \sigma T_S^4,$$

$$L^{\uparrow}(h_2) = X_1 \sigma T_S^4 + (1 - X_1)\sigma T_{C1}^4,$$

and

$$L^{\uparrow}(h) = X_1 X_2 \sigma T_S^4 + X_2(1 - X_1)\sigma T_{C1}^4$$
$$+ (1 - X_2)\sigma T_{C2}^4.$$

In the preceding, L_a is the downward atmospheric radiation incident on the canopy, σ Stefan–Boltzmann constant, and T_S the snow surface temperature. The transmittance for infrared radiation X_i is the same as that for the reflected solar radiation [given by Eq. (10)].

The sensible and latent heat fluxes are described that they are proportional to the potential differences of

temperature and specific humidity, respectively. For example, H_{Ci} and lE_{Ci} can be written as

$$H_{Ci} = A_h c_h (T_{Ci} - T_i), \qquad (13)$$

and

$$lE_{Ci} = A_e j c_h [q^*(T_{Ci}) - q_i], \quad (i = 1, 2). \quad (14)$$

Here, A_h and A_e are coefficients that are proportional to the wind speed and leaf-area density, respectively, c_h the heat-transfer coefficient of individual leaves, $j(0 \leqslant j \leqslant 1)$ the evapotranspiration factor of a leaf, and $q^*(T)$ the saturation specific humidity at temperature T.

The wind-speed profile is described by the nondimensional canopy density defined by

$$c_* = c_d a h, \qquad (15)$$

where c_d is the drag coefficient of individual leaves ($=0.2$) and h the canopy height. The value of c_* falls between 0.01 and 0.1 for a leafless deciduous forest, between 0.1 and 1 for a medial density, and greater than 1 for a tall coniferous forest. The effect of atmospheric stability is taken into account above the canopy. To consider free convection, a minimum wind speed is introduced in the crown space (0.05 m s^{-1}).

Since there are six equations for the six unknown variables, these equations can be easily solved to obtain the fluxes after making some approximations (in the original canopy model, the heat-balance equation for the ground surface is necessary, since the ground surface temperature, T_S, is also unknown). In practice, it is difficult to give suitable parameters in the model rather than to compute.

b. The snowmelt model

The details of the snowmelt model were described in Kondo and Yamazaki (1990). Schematic profiles of the snow temperature T and water content W are shown in Fig. 2. The depth Z where the temperature profile intersects $T = 0°C$ represents the freezing depth. The liquid water content W is taken as $W = 0$ ($z < Z$), $W = W_0$ ($z \geqslant Z$), where W_0 is the maximum water content (assumed to be 0.1). Water content is defined as the ratio of the mass of liquid water to the mass of the wet snow.

After an amount of time equivalent to the length of a time step Δt passes, the snow surface temperature T_S changes to T_{Sn} and Z to Z_n. Then the equation for the conservation of energy for the entire snow cover can then be written as (see Fig. 2)

$$\frac{1}{2} C_S \rho_S [Z(T_0 - T_S) - Z_n(T_0 - T_{Sn})]$$

$$+ W_0 \rho_S l_f (Z - Z_n) + M_0 \Delta t = G \Delta t. \quad (16)$$

Here, C_S is the specific heat of snow, ρ_S the snow density (0.3 g cm^{-3}), $T_0 = 0°C$, and l_f the heat of fusion for ice. The first term on the left-hand side of Eq. (16)
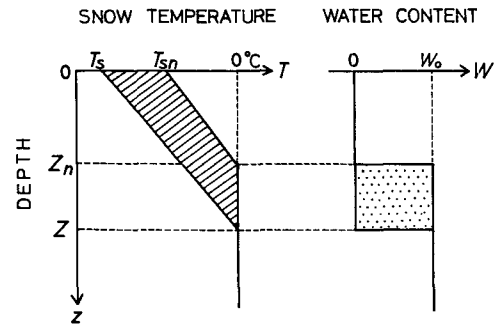
FIG. 2. Schematic diagram showing the snow temperature and water-content profiles of the snowmelt model.

describes the energy associated with the heating or cooling of the snow (the shaded portion in Fig. 2) while the second term accounts for the energy of fusion or freezing of the liquid water remaining in snow cover (the stippled portion in Fig. 2). In the third term, M_0 is the energy required to create runoff from the snow cover in surplus of the maximum water content. On the right-hand side, G represents the energy received by the entire snow cover, and is given by

$$G = S_{nS} + L_{nS} - H_S - lE_S \qquad (17)$$

where $S_{nS} [=S^\downarrow(h_1) - S^\uparrow(h_1)]$ and $L_{nS} [=L^\downarrow(h_1) - L^\uparrow(h_1)]$ are the net solar radiation and net infrared radiation at the snow surface, respectively. The sensible heat H_S and latent heat lE_S are obtained by bulk formulas.

When $Z = Z_n$ and $T_S = T_{Sn}$, the heat-balance equation of a snow surface with an infinitesimal thickness can be written as

$$L_{nS} - H_S - lE_S + \lambda_S \frac{T_0 - T_{Sn}}{Z_n} = 0. \quad (18)$$

The upward thermal conduction through the snow surface is approximated by the last term on the left-hand side of Eq. (18), where λ_S is the thermal conductivity of the snow (assumed to be 0.42 W m^{-1} K^{-1}). Note that since snow transmits solar radiation, Eq. (18) has no terms associated with solar radiation. By solving Eq. (16) (the value of M_0 is assumed to equal zero) and Eq. (18) with bulk formulas for H_S and lE_S, it is possible to simultaneously obtain the snow surface temperature T_{Sn} and freezing depth Z_n. If $Z_n < Z_{min}$ ($=1$ cm), Z_n is set equal to Z_{min}, and if $T_S > 0°C$, it is replaced by $0°C$. The value of M_0 is obtained by the substitution of Z_n and T_{Sn} into Eq. (16).

c. Implementing the coupled model

The practical procedure of the coupled model is as follows.

1) Calculate the wind-speed profile, T_{C1}, T_{C2}, T_1, T_2, q_1, and q_2 using the canopy model with the value

of T_S from the last time step (initially assume neutral atmospheric stability).

2) Calculate the downward solar and infrared radiation at the snow surface.

3) Obtain Z and T_S using the snowmelt model with the value of G from the last time step.

4) Recalculate the wind-speed profile, T_{C1}, T_{C2}, T_1, T_2, q_1, and q_2 (taking into account the atmospheric stability above the canopy) with the new value of T_S.

5) At this point, if T_{C1} and T_{C2} do not converge, return to 2).

6) Calculate the fluxes and snowmelt.

3. Results and discussion

Figures 3a–f show the amount of daily snowmelt calculated for several typical diurnal variations of meteorological conditions, and for various values of snow albedo. The height of the canopy and the reference
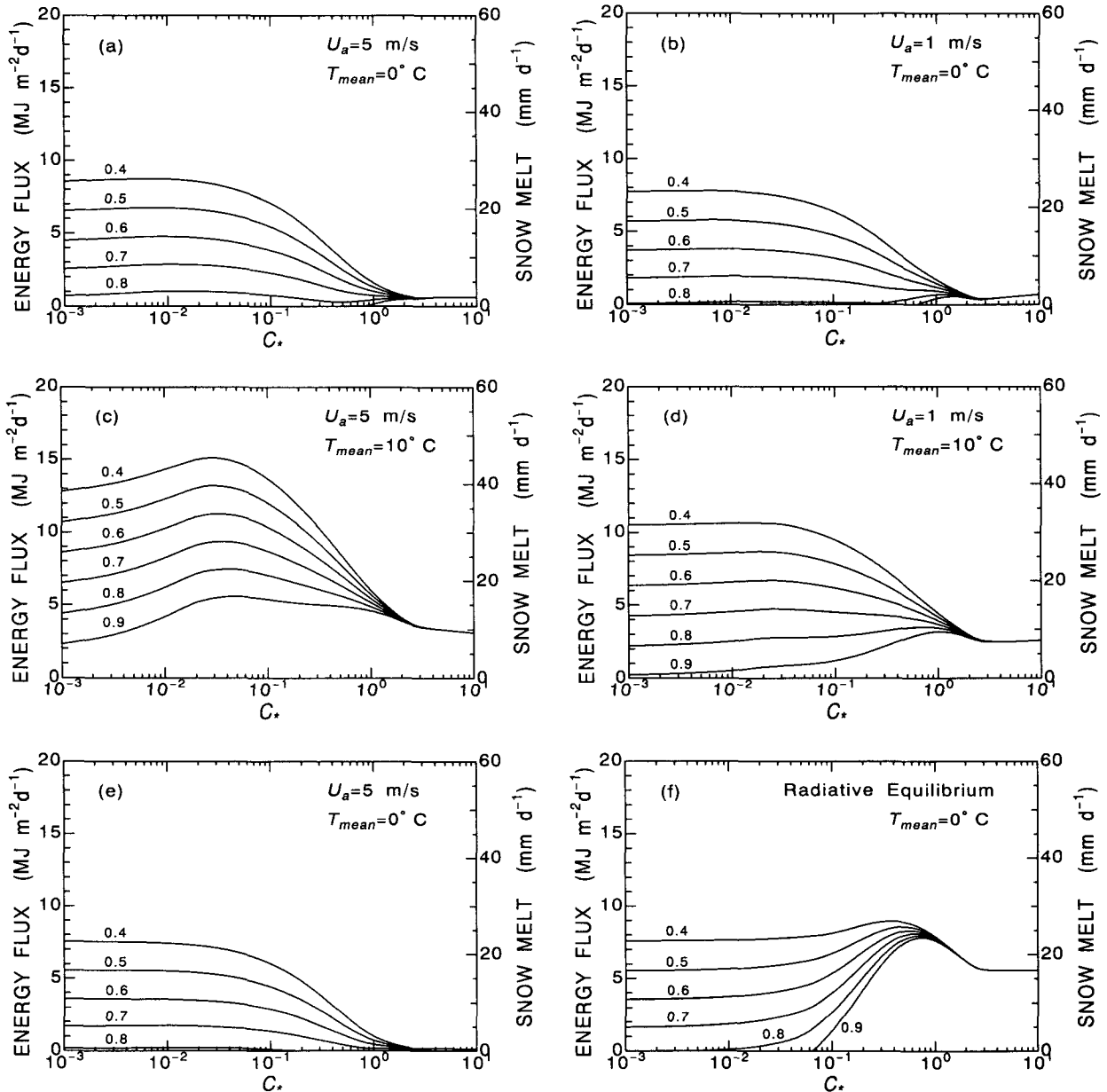

FIG. 3. The amount of daily snowmelt as function of canopy density c_*. Each line corresponds to different values of snow albedo, which are specified at an interval of 0.1 between 0.4 and 0.9. Here U_a is the wind speed at the reference height. (a) The standard conditions: U_a = 5 m s^{-1}, T_{mean} = 0°C. (b) The same as (a) except for weak wind speed (U_a = 1 m s^{-1}). (c) The same as (a) except for high air temperature (T_{mean} = 10°C). (d) High temperature and weak wind. (e) The same as (a) except for no temperature amplitude. (f) Radiative equilibrium condition (c_h = 0 and H_S = lE_S = 0).

height—where the wind speed, air temperature, and specific humidity are given—are set at $h = 5$ m and $z_a = 20$ m, respectively. The reference height involved in snow surface flux is h_1 in Fig. 1 (set at 2.5 m). The diurnal variations of incident solar radiation S_0 and air temperature T are given as graphically displayed in Fig. 4. The maximum value of solar radiation is 800 W m^{-2} as a clear condition. The daily amplitude of air temperature is 5°C except for Fig. 3c, which has no amplitude. The daily mean air temperature T_{mean} is 0°C except for Figs. 3c and 3d, where it is fixed at 10°C.

The wind speed U_a, downward atmospheric radiation L_a, and specific humidity q at the reference height are assumed to be independent of time. The value of L_a is set at 215 W m^{-2} for Figs. 3a, 3b, 3e, and 3f, while at 270 W m^{-2} for Figs. 3c and 3d. The specific humidity q is determined from the minimum relative humidity RH$_{min}$ (set at 50%) and the maximum air temperature. stant, the relative humidity is minimum when the air temperature is maximum value. The evapotranspiration factor of a leaf j is set to 0.1.

For the standard case shown in Fig. 3a, as expected, when the canopy density increases the snowmelt decreases except for cases of large snow albedo. In a sparse canopy, when the air temperature is low, the snowmelt is controlled by the albedo and solar radiation. For a very dense canopy, the snowmelt is independent of the snow albedo, since little solar radiation reaches the snow surface. In this case, snow is melted by infrared radiation emitted from canopy elements (the maximum value of L_{nS} is 62 W m^{-2} when $c_* = 10$). Note that snow albedo depends on specific surface area of snow particles (related to snow density), liquid water content, and impurities, and changes every moment. Therefore, we are studying to parameterize albedo and will include this parameterization in the snowmelt model.

The influence of RH$_{min}$ and j are slight when the air temperature is in the neighborhood of 0°C. If the value of RH$_{min}$ is changed from 50% to 30%, the amount of snowmelt decreases by 0.6 to 1 MJ m^{-2} day^{-1} for a sparse canopy, and by 0.2 MJ m^{-2} day^{-1} for a dense canopy. If the value of j is changed by ±0.1, the snowmelt changes by ∓0.5 MJ m^{-2} day^{-1} for a dense canopy only.

The snowmelt for weak wind speed is shown in Fig. 3b. The result is not much different from that of 3a; however, the amount of snowmelt increases by 1 MJ m^{-2} day^{-1} for a sparse canopy. If the wind speed is equal to 10 m s^{-1}, the result is not different from that of 3a again, however, the amount of snowmelt decreases by 1 MJ m^{-2} day^{-1} for a sparse canopy in this case (the figure not shown). In any case, the wind speed hardly influences the snowmelt in these conditions.

If a cloudy condition is assumed (the maximum value of S_0 is 145 W m^{-2}, $L_a = 305$ W m^{-2}, the other components are the same as Fig. 3a), the snowmelt decreases markedly. The amount of snowmelt is less than 1.5 MJ m^{-2} day^{-1} for a sparse canopy and almost equal to zero for a dense canopy.

When the air temperature is high (see Fig. 3c and 3d), the snowmelt strongly depends on the wind speed in a sparse or medial canopy; the sensible and latent heat are also important. Figure 3d shows that the snowmelt is large in a dense canopy when compared with a sparse canopy for cases of large snow albedo. The reason for this is that, when the canopy is sparse, the absorption of solar radiation—along with sensible and latent heat—is small for high albedo and weak wind conditions. On the other hand, for a dense canopy, the canopy elements are heated, resulting in the snow effectively being melted by infrared radiation emitted from the canopy elements. Forests transform solar radiation into infrared radiation, which in turn can be absorbed even by white snow. In Fig. 3c, the snowmelt decreases in a very sparse canopy, because the sensible and latent heat are limited by strong stable stratification. Although a surface layer is more stable in weak wind condition (Fig. 3d), the snowmelt almost does not decrease in a sparse canopy since the sensible and latent heat are small from the first.

Figure 3e is the same as Fig. 3a except for no temperature amplitude. Although the amount of snowmelt decreases by 0.5 to 1.5 MJ m^{-2} day^{-1}, the influence of the temperature amplitude are not serious.

Figure 3f shows the special case of radiative equilibrium condition (the sensible and latent heat terms are set at zero: $c_h = 0$, $H_S = 0$, and $lE_S = 0$). When the canopy is sparse, the snowmelt is almost the same as in case of 3a and 3b, because the sensible and latent heat are small in 3a and 3b. In a medial or dense canopy, however, the temperature of the canopy elements rises and emits a large amount of infrared radiation and as a result the snowmelt increases.

Figure 4 shows examples of the time series of temperatures and heat-balance components. The meteorological conditions are the same as those in Fig. 3a and the snow albedo is set to 0.6. The calculated values of albedo above the canopy are 0.58 in Fig. 4a and 0.24 in Fig. 4b at noon. The air temperature T and incident solar radiation S_0 are prescribed on the assumption of clear conditions, while the other components (T_2, T_S, R_n, R_{nS}, etc.) are calculated. Although $G > 0$, $M_0 = 0$ during the morning hours, because the energy G is used to dissipate the freezing layer.

When the canopy is very sparse (Fig. 4a), H and H_S are almost identical, since the heat exchange between the air and the canopy elements is negligible. Since the temperature of the snow is never greater than 0°C, the sensible heat H over a snow surface without a forest must be downward when $T > 0$°C. The direction of H is always downward in Fig. 4a, thus the atmosphere is cooled during these conditions.

However, in a canopy of medial dense (Fig. 4b), the direction of H is upward in the daytime and the maximum value H_{max} reaches 240 W m^{-2} (if the value
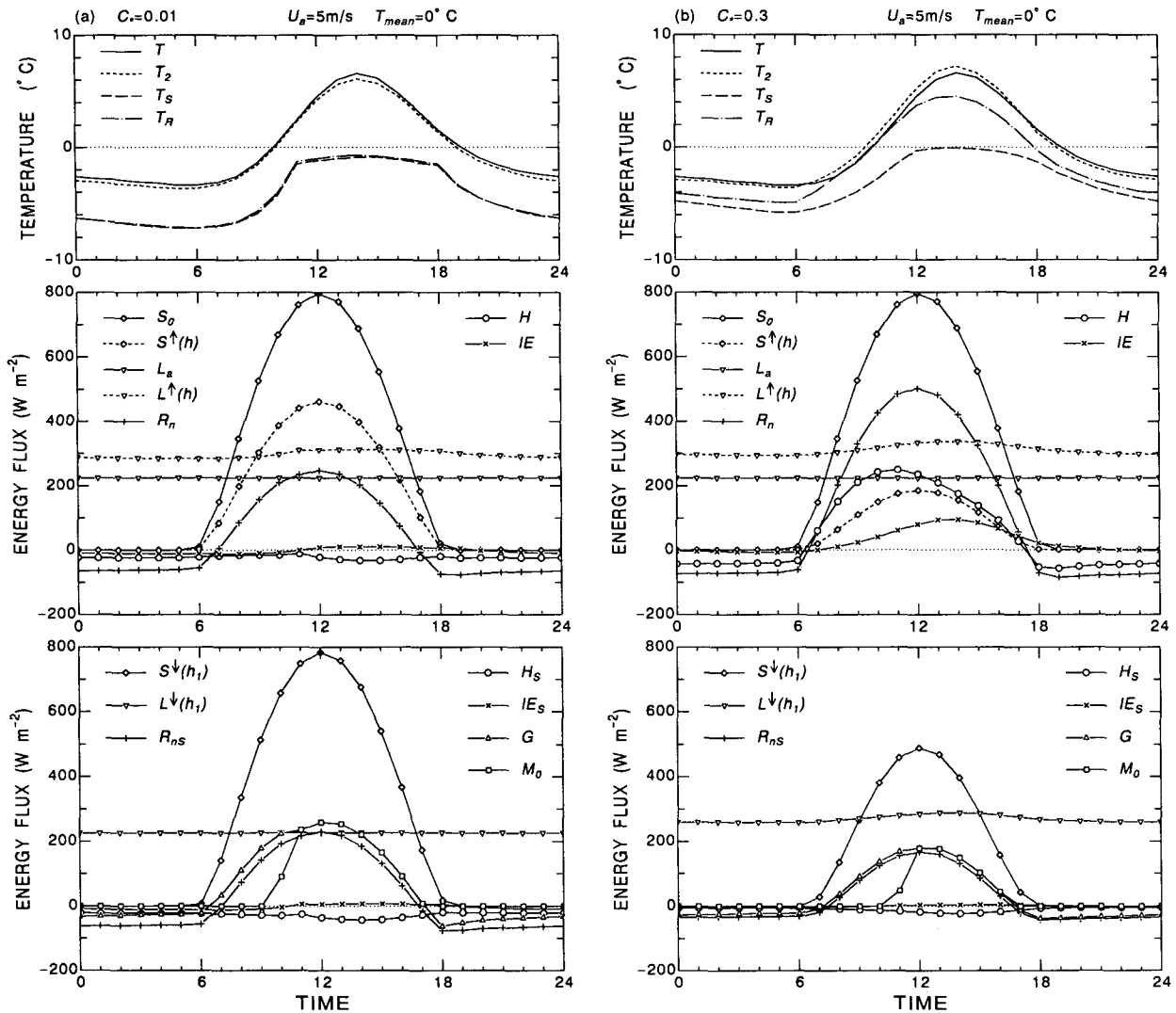
FIG. 4. Examples of the daily march of temperatures and heat-balance components. The variable T_R is the radiative temperature measured from above the canopy, S_0 the incident solar radiation, R_n the net radiation above the canopy, and R_{nS} the net radiation at the snow surface ($= S_{nS} + L_{nS}$). (a) Very sparse canopy ($c_* = 0.01$). (b) Medial dense canopy ($c_* = 0.3$).

of T is held constant at 0°C, $H_{max} = 260$ W m^{-2}). Therefore, the atmosphere is obviously heated, even when daytime air temperatures are greater than 0°C. The snow cover also simultaneously receives sensible heat from the canopy layer. Note that the total latent heat flux lE is provided by the forest transpiration, which depends on evapotranspiration factor j. The suitable value of j is not well known in winter or snow-melt season. It influences the Bowen ratio (H/lE); however, it is not important to snowmelt as mentioned previously. If a canopy is very dense ($c_* = 10$), the maximum values of $L^{\downarrow}(h_1)$, $L^{\uparrow}(h)$, H, and T_R are 377 W m^{-2}, 381 W m^{-2}, 356 W m^{-2}, and 13.2°C, respectively (the figure not shown). The values of $S^{\downarrow}(h_1)$, H_S, and lE_S are almost zero.

The foregoing results indicate that forests probably influence the climate during the snowmelt season. All the results in this paper are obtained from model cal-

culations. It is necessary to observe the snowmelt and heat balance in forested areas in consideration of the forest density.

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