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Heat Balance Model over a Vegetated Area and Its Application to a Paddy Field

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Abstract

A new canopy model having a simple parameter is here presented for estimating the diurnal or seasonal variations of the heat balance, employing meteorological data of temperature, wind, humidity, precipitation, and solar radiation gathered by AMeDAS near the experimental field. The calculated results were compared with observations over the rice paddy field for the two-year period of 1993 and 1994. Simulations using the presently developed canopy model agreed well with the daily-mean observed values. The rms error in the daily mean values between the observed and calculated values of surface temperature, sensible heat flux, and latent heat flux were 1.7°C, $\pm 13\ \mathrm{Wm^{-2}}$, and $\pm 16\ \mathrm{Wm^{-2}}$, respectively, during the overall observational period. The annual evapotranspiration was evaluated as about 600 $\mathrm{mmy^{-1}}$. This value is found between the values for shallow water (564 $\mathrm{mmy^{-1}}$) and forests (692 $\mathrm{mmy^{-1}}$). The ratio of rainfall interception to evapotranspiration during the vegetated periods are about 18 % and 9 % for the year 1993 (cool summer) and 1994 (hot summer), respectively. The precipitation amount in 1993 is large, while evapotranspiration from the vegetation is small, resulting in a large rate of rainfall interception.

1. Introduction

As vegetation types adapt to given climatic conditions, plant life has a great influence on the climate. A realistic representation of the energy exchange between a vegetated area and the atmosphere is one of the most important issues in considering a climatic variation on a landscape scale.

The bulk transfer method for estimating the energy flux is mainly developed for the ocean, snow, and bare soil surface. The bulk transfer coefficient of the sensible heat flux C_H and the latent heat flux C_E are required to evaluate the simultaneous calculated values of the surface temperature, the sensible heat flux, and the latent heat flux. The value of C_H and C_E for a vegetated surface has been physically investigated and determined from the plant height, leaf area index, and the roughness length (Kondo and Watanabe, 1992). However, a heat balance model for the vegetated area considering these analyses is not conducted to simulate the energy exchange between the vegetated area and the atmosphere over a long period. In the present study, by employing a new canopy model combined with the bulk transfer equation, the daily and seasonal variation of heat balance over a long period is simulated. The advantage of the present model is that the estimation of heat balance can be conducted by using only the daily averaged meteorological data, when the value of C_H and C_E are given using vegetation parameters. In this study, input data for the model simulation is routine meteorological data that is easy to obtain, as opposed to the detailed data from the experimental field, and the leaf bulk transfer coefficient for latent heat flux c_e that is the most important parameter for estimation of C_E is simply represented by a function of solar radiation and minimum stomatal resistance, although the value of c_e is dominated by complex plant physiology and meteorological conditions. The object of the present study is to see how the accuracy of this model with the use of a simple routine meteorological data and simple parameterization compares to observed results.

In this paper, the seasonal variation of evapotranspiration in a paddy field has been simulated over the two-year period of 1993 and 1994, employing routine meteorological data (AMeDAS; Automated Meteorological Data Acquisition System, operated by the Japan Meteorological Agency, about 1300 stations distributed in Japan). The simulated results were then compared with the micrometeo-

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rological observations. The routine meteorological variables required for the calculations are the daily amount of precipitation, the duration of sunshine, the diurnal maximum, minimum and averaged air temperatures, and the daily means of wind speed, specific humidity of the air. Parameters required for the calculations are the canopy height, leaf area index, surface albedo, individual leaf transfer coefficients, and roughness lengths of the underlying ground surface.

In Section 2, we describe the observational data of the heat balance in a paddy field. In Section 3, the presently developed canopy model and parameterizations are discussed. In Section 4 and 5, the observed values of the heat balance are compared with the model calculations for diurnal and seasonal variations, respectively.

2. Heat balance observational data

The observations by Ishida *et al.* (1997) were conducted at Kitaura (38°33′N, 141° 02′E), about 35 km north of Sendai in Miyagi Prefecture, Japan (Fig. 1). A horizontally homogeneous rice paddy field surrounds the observation site. The observational data were continuously obtained during the period from 1 December 1992 to 31 October 1994, a total of 700 days. The periods of irrigation, rice planting, and harvesting in 1993 and 1994 are listed in Table 1.

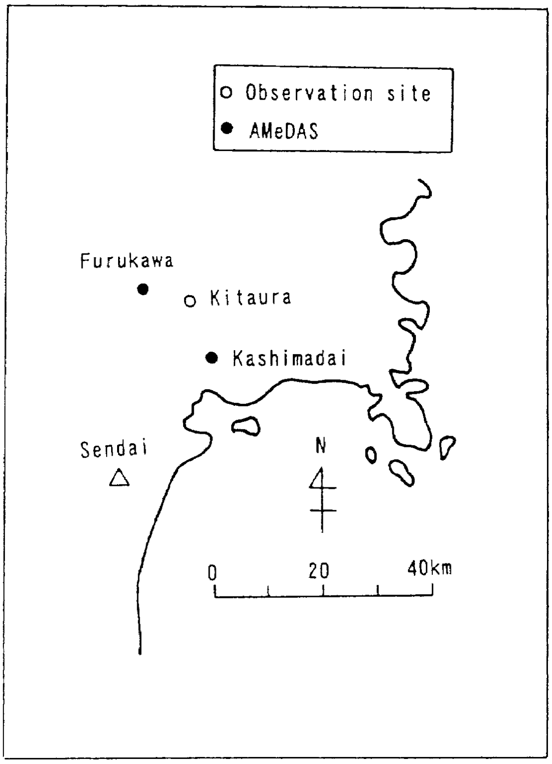

Fig. 1. A map showing the location of the observation and AMeDAS sites in Miyagi Prefecture, Japan.

The elements observed to determine the heat balance were the surface temperature T_s, air temperature T, specific humidity of air q, wind speed U, solar radiation $S\downarrow$, downward longwave radiation $L\downarrow$, upward longwave radiation $L\uparrow$, and the soil heat flux G, the sensible heat flux H and latent heat flux lE. The value of T_s was estimated by the followings equation,

$$L\uparrow = \varepsilon\sigma T_s^4 + (1-\varepsilon)L\downarrow, \tag{1}$$

where the value of $L\uparrow$ was observed by a pyrgeometer in the overall infrared region. ε is the emissivity and assumed as unity in this study, and σ the Stefan-Boltzmann constant (5.67×10^{-8} Wm^{-2}K^{-4}). Values of T and q were observed at elevations of 0.5, 1.8, 3.8, and 10 m. Values of U were observed at elevations of 1 m, 3.6 m, and 10 m. The surface fluxes H and lE were estimated by the gradient method using all elevations data of T, q and U in which atmospheric stability was considered (Ishida *et al.*, 1997).

The definition of B is given by the following equation as

$$B = (1-ref)S_M\downarrow + (L_M\downarrow - L_M\uparrow) - (H_M + lE_M + G_M). \tag{2}$$

Here *ref* is the albedo of the rice paddy, $S_M\downarrow$, $L_M\downarrow$, $L_M\uparrow$, H_M, lE_M, and G_M are the daily means of solar radiation, downward longwave radiation, upward longwave radiation, sensible heat flux, latent heat flux, and soil heat flux, respectively. The observed values of H and lE may have some errors when the absolute value of B exceeds 30 Wm^{-2}. These erroneous data of observed flux values under these conditions are not used in the present analysis.

3. Model of heat balance simulation

In this chapter, the presently developed canopy model for the heat balance simulation, employing simple routine meteorological data, and suitable parameterizations for this model are described.

3.1 Principal equations (single source model)

The basic equation for the heat balance of a land surface is given as

$$R\downarrow - G = \sigma T_s^4 + H + lE, \tag{3}$$

where

$$R\downarrow = (1-ref)S\downarrow + L\downarrow. \tag{4}$$

Here, $R\downarrow$ is the total incident radiation. The fluxes H and lE can be expressed by the bulk transfer equations as

$$H = c_p\rho C_H U(T_s - T), \tag{5}$$

and

$$lE = l\rho C_H U\beta\{q_{sat}(T_s) - q\}, \tag{6}$$

Table 1. Agricultural calendar for the years 1993 and 1994

	1993	1994	
Irrigation	5 May (125)	5 May (125)	Irrigation period
Rice planting	12 May (132)	8 May (128)	
Harvest	31 October (304)	20 September (263)	Vegetated period

where c_p and ρ are the specific heat and air density, respectively, and C_H is the bulk transfer coefficient for sensible heat. The variables U, T, and q are the wind speed, temperature, and specific humidity at reference level z, respectively, β is the surface moisture availability, and $q_{sat}(T_s)$ the saturation specific humidity at the temperature T_s. Substitution of Eqs. (5) and (6) into Eq. (3) yields an equation for T_s, given as

$$R\downarrow - G - \sigma T_s^4 - c_p\rho C_H U(T_s - T) - l\rho C_H U\beta\{q_{sat}(T_s) - q\} = 0. \quad (7)$$

In Eq. (7), it is assumed that the canopy and ground can be regarded as a single palne surface.

The calculation of G will be explained later in this paper (see Appendix C). The solution of T_s can be found through successive approximations of Eq. (7), with the fluxes H and lE evaluated from Eqs. (5) and (6) (Kondo and Watanabe, 1992). In the present study, the time interval used in the calculation is 30 minutes.

3.2 Meteorological data required for the simulation

The meteorological data required for the simulation (T_s, H, and lE) are:

(1) the sunshine duration N (hour)

(2) the daily-mean air temperature, the daily maximum and minimum air temperatures T_M, T_{Max}, T_{Min}, respectively

(3) the daily-mean wind speed U_M

(4) the daily-mean specific humidity q_M

(5) the daily amount of precipitation Pr.

Two meteorological stations (AMeDAS) located near the experimental field site are Kashimadai and Furukawa (see Fig. 1). The average values of the meteorological variables at the two stations are used for the simulation. The specific humidity, however, is not observed at AMeDAS sites. Therefore, it is evaluated from the daily-mean vapor pressure e observed at the Sendai meteorological observatory.

For the model simulation, diurnal variations of meteorological data are required. In this model, these are estimated using daily meteorological data (AMeDAS), as follows.

3.3 Solar radiation and downward flux of longwave radiation

The duration of sunshine N is observed at AMeDAS sites. According to Kondo (1994), the daily mean of solar radiation $S_M\downarrow$ can be estimated by the empirical formulae (see Appendix A for more details).

Using $S_M\downarrow$, the time series of the diurnal variation of solar radiation $S\downarrow$ is given as (Kondo and Xu, 1997).

$$S\downarrow(t) = S_M\downarrow[1 + S_1\cos(\omega t) + S_2\cos(2\omega t) + S_3\cos(3\omega t) + S_4\cos(4\omega t)], \quad (8)$$

where t is the local time, $\omega = 0.727\times10^{-4}$ s^{-1}, $S_1 = -1.503$, $S_2 = 0.584$, $S_3 = -0.058$, and $S_4 = -0.023$.

For the albedo of the rice paddy *ref*, the observed value obtained by Ishida *et al.* (1997) is used. Figure 2a shows the seasonal variation of albedo for the year 1993.

Calculated methods of the daily-mean downward longwave radiation $L_M\downarrow$ is given by Kondo and Xu (1997). It is assumed that the daily variation of downward longwave radiation is small and negligible when atmospheric conditions change only to a slight degree, since the usual amplitude of the diurnal variation is as small as 10 Wm^{-2}.

3.4 Diurnal variation of the air temperature

Using values of T_M, T_{Max}, and T_{Min}, the diurnal variation of air temperature is given as (Kondo and Xu, 1997).

$$T(t) = T_M + B_1\cos(\omega t - \alpha_1) + B_2\cos(2\omega t - \alpha_2). \quad (9)$$

where the coeffients are determined by

$$B_1 = -(T_{Max} - T_{Min})/2.09, \quad B_2 = -0.2B_1,$$

and

$$\alpha_1 = \alpha_2 = \pi/4$$

In this model, wind speed and specific humidity are assumed to be constant. The estimated result of the diurnal average fluxes is about the same as the result when the diurnal variation of these are given (Kondo, 1994). When a strict diurnal variation of the fluxes is calculated, this model should be applied making use of the detailed hourly data sets.

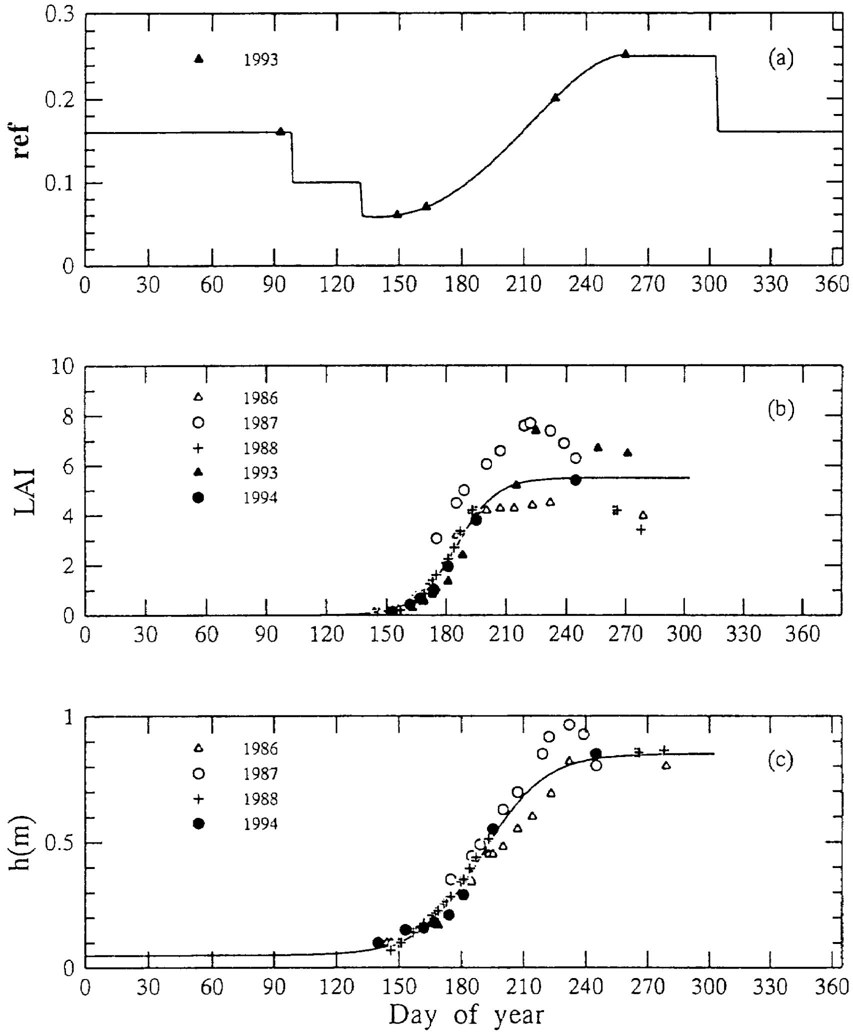

Fig. 2. The seasonal variation of (a) the albedo *ref*, (b) the leaf area index LAI, and (c) the canopy height h. The legends within the panels indicate the years the observations were taken.

3.5 Bulk transfer coefficients for sensible heat C_H, latent heat C_E, and surface moisture availability β

The values of C_H, C_E, and β have seasonal variation. In this model, the season is divided into three stages, that is, the vegetated period, the irrigation period, and periods before rice planting or after harvest. In the following discussion, the parameterization of C_H, C_E, and β is described at every stage.

3.5.1 Vegetated period of the rice paddy

Based on numerical experiments (Kondo and Watanabe, 1992), Watanabe (1994) proposed that the zero-plane displacement d, the roughness lengths for momentum z_0, sensible heat z_T, and latent heat z_q, can be parametrized by the empirical formulae (see Appendix B). The formulae employ the vegetation parameters, that is, the leaf area index LAI, plant height h, and the individual leaf transfer coefficients for momentum c_d, sensible heat c_h, and water vapor c_e. Also used are the roughness lengths of the underlying ground surface for momentum z_{0s}, sensible heat z_{Ts}, and water vapor z_{qs}.

The values of LAI and h in the calculations are obtained from observation. Figures 2b and 2c show the seasonal variation of LAI and h, respectively. It should be noted that the curves of *ref*, LAI, and h shown in Fig. 2 are for the year 1993. In 1994, the agricultural calendar (the days of rice planting, harvest, *etc.*) differed from that in 1993.

Table 2 lists the typical values of c_d, c_h, z_{0s}, and z_{Ts} that can be found in numerous vegetation types (*e.g.*, Baldocchi and Meyers, 1998; Yamazaki *et al.*, 1992; Kondo and Watanabe, 1992). Estimated by the values of d, z_0, and z_T, the seasonal variation of the bulk transfer coefficient for sensible heat C_H can be obtained from the following equation, expressed as

$$C_H = \frac{k^2}{\ln\left[(z-d)/z_0\right] \ln\left[(z-d)/z_T\right]}, \tag{10}$$

where k is the Von Karman constant ($= 0.4$).

Table 2. Parameters required for the model calculation

Element
(a) Vegetated period
Canopy hight h (m)
Leaf area index LAI
Albedo ref
Leaf drag coefficient c_d: 0.2
Leaf bulk heat transfer coefficient c_h: 0.06
Leaf bulk vapor transfer coefficient c_e: Eq. (13)
Roughness length of the ground surface
For momentum z_{0s} (m): 0.001
For sensible heat z_{Ts} (m): 0.001
For specific humidity z_{qs} (m): 0.001
(b) Irrigation period
Roughness length
For momentum z_0 (m): 0.001
For sensible heat z_T (m): 0.001
Surface moisture availability β: 1.0
(c) Except for (a), (b)
Roughness length
For momentum z_0 (m): 0.01
For sensible heat z_T (m): 0.001
Surface moisture availability β: 0.6

Equation (10) is valid for the case of neutral atmospheric stability. In fact, atmospheric stability should be considered when fluxes are calculated in the hourly times interval. The stability of the lower atmosphere, however, is not considered in the present study, since the simultaneous calculated values of the surface temperature and heat balance components by the principal equation (Eq. (7)) are not sensitive to $C_H U$ (Kondo and Watanabe, 1992). For instance, when the value of $C_H U$ is doubled, the daily averaged latent heat flux lE_M have an increase by 45 $\mathrm{Wm^{-2}}$.

The underlying ground surface value of z_{qs} required for calculating z_q is taken from Table 2. For a rice paddy field, it can be assumed that $z_{Ts} \fallingdotseq z_{qs}$, since the soil surface is sufficiently wet during the vegetated period.

In calculating z_q, the value of c_e (individual leaf transfer coefficient for water vapor) is important. The relationship between the resistances and the transfer coefficient are written as

$$c_e U(Z_{canopy}) = \frac{1}{r_a + r_s}, \quad (11)$$

$$c_h U(z_{canopy}) = \frac{1}{r_a}, \quad (12)$$

where r_a is the aerodynamic resistance of the leaf surface, r_s the stomatal resistance and z_{canopy} the representative heights for the wind speed within the canopy. Hsiao *et al.* (1973) verified that the stomatal aperture of C3 plants such as rice reaches a maximum when the intensity of solar radiation is about 10 $\mathrm{Wm^{-2}}$. The aperture then remains constant under conditions of increased solar radiation. In fact, c_e is dominated by various conditions (plant physiology and meteorological conditions). Dickinson (1984), Sellers *et al.* (1986) and Sellers *et al.* (1995) parametrized stomatal resistance considering these conditions in detail. In this research, however, the diurnal variation of c_e is simply given as a function of solar radiation and c_e in midday, and can be described by

$$c_e = 0.048 \left[1 - \frac{1}{1 + \tan h(0.2 S \downarrow)} \right]. \quad (13)$$

Using this equation, the maximum value of c_e in midday is 0.024, while the minimum stomatal resistance r_{sm} is about 50 $\mathrm{sm^{-1}}$. The maximum value of c_e in midday was estimated by substituting the observed data of β into empirical formulae (see Appendix B). The solution for c_e in midday can be found through successive approximations, with the use of Appendix B, so that the calculated value of β agree with the observed values. The values assigned for these estimates are listed in Table 2.

Using the value of c_e, z_q is calculated similar to the calculation for z_T. The bulk transfer coefficient for latent heat C_E is calculated using the value of z_q by the following equation, given as

$$C_E = \frac{k^2}{\ln[(z-d)/z_0] \ln[(z-d)/z_q]}. \quad (14)$$

The moisture availability β is calculated using Eqs. (10) and (14), expressed in the following equation as

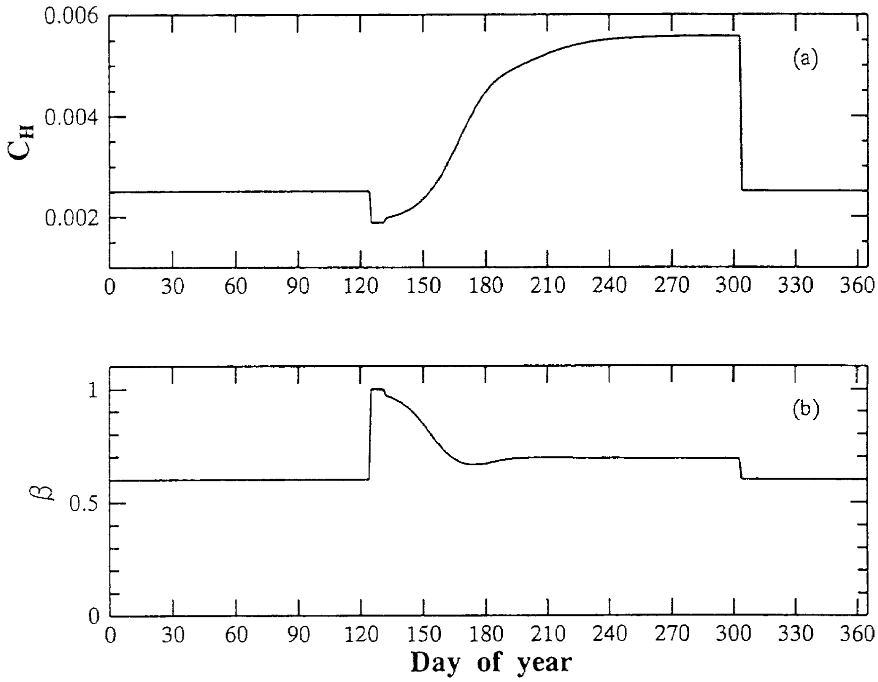

Fig. 3. The seasonal variation of (a) the bulk transfer coefficient for sensible heat C_H, and (b) the surface moisture availability β in the rice paddy field for the year of 1993.

$$\beta = \frac{C_E}{C_H}. \qquad (15)$$

β is the moisture availability for the overall canopy, not for individual leaf.

3.5.2 Periods before rice planting, during irrigation, and after harvest

The observed values of roughness lengths (z_0 and z_T), estimated from the wind speed and air temperature profiles by Ishida *et al.* (1997), are adopted for the periods before the rice planting, during irrigation, and after the harvest, as listed in Table 2.

For the period prior to the rice planting and after the harvest, rice straw is found scattered over the entire paddy. The rice paddy field can not be regarded as a completely bare soil surface, and consequently the value of β is difficult to determine. According to observations by Ishida *et al.* (1997), there is a great deal of complexity about β for this period. For this reason, the calculation of the heat balance for this period is conducted by assuming β as a constant ($= 0.6$) that is averaged value of β for this period.

During periods of irrigation, the value of β is regarded as unity, since the soil surface is covered by water. Strictly speaking, the value of β has not reached unity because the roughness lengths for sensible heat and water vapor are not the same, but the difference is negligible (Kondo and Watanabe, 1992).

Figures 3a and 3b show the seasonal variation of C_H and β (for the year of 1993), respectively, calculated from Eqs. (10) and (15).

3.5.3 Calculations under conditions of nocturnal condensation

When condensation occurs ($q_{sat}(T_s) < q$) during nighttime, the results are recalculated with the value of β being set to unity.

3.6 Calculation for soil heat flux G

The soil heat flux G is an unknown quantity when estimating T_s, H, and lE using Eqs. (5)–(7). In the present paper, the value of G is defined as

$$G = G_D + G_Y, \qquad (16)$$

where G_D and G_Y are the diurnal and seasonal variations of the soil heat flux, respectively.

The value of G can be analytically determined using meteorological data, bulk transfer coefficient C_H, evapotranspiration efficiency β, albedo *ref*, and soil parameters (see Appendix C).

3.7 Rainfall interception

The interception of rainfall by the leaves is defined as following;

(1) evaporation from wetted leaves during rainfall duration

(2) evaporation of water storage on leaves S, after the precipitation.

In these conditions, the calculation of the heat balance is conducted by assuming β as unity. In the case of (2), assumption of $\beta = 1$ is applied to only

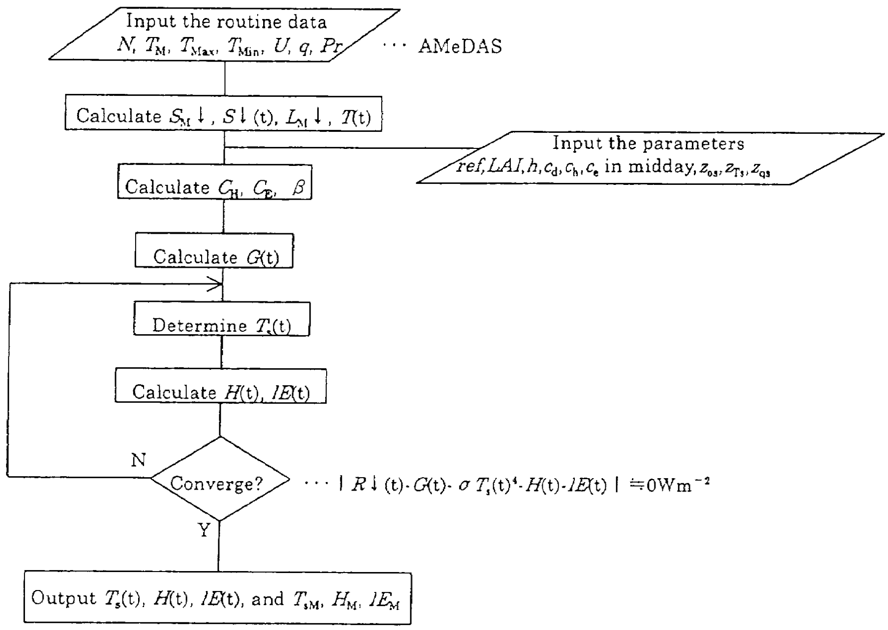

Fig. 4. Algorithm for the presently developed canopy model.

when $S > 0$, that is,the calculation is conducted as $\beta = 1$ until the value of S equals zero.

Water storage of the leaves S is evaluated as a function of precipitation (Kondo *et al.*, 1992), as

$$S = S_{Max}\left[1 - \exp(-Pr/S_{Max})\right], \qquad (17)$$

where

$$S_{Max} = 0.15 \cdot LAI. \qquad (18)$$

Here, S_{MAX} is the maximum water storage of the leaves of vegetation, and Pr the precipitation rate (mm d^{-1}). The coefficient of 0.15 in Eq. (18) can be adopted for many types of vegetation. Rainfall duration τ(hr) can be statistically expressed by the following equation (Kondo and Xu, 1997), as

$$\tau = 1.67 Pr^{0.5} \qquad (19)$$

In Eq. (19), $\tau = 12$(hr) when $Pr = 50$ mmd^{-1}. In this study, it is assumed that it begin to rain at midday when $Pr \leq 50$ mmd^{-1}, and at midnight when $Pr > 50$ mmd^{-1}.

The flow chart for this calculation is shown in Fig. 4. In this study, Eq. (7) can be rewritten as

$$\begin{aligned}(1 - ref)S\downarrow(t) &+ L_M\downarrow - G(t) - \sigma T_s(t)^4 \\ &- c_p\rho C_H U_M (T_s(t) - T(t)) \\ &- l\rho C_H U_M \beta\left\{q_{sat}(T_s(t)) - q_M\right\} = 0. \qquad (20)\end{aligned}$$

4. Calculated results for diurnal variations

In this section, the diurnal variations for 13 days, from 2 July to 14 July 1994, are examined. Figure 5 shows a comparison between the observed (open circles) and calculated (solid lines) values for various quantities. It should be noted that a rigid calculation of diurnal variation can not be conducted by using diurnal averaged data, since the input data (solar radiation and air temperature) are given as trigonometric functions having four and two temporal harmonics. When calculating a strict diurnal variation of the heat fluxes, this model should be applied, making use of the detailed hourly data sets.

First, an examination is made of the diurnal variations of solar radiation $S\downarrow$ (Fig. 5 top panel). The time series used in the calculation is represented by a trigonometric function having four temporal harmonics, as in Eq. (8). The harmonics vary smoothly compared with the observed midday values. On cloudy days, observed values fluctuate only slightly, with the diurnal average of the calculated values coinciding with observed values. On fair weather days, however, the diurnal variation is large and the calculated values almost coincide with those observed.

Next, the surface temperature T_s (Fig. 5, middle panel) is examined. The calculated values of T_s are somewhat overestimated when compared with observed values, being within an error of 2.2°C. It is considered that this overestimation is caused by the hemispherical radiometric temperature of the surface differing from the aerodynamic surface temperature based on the definition of C_H in this paper. The hemispherical radiometric temperature was evaluated by a measurement using a dome-type pyrgeometer of the entire infrared region with an adjustment to account for the surface emissivity and downward longwave radiation (Eq. (1)).

Concerning the soil heat flux G (Fig. 5 second panel), the sensible heat flux H (Fig. 5, second panel

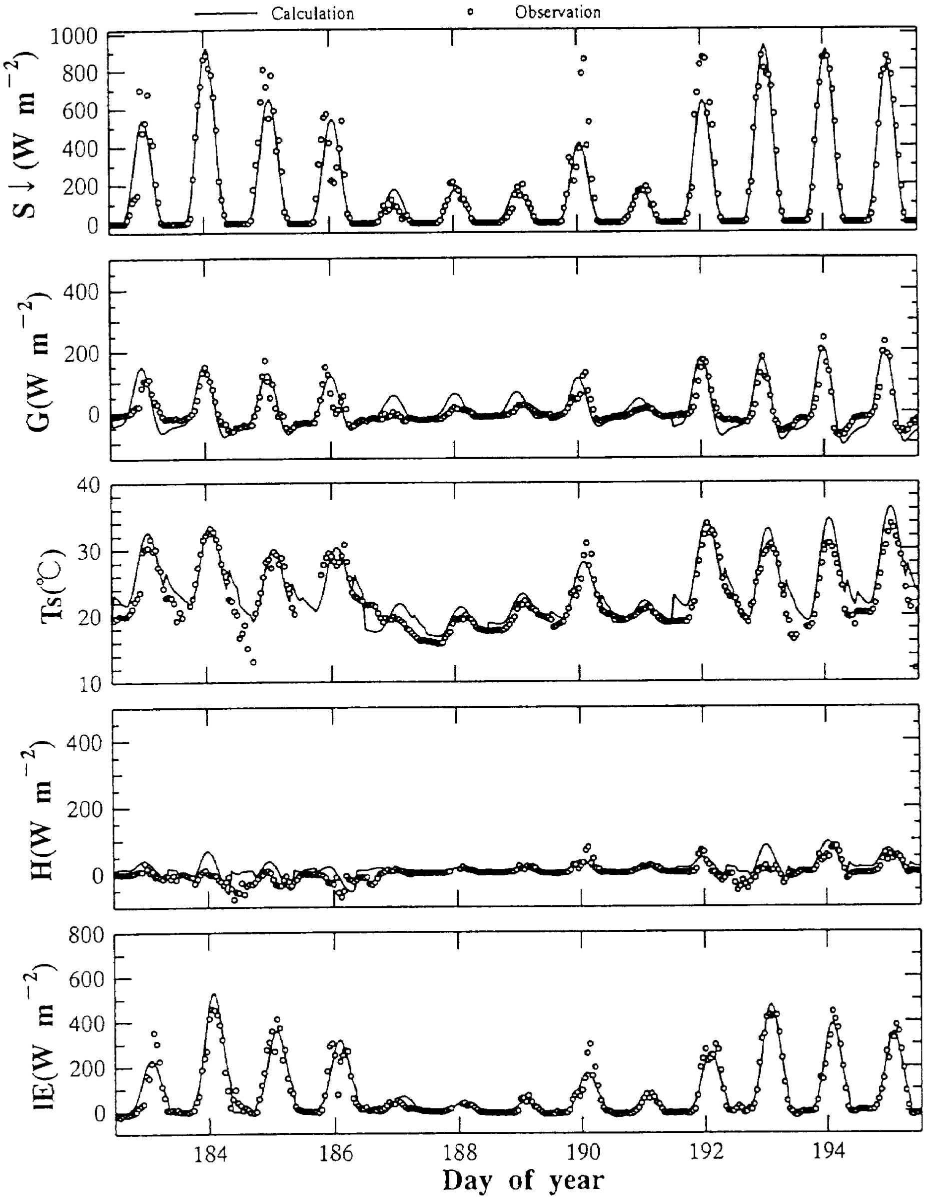

Fig. 5. Diurnal variations of the solar radiation $S\downarrow$, the soil heat flux G, the surface temperature T_S, the sensible heat flux H, and the latent heat flux lE during day 183 to day 195 (2 July to 14 July 1994). Open circles denote observed values, and solid lines the calculated values.

from bottom), and latent heat flux lE (Fig. 5, bottom panel), the trend of the diurnal variation of the calculated values almost agrees with those observed.

5. Calculated results for seasonal variations

5.1 Comparison with observations

Figures 6a and 6b show the seasonal distributions of the observed (open circles) daily means of solar radiation $S_M\downarrow$, downward longwave radiation $L_M\downarrow$, air temperature T_M, and the daily precipitation rate Pr for 1993 and 1994, respectively. For the $S_M\downarrow$ (top panels) and $L_M\downarrow$ (second panels from top), the calculated values (solid lines) are also shown. The rms errors between the observed $S_M\downarrow$ and calculated $S_M\downarrow$ are within ± 26 Wm^{-2} and ± 22 Wm^{-2} for 1993 and 1994, respectively. On the other hand, those for $L_M\downarrow$ are ± 15 Wm^{-2} and ± 16 Wm^{-2}, respectively.

The year 1993 exhibited more days of successive low temperatures compared with that of 1994. In particular, there were summer (July to August) days when the daily averaged temperature T_M was under 20°C. These cool days resulted in unusual agricultural damage. In contrast, the summer period of 1994 was hot, with no unusual damage occurring.

Figures 7a and 7b show a comparison of the observed (open circles) and the calculated (solid lines) values of the daily averaged surface temperature T_{sM} (top panels), the daily averaged sensible heat flux H_M (middle panels), and the daily averaged la-

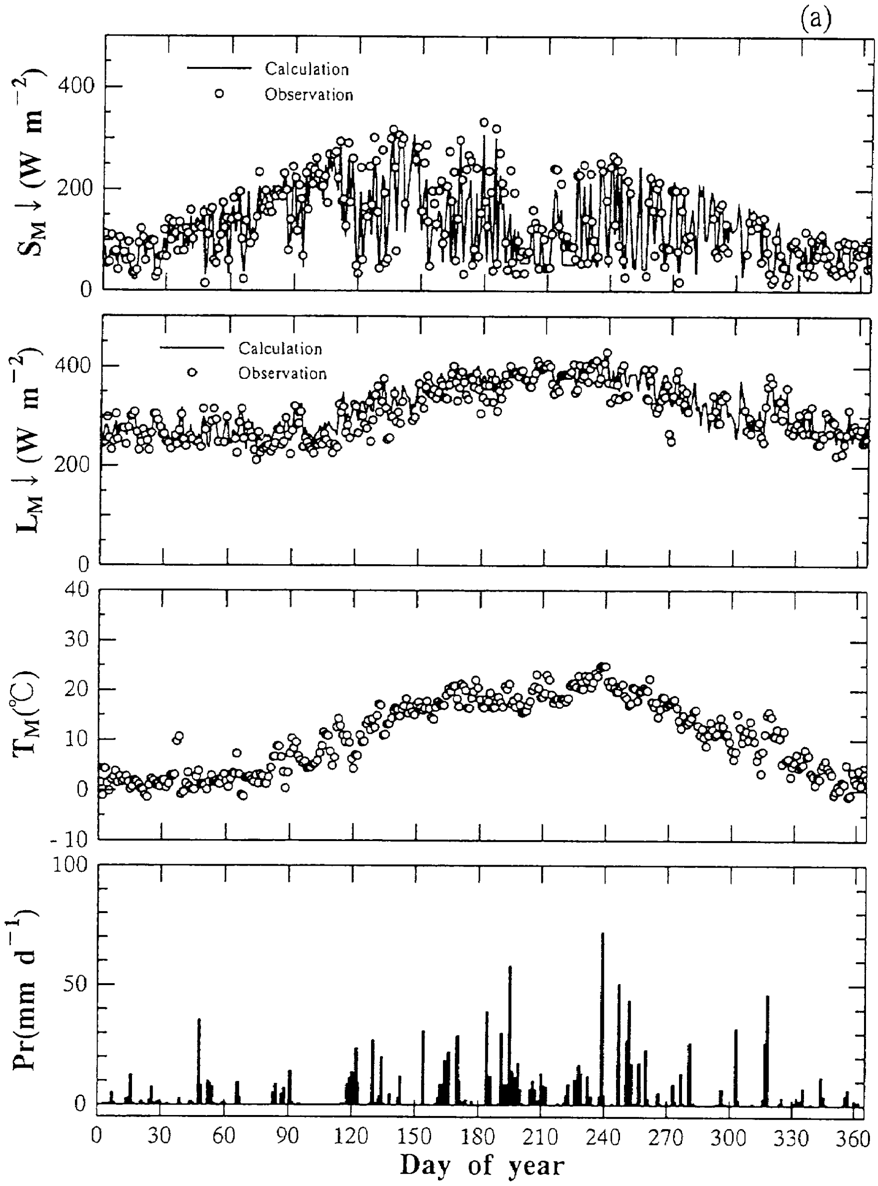

Fig. 6. The seasonal variation of the daily mean values $S_M \downarrow$, $L_M \downarrow$, T_M, and daily amount of precipitation Pr for the year (a) 1993 and (b) 1994. The open circles indicate observed values of $S_M \downarrow$, $L_M \downarrow$, and T_M, while the solid straight lines denote Pr (bottom panel). Solid lines represent calculated values of $S_M \downarrow$ and $L_M \downarrow$.

tent heat flux lE_M (bottom panels) for 1993 and 1994, respectively. For the winter period (day 1 to 124 and 304 to 365 in 1993; 1 to 124 and 263 to 365 in 1994), the calculations were performed with β held constant (= 0.6).

Figures 8a and 8b show a comparison of the observed and calculated values for the period from rice panting to harvest, *i.e.*, day 132 to 303 in 1993 (Fig. 8a) and 128 to 262 in 1994 (Fig. 8b).

The values of the calculated surface temperature T_{sM} (Figs. 7 and 8, top panels) are overestimated in both yearly periods. The rms error between the observed and calculated T_{sM} during the vegetated period are within 1.2°C and 1.3°C , while those over the entire year are within 1.4°C and 1.9°C, for the periods of 1993 and 1994, respectively. As previously mentioned, it is considered that the observed values of T_{sM} are different from the calculated values that are based on the bulk transfer coefficient C_H defined in this paper. That is, the leaf temperature has a vertical profile within the canopy. Matsushima and Kondo (1997) indicated that a range of optimum viewing angles exist for the measurement of canopy surface temperature by a thermal infrared radiometer, based on a simulation with a multilayer canopy model. The optimum viewing angle was found to be between 50° and 70° of the nadir angle. In the future, estimated results should be

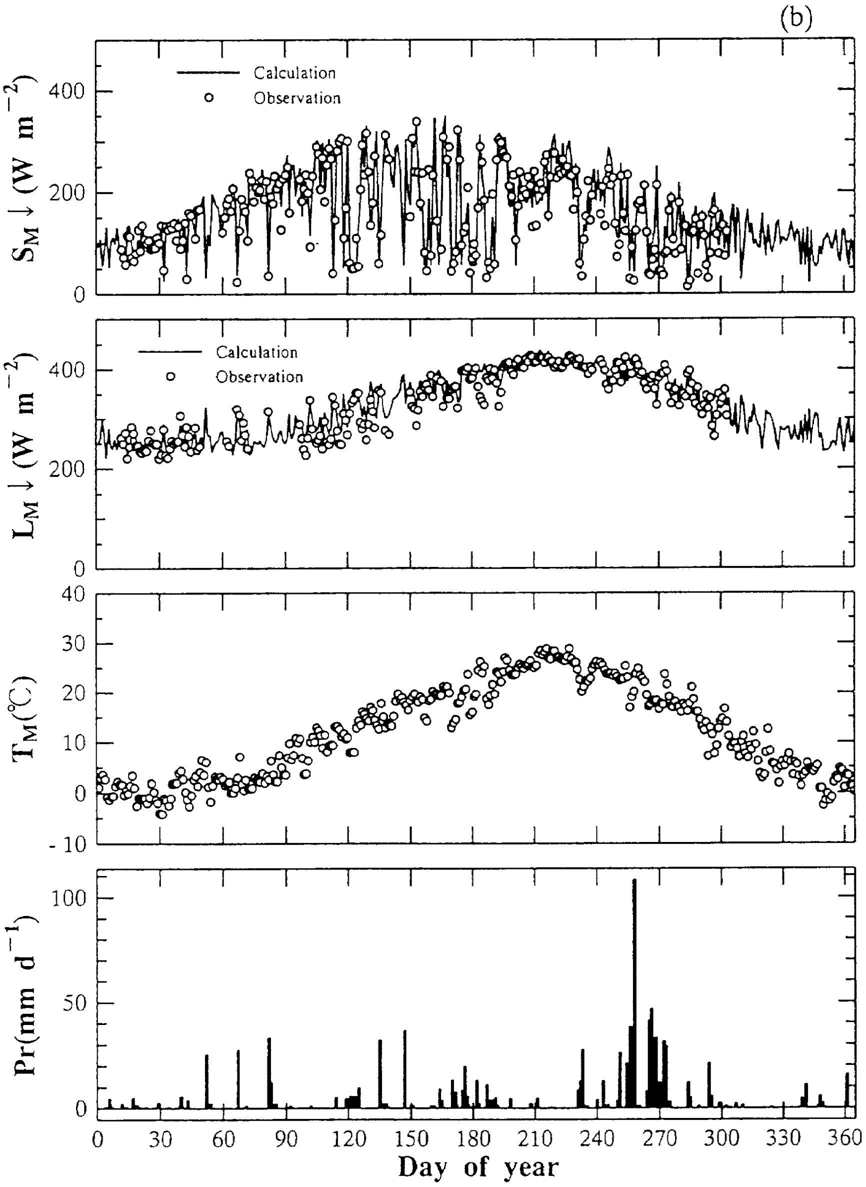

Fig. 6 (Continued)

compared with observation using a thermal infrared radiometer.

The daily mean observed sensible heat flux H_M (Fig. 7 middle panels) is about 20–30 Wm^{-2} over the two year period. The rms error between the observed and the calculated values of H_M during the vegetated period are ± 9 Wm^{-2} and ± 11 Wm^{-2}, while those over the entire year are ± 10 Wm^{-2} and ± 15 Wm^{-2}, for 1993 and 1994, respectively.

The latent heat flux lE_M values are shown in the bottom panels of Figs. 7 and 8 for the entire year and vegetated period, respectively. The rms errors between the observed and calculated values are ± 14 Wm^{-2} and ± 17 Wm^{-2} for the vegetated period of the year 1993 and 1994, respectively. Furthermore, those during the overall yearly experimental periods are ± 14 Wm^{-2} and ± 17 Wm^{-2}, respectively. At the present time, values of the diurnal averaged latent heat flux by various observational methods (gradient method, Bowen ratio method, and so on) may have an error of about ± 20 Wm^{-2}. This result shows that the calculated error is within the observational error.

5.2 The rate of rainfall interception

In this section, the rate of rainfall interception to rainfall is examined. With respect to evapotranspiration, rainfall interception is also included.

Table 3 indicates that the ratio of rainfall interception to evapotranspiration during the vegetated periods are about 18 % and 9 % for the years 1993 (cool summer) and 1994 (hot summer), respectively. The precipitation amount in 1993 is large, while evapotranspiration from the vegetation is small, re-

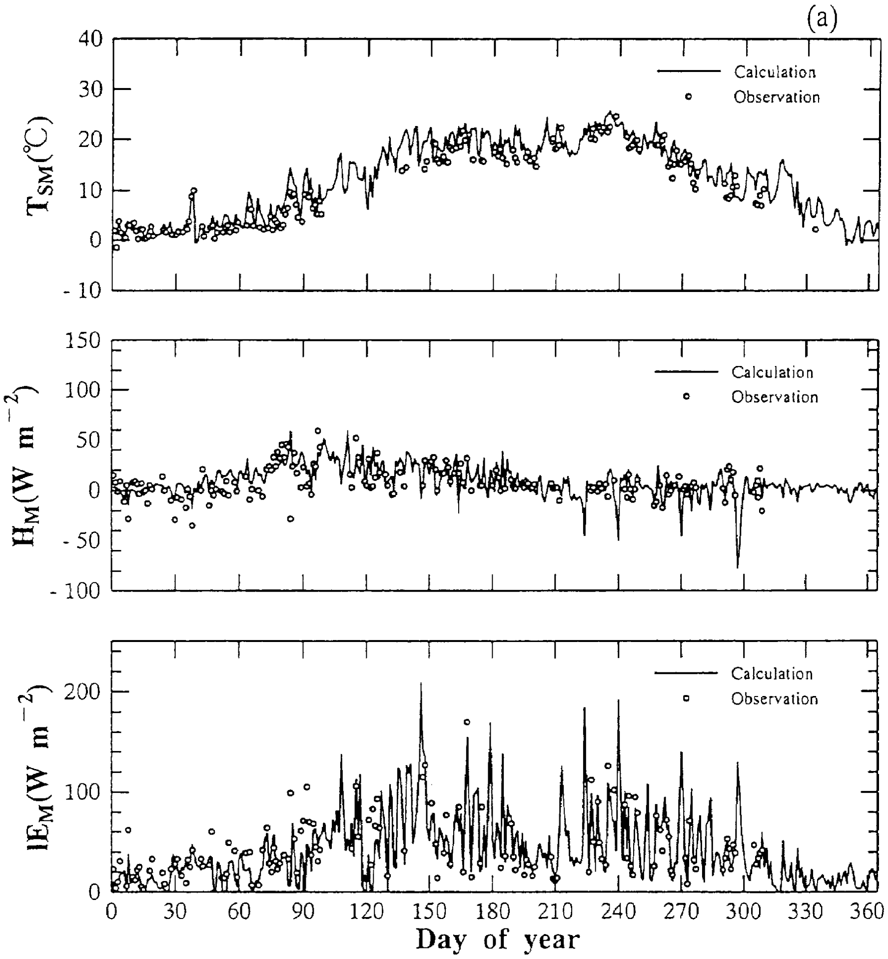

Fig. 7. The seasonal variation of the daily mean values T_{SM}, H_M, and lE_M for the year (a) 1993 and (b) 1994. Open circles denote observed values, and solid lines the calculated values.

sulting in a large rate of rainfall interception.

The amount of interception loss has been reported as about 246 mmy^{-1} for an annual rainfall amount of 2000 mmy^{-1} in a typical Japanese (LAI = 6) located in Sendai (Kondo *et al.*, 1992). The results in Table 3 indicate that the rate of interception to annual rainfall was about 5 %, being less than that of the forest by 7 %. This results from the fact that the vegetation does not exist after the autumn harvest in the rice paddy field. In order to explain this in greater detail, the rate of interception to the rainfall during the vegetated period is also considered. It was found that the rate for the paddy field during the vegetated period is less than that of the forest by 4 %. The constant in Eq. (22) was determined as 0.15 being less than that in the forest. This results from the fact that not only the leaves, but also the trunk and branches of trees can easily intercept the raindrops in the forest. It should be noted that this constant was determined to be about 0.3 to 0.4 in typical Japanese forests.

5.3 Annual evapotranspiration

The annual evapotranspiration in the rice paddy field is about 600 mmy^{-1} as listed in Table 3. Kondo and Kuwagata (1992) and Kondo *et al.* (1992) conducted a comparison of the annual evapotranspiration from the Japanese forests with that from shallow bodies of water. It was reported that these values become large, on the order of 564 mmy^{-1} for shallow water and 692 mmy^{-1} for forests. The value of evapotranspiration from the rice paddy field is found between the values for shallow water and forest. The rice paddy value is less than those in the forest by about 100 mm. As previously mentioned, this occurs from the fact that the evaporation rates are relatively large due to the rainfall interception in the forests.

The total annual rainfall was large in 1993, while the annual evapotranspiration was less by 143 mm compared with those in 1994. This result reflects the fact that 1993 was a year with a cool moist summer in which agricultural damage occurred.

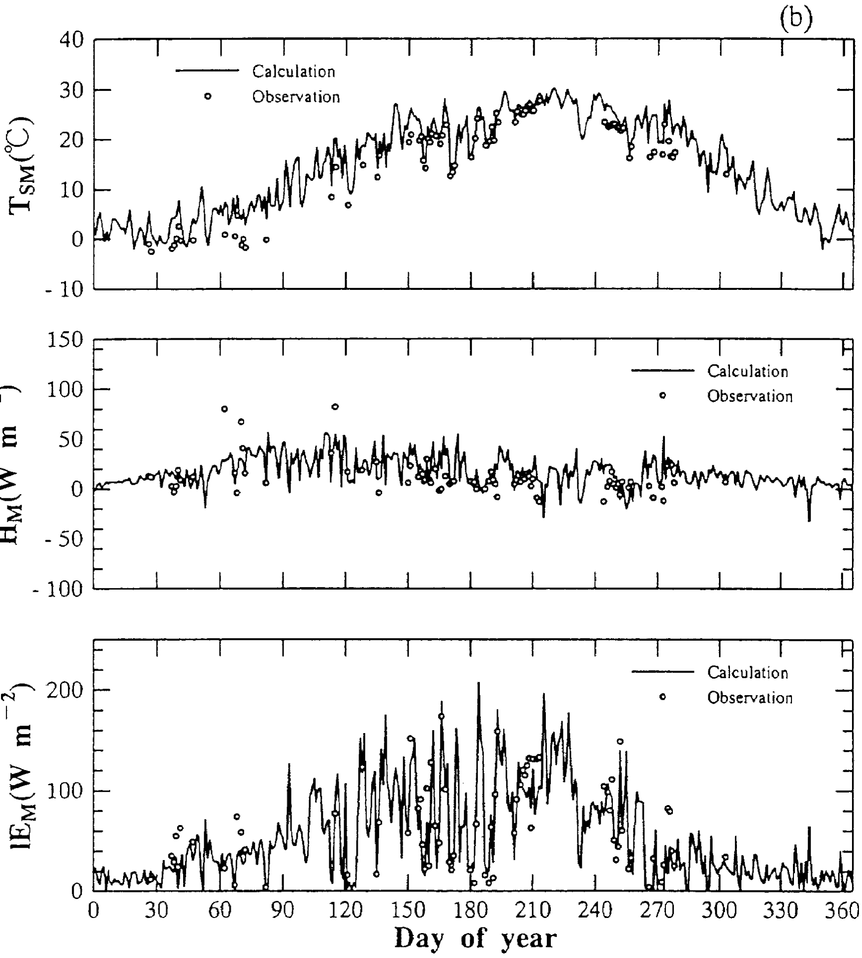

Fig. 7. (Continued)

6. Conclusions

In the present study, a new canopy model is presented for estimating the diurnal or seasonal variations of heat balance. Our main objective with this study is to evaluate the heat balance by model calculation with the use of simple routine meteorological data, and to evaluate the applicability of this model. The calculated results were compared with observations over a paddy field.

The accuracy of heat balance on an hourly basis is not good when the cloud amount varies, since the diurnal variation of solar radiation and air temperature is approximated by a trigonometric function having temporal harmonics. When calculating a strict diurnal variation of the heat fluxes, this model should be applied, making use of the detailed hourly data sets.

On a daily basis, however, the daily averaged heat balance can be well estimated by using simple routine meteorological data. The rms error in the daily mean values between the observed and calculated values of surface temperature T_{sM}, sensible heat flux H_M, and latent heat flux lE_M during vegetated periods were 1.3°C, ± 10 Wm^{-2} and ± 16 Wm^{-2}, respectively, while those over the entire two year period were 1.7°C, ± 13 Wm^{-2}, and ± 16 Wm^{-2}, respectively. These results indicate that seasonal variations of evapotranspiration can be simulated within the observational error, although simple routine meteorological data is used.

In this model, required parameters for calculation are plant height h, leaf area index LAI, surface albedo *ref*, individual leaf transfer coefficients c_d, c_h, c_e, and roughness lengths of the underlying ground surface z_{0s}, z_{Ts}, and z_{qs}. With regard to the values of c_d, c_h, z_{0s}, z_{Ts}, and z_{qs}, the typical values that can be found in numerous vegetation types are used, while, the maximum value of c_e in midday is determined by observation as 0.024 (minimum stomatal resistance $\fallingdotseq$ 50 sm^{-1}). The diurnal variation of c_e is simply represented by a function of solar radiation. In fact, the stomatal resistance is dominated by various conditions (air humidity, CO_2 concentration *etc.*). The results, however, indicate that the evapotranspiration from a wet area like a paddy field can be estimated even if a simple parameterization is conducted.

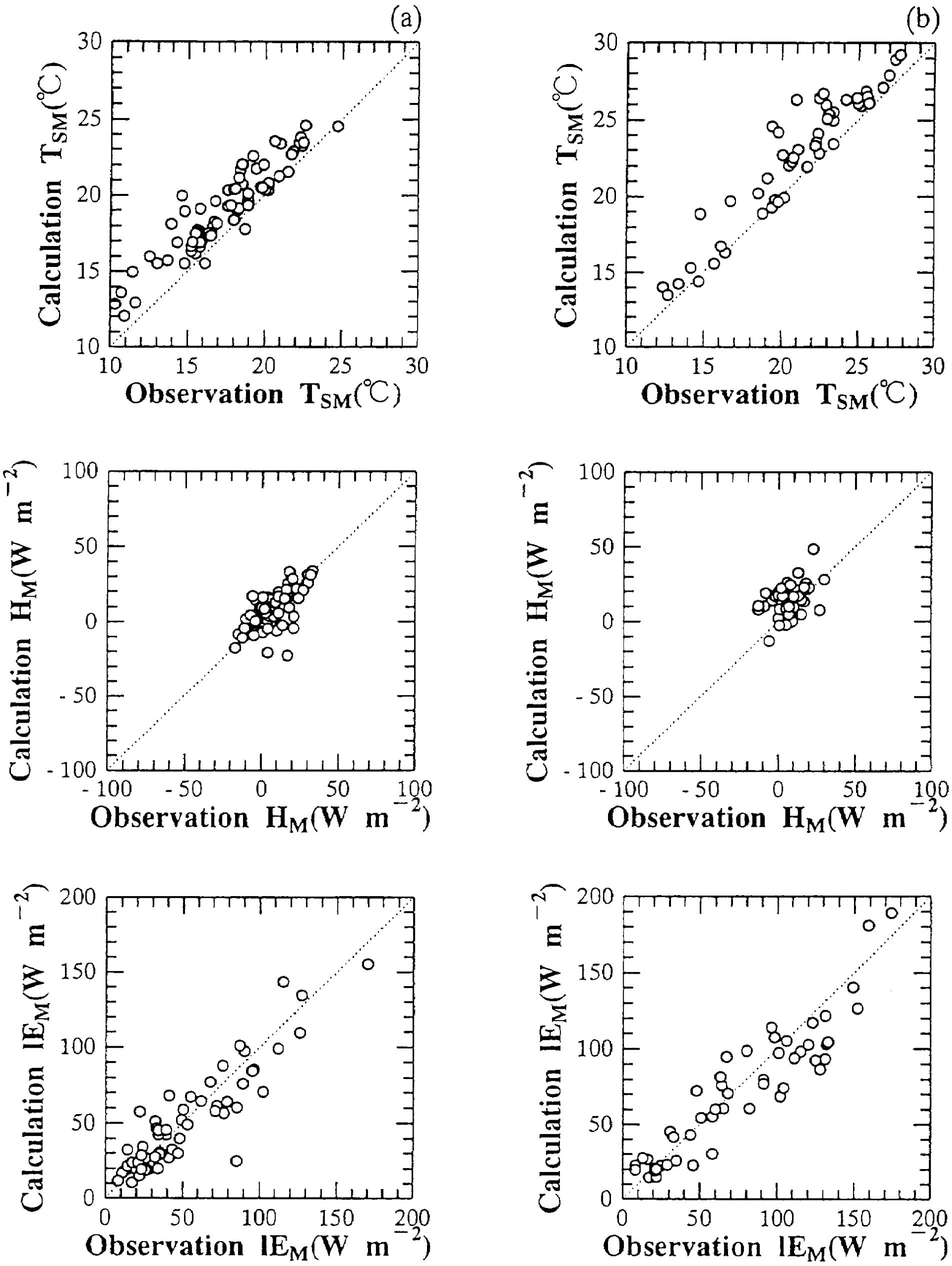

Fig. 8. Comparison of observed and calculated values of T_{SM}, H_M, and lE_M during the vegetated periods for the year (a) 1993 and (b) 1994.

The percentage of the rate of interception to the annual rainfall in a paddy field was about 5 % and less than that in a forest by 7 %. The annual evapotranspiration in the rice paddy field was about 600 mmy^{-1}, and had a value found between those for shallow water (564 mmy^{-1}) and forests (692 mmy^{-1}).

How general is this study for other vegetation? The same calculations have been conducted on a sorghum field (C4 plant) during the summer of 1994 in the Tottori sand dunes (Kimura *et al.*, 1997). The result shows that the value of c_e in midday is 0.026 (minimum stomatal resistance r_{sm} = 63 sm^{-1}), and the rms error in the daily mean values between the observed and calculated values of T_{sM}, H_M, and lE_M were 1.2°C, ±25 Wm^{-2}, and ±17 Wm^{-2}, respectively. These results are almost as same as those in the present study.

Can we use these results to calculate global land surface fluxes from standard routine meteorological observations? In the future, we would like to evaluate the heat balance of a large area using simple routine meteorological data, by accumulating model simulations.

Acknowledgments

The authors wish to thank Mr. S. Ishida and Ms. T. Ishida for their support during this study.

Table 3. Values of precipitation, evapotranspiration, and rainfall interception for the years 1993 and 1994

	1993	1994
Overall period (day)	365	365
Rainfall days (day)	145	114
Precipitation (mm)	1253	991
Evapotranspiration (mm)	559	702
Vegetated period	172	135
Rainfall days (day)	50	27
Precipitation (mm)	860	495
Evapotranspiration (mm)	361	438
Rainfall interception (mm)	66	41
Except for vegetated period	193	230
Rainfall days (day)	95	87
Precipitation (mm)	393	496
Evaporation (mm)	198	264

Appendix A

Solar radiation

According to Kondo and Xu (1998), the daily mean of solar radiation $S_M\downarrow$ can be estimated by the use of the following equations

$$\frac{S_M\downarrow}{S_{0M}\downarrow} = a + b\frac{N}{N_0}, \quad \text{for} \quad 0 < \frac{N}{N_0} \le 1, \tag{A1}$$

$$= c, \quad \text{for} \quad \frac{N}{N_0} = 0, \tag{A2}$$

where

$$N_0 = \frac{2\zeta}{0.2618},$$

$$\sin(\zeta/2) = \left(\frac{A_S}{\cos\phi\cos\delta}\right)^{1/2}, \tag{A3}$$

$$A_S = \sin\left(\frac{\pi}{4} + \frac{\phi - \delta + r}{2}\right)\sin\left(\frac{\pi}{4} - \frac{\phi - \delta - r}{2}\right), \tag{A4}$$

$$S_{0M}\downarrow = \frac{I_{00}}{\pi}\left(\frac{d_0}{d}\right)^2 (\zeta\sin\phi\sin\delta + \cos\phi\cos\delta\sin\zeta), \tag{A5}$$

$$\zeta = \cos^{-1}(-\tan\phi\tan\delta), \tag{A6}$$

$$\left(\frac{d_0}{d}\right)^2 = 1.00011 + 0.034221\cos\eta + 0.00128\sin\eta + 0.000719\cos 2\eta + 0.000077\sin 2\eta, \tag{A7}$$

$$\delta = \sin^{-1}(0.398\sin a_2), \tag{A8}$$

$$a_2 = 4.871 + \eta + 0.033\sin\eta, \tag{A9}$$

and

$$\eta = (2\pi/365)day. \tag{A11}$$

In the above, $S_{0M}\downarrow$ is the daily-mean solar radiation at the top of the atmosphere, N the duration of sunshine (hr), N_0 the duration of possible sunshine (hr), ζ the hourly angle from sunrise to meridian transit time in which a correction $r(= 0.01\text{rad})$ is considered, ϕ the latitude, and δ the solar declination. The coefficients a, b, and c have different values depending on the type of sunshine recorder; $a = 0.244$, $b = 0.511$, and $c = 0.118$ at the AMeDAS sites. In addition, I_{00} is the solar constant, and day the number of days from 1 January of the given year.

Appendix B

Empirical formulae for the aerodynamic parameters

1) The zero-plane displacement d

$$\frac{d}{h} = 1 - \frac{1 - \exp(-A^+)}{A^+}, \tag{B1}$$

with

$$A^+ \equiv \frac{C_*}{2k^2}, \tag{B2}$$

and

$$C_* = c_d \cdot LAI. \tag{B3}$$

2) The roughness length for momentum z_0

$$\left(\ln\frac{h-d}{z_0}\right)^{-1} = \left[1 - \exp(-A^+) + \left(-\ln\frac{z_{0s}}{h}\right)^{-1/0.45} \cdot \exp(-2A^+)\right]^{0.45}. \tag{B4}$$

3) The roughness lengths for sensible heat z_T and water vapor z_q

By assigning $z_x = z_{T,q}$ (canopy values), $z_{xs} = z_{Ts,qs}$ (underlying ground surface values), and $F_x = c_{h,e}/c_d$, the scalar roughness length for the case of $F_x = 0$ (represented by $z_x{}^+$), can be calculated by the following equations as

$$\left(\ln\frac{h-d}{z_x{}^+}\right)^{-1} = \frac{1}{-\ln(z_x{}^+/h)}\left[\frac{P_1}{P_1 + A^+\exp(A^+)}\right]^{P_2}, \tag{B5}$$

where

$$P_1 = 0.0115\left(\frac{z_{xs}}{h}\right)^{0.1}\exp\left[5\left(\frac{z_{xs}}{h}\right)^{0.22}\right], \tag{B6}$$

and

$$P_2 = 0.55\exp\left[-0.58\left(\frac{z_{xs}}{h}\right)^{0.35}\right]. \tag{B7}$$

Using the value of z_x^+, the scalar roughness length z_x for $0 < F_x \leq 1$ can be calculated by the following equations, given as

$$\left(\ln\frac{h-d}{z_0}\right)^{-1}\left(\ln\frac{h-d}{z_x}\right)^{-1} = C_x$$
$$\left[1-\exp(-P_3A^+)+\left(\frac{C_x{}^0}{C_x^\infty}\right)^{1/0.9}\cdot\exp(-P_4A^+)\right]^{0.9}, \tag{B8}$$

where

$$\frac{1}{C_x^0} \equiv \ln\frac{h-d}{z_0}\ln\frac{h-d}{z_x^+}, \tag{B9}$$

$$C_x^\infty \equiv \frac{(1+8F_x)^{1/2}-1}{2}, \tag{B10}$$

$$P_3 \equiv [F_x + 0.084\exp(-15F_x)]^{0.15}, \tag{B11}$$

and

$$P_4 \equiv 2F_x^{1.1}. \tag{B12}$$

Appendix C

Soil heat flux

1) Diurnal variation of soil heat flux G_D

Kondo (1994) parametrized the value of G_D by using the time t as a trigonometric function having four temporal harmonics for the diurnal variation of the air temperature and radiation, as in the following equation. The daily time series of surface temperature, using the daily averaged surface temperature T_{sM}, is given as

$$T_S(t) = T_{SM} + \sum_{n=1}^{4} A_n\cos(n\omega t - \phi_n). \tag{C1}$$

The time series of G_D can be predicted by the following equation, expressed as

$$G_D(t) = \sum_{n=1}^{4} G_n\cos\left(n\omega t - \phi_n + \frac{\pi}{4}\right), \tag{C2}$$

where

$$G_n = A_n\left(n\omega c_G\rho_G\lambda_G\right)^{0.5}. \tag{C3}$$

The volumetric thermal capacity $c_G\rho_G$ and the thermal conductivity λ_G of the soil in the paddy field (Ishida *et al.*, 1997) are defined as

$$c_G\rho_G = (1-\theta_{sat})c_s\rho_s + c_w\rho_w\theta, \tag{C4}$$

where

$$c_s\rho_s = 1.447\times10^6\ \mathrm{Jm^{-3}K^{-1}},$$
$$c_w\rho_w = 4.2\times10^6\ \mathrm{Jm^{-3}K^{-1}},$$

and

$$\lambda_G = 0.25 + 0.5\theta^{1/3}\ \mathrm{Wm^{-1}K^{-1}}.$$

Here, $c_s\rho_s$ and $c_w\rho_w$ are the thermal capacity of the soil and water, respectively, θ_{sat} is the saturated volumetric water content (0.77 $\mathrm{m^3m^{-3}}$), and θ the volumetric water content.

According to Kondo (1994), the amplitude A_n and phase ϕ_n of the surface temperature for wavenumber n are calculated by

$$A_n = \frac{R_n\cos\phi_n + B_n\xi\cos(\alpha_n-\phi_n)}{\mu+\Gamma_n}, \tag{C5}$$

where

$$R_n = (1-ref)S_n + L_n, \tag{C6}$$

and

$$\phi_n = \arctan(X_n). \tag{C7}$$

In Eqs. (C5) and (C7), the variables are defined as

$$X_n = \frac{\Gamma_nR_n + \xi B_n\left[\Gamma_n\cos\alpha_n + (\mu+\Gamma_n)\sin\alpha_n\right]}{(\mu+\Gamma_n)R_n + \xi B_n\left[(\mu+\Gamma_n)\cos\alpha_n - \Gamma_n\sin\alpha_n\right]}, \tag{C8}$$

$$\Gamma_n = (n\omega c_G\rho_G\lambda_G)^{0.5}\cos\left(\frac{\pi}{4}\right), \tag{C9}$$

$$\mu = 4\sigma T_M{}^3 + c_p\rho C_HU_M + l\rho\beta C_HU_M\Delta, \tag{C10}$$

$$\xi = 4\sigma T_M{}^3 + c_p\rho C_HU_M, \tag{C11}$$

where

$$\Delta = \left(\frac{dq_{sat}}{dT}\right)_{T_M}. \tag{C12}$$

In this equation, B_n and α_n are the amplitude and phase of air temperature, respectively, (defined by Eq. (9)), S_n is the amplitude of the solar radiation (Eq. 8)), and L_n the amplitude of the downward longwave radiation (in the present paper, L_n is set to zero). Note that the surface temperature by Eq. (Cl) is used only for estimating the soil heat flux G_D.

The concept of calculating G is almost same as those of Brutsaert (1982, p. 145–152). In this study, estimation of the heat balance, however, is conducted by using the bulk transfer method. Hence, the salient points of difference between them is to use Eqs. (C5)–(C12) that correspond to the bulk transfer equation.

2) Seasonal variation of soil heat flux G_Y

The values of G_D are calculated under the assumption that the daily averaged value of the soil heat flux G_M is zero. In fact, G_M actually has a tendency to vary seasonally by about ±10 $\mathrm{Wm^{-2}}$. Using observed values of T_s, the time series of the seasonal variation of surface temperature T_{sY} is given as

$$T_{SY} = T_{SYM} + \sum_{n=1}^{5} A_{Yn} \cos(n\omega_Y t - \phi_{Yn}). \quad (C13)$$

The values of G_Y can then be estimated by the following equation

$$G_Y = \sum_{n=1}^{5} G_{Yn} \cos\left(n\omega_Y t - \phi_{Yn} + \frac{\pi}{4}\right), \quad (C14)$$

where

$$G_{Yn} = A_{Yn}(n\omega_Y c_G \rho_G \lambda_G)^{0.5}, \quad (C15)$$

and

$$\omega_Y = 1.992 \times 10^{-7}\ \mathrm{s}^{-1}. \quad (C16)$$

Here, A_{Yn} and ϕ_{Yn} are the amplitude and phase of T_{sY}, respectively.

The daily averaged values T_{sM}, H_M, and lE_M can be calculated by averaging diurnal values obtained by substituting values of G into Eq. (7). The consideration of incorporating G_M does not seriously affect the accuracy of lE_M, but raises the accuracy of T_{sM} to within about ±0.5°C.

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植生地における熱収支モデルおよび水田地帯への適用

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熱収支の日変化および季節変化を計算する新たな植生モデルを示した。モデル内のパラメーターは簡略化されており、計算に必要なデータは、観測地点近くの温度、風、湿度、降水量、日照時間であり、アメダスデータから読み取ったものである。モデルの適用性が 1993, 1994 年の 2 年間にわたる水田地帯の観測結果によって検証された。本モデルにおける計算結果は日平均観測値とよく対応している。全期間において、本モデルの日平均地表面温度、日平均顕熱フラックス、および日平均潜熱フラックスの計算値と観測値の差はそれぞれ 1.7 °C, ±13 $\mathrm{Wm^{-2}}$、および±16 $\mathrm{Wm^{-2}}$であった。また、年蒸発散量は約 600 $\mathrm{mmy^{-1}}$と推定された。水田の年蒸発散量は浅い水面からの年蒸発量 (564 $\mathrm{mmy^{-1}}$) と森林からの年蒸発散量 (692 $\mathrm{mmy^{-1}}$) の中間に入ることが分かった。水田の夏期植生期間における遮断蒸発量の蒸発散量に対する割合は、18 % (1993 年：冷夏) と 9 % (1994 年：暑夏) である。1993 年の降水量は多いために遮断蒸発量は多いが、蒸発散量は少ない。