BULK PARAMETERIZATION FOR A VEGETATED SURFACE AND ITS APPLICATION TO A SIMULATION OF NOCTURNAL DRAINAGE FLOW

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Abstract. Two simple models are presented for describing the surface energy budget above vegetated surfaces. One is the traditional single-source model that includes only one energy budget equation for the entire canopy-soil system, and the other is the double-source model that includes separate energy budget equations for the vegetation canopy and the underlying soil surface. In both models, the bulk transfer coefficients needed to solve the energy budget equations are parameterized as functions of leaf area index, leaf transfer coefficients, and soil surface roughnesses to obtain the best fit to values calculated by a standard multilayer-canopy model. The validity of these models was tested by comparing their performance with that of the multilayer-canopy model for simulation of the surface energy balance and nocturnal drainage flow above vegetation. Results show that the double-source model gives reliable estimations for all cases ranging from sparse to dense vegetation covers; the single-source model is only applicable to dense, fully-covered vegetation. It is also shown that sparse vegetation weakens nocturnal drainage flow, since it isolates the cool underlying soil surface from the atmosphere above the canopy. This phenomenon cannot be described by a traditional single-source model incorporated commonly in many atmospheric models; however, the double-source model adequately describes this process.

1. Introduction

Since half of the earth's land-surface is covered by vegetation, it is very important to explore a realistic representation of the energy budget of a vegetated surface. In a canopy, however, there are complicated interactions among the energy budgets of individual leaves and the soil surface, radiative energy transfer, and turbulent transfer of energy and momentum. Hence, modeling of the energy budget of a vegetated surface is more difficult than for a single plain surface. Recently, many numerical models have been developed for describing the energy budget of a vegetated surface. Almost all of them have a canopy structure to represent the vegetation cover and to incorporate the complicated interactions previously mentioned. The most realistic expression of canopy micrometeorology is given by a multilayer canopy model. However, it must be accompanied by some intricate calculations.

A rough estimation of the total energy balance can be obtained by combining the energy budget equation and the bulk transfer equations, considering the vegetation and the underlying soil surface as a lumped system. In fact, the surface properties of many atmospheric numerical models are represented only by the roughness length for momentum, sensible heat, and water vapor (or using the

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Stanton number and the Dalton number) without considering canopy structure. This type of model is referred to as a single-source model. The well-known Penman-Monteith equation is an example of such a model. Thus, it is worth investigating whether such single-source models can give realistic representations of the energy balance for vegetated surfaces.

For a sparse vegetation cover, however, the energy balance of the underlying soil surface is rather different from that of the vegetation canopy because of differences in available radiation, wind speed, and evaporative efficiency. Evaporative efficiency is the most serious problem because the Bowen ratio is strongly dependent on this parameter. For moderately dense vegetation where the leaf area index (LAI) is near unity, about half of the incident solar radiation can penetrate to the soil surface. Thus, a very large temperature difference can exist between transpiring vegetation and a dry soil surface. Also, the radiative energy absorbed by the soil surface is less easily transformed into sensible and latent heat than that absorbed by the vegetation canopy because of the low wind speed near the soil surface. In such cases, large errors may occur using single-source models. The double-source model incorporates this feature by using respective energy budget equations for the vegetation canopy and for the soil surface.

In this paper, single-source and double-source models are compared with results from a multilayer canopy model for simulating the surface energy balance and the nocturnal drainage flow above vegetation.

2. Models

2.1. Multilayer canopy model

The multilayer canopy model used in the present study is modified from the version presented by Kondo and Watanabe (1992; hereafter Paper A). Governing equations are the energy budget equations for individual canopy layers and the soil surface, equations for the turbulent and radiative energy transfer between layers, and the momentum transfer equation. Exchanges of momentum, sensible heat, and water vapor between individual leaves and the ambient atmosphere are expressed using the transfer coefficients of a leaf for momentum (c_d), sensible heat (c_h), and water vapor (c_e). Also, exchanges between the soil surface and the lowest atmosphere within the vegetation are expressed in terms of the roughness lengths for momentum (z_{0s}), sensible heat (z_{Ts}), and water vapor (z_{qs}). The radiative energy transport is approximated by a 2-stream model, which is modified from the version presented in Paper A to take solar zenith angle into consideration. The turbulent transfers of momentum, sensible heat, and water vapor are expressed by K-theory. From the model, profiles of wind speed, air and leaf temperature, specific humidity, and fluxes of momentum and energy can be calculated within and above the vegetation canopy, given meteorological conditions, profiles of leaf-area density and leaf-transfer coefficients, and roughness lengths of the soil surface. The validity of the model has already been confirmed by comparing calculated results with observations. A more detail description is presented in Paper A.

2.2. Single-source model

Only the energy budget of the entire system is considered in the single-source model. The equation is

$$R^{\downarrow} = \sigma T_{sfc}^4 + H + lE + G\,, \tag{1}$$

where

$$R^{\downarrow} = (1 - ref)S^{\downarrow} + L^{\downarrow}\,. \tag{2}$$

Here $R^{\downarrow}$ is available incident radiation, σ is the Stefan–Boltzmann constant, T_{sfc} is apparent surface temperature, H is sensible heat flux, lE is latent heat flux (l is specific latent heat, E is water vapor flux), and G is energy storage rate. Also, *ref* is the albedo of the entire canopy-soil system, $S^{\downarrow}$ is downward shortwave radiation, and $L^{\downarrow}$ is downward longwave radiation. The emissivity of the system is assumed to be unity because of its structural complexity. Fluxes of momentum, sensible heat, and latent heat are written by the following bulk transfer equations

$$\tau = \rho C_M u^2\,, \tag{3}$$

$$H = c_p \rho C_H u(T_{sfc} - T)\,, \tag{4}$$

$$lE = l\rho C_E u[q_{\text{sat}}(T_{sfc}) - q]\,, \tag{5}$$

where c_p and ρ are specific heat and density of the air, respectively, $q_{\text{sat}}(T_{sfc})$ is saturation specific humidity at a temperature of T_{sfc}, u, T, and q are wind speed, air temperature, and specific humidity at a reference level z_a, respectively, and C_M, C_H and C_E are the bulk transfer coefficients for momentum, sensible heat and latent heat, respectively. When the meteorological conditions ($S^{\downarrow}$, $L^{\downarrow}$, u, T, q) and the energy storage rate G are given, the apparent surface temperature T_{sfc}, the sensible heat flux H, and the latent heat flux lE can be evaluated from Equations (1), (2), (4), and (5) in terms of the bulk transfer coefficients C_H and C_E.

The bulk transfer coefficients can be written by

$$C_M = k^2 \left[\ln \frac{z_a - d}{z_0} + \Psi_m(\zeta)\right]^{-2}, \tag{6}$$

$$C_H = k^2 \left[\ln \frac{z_a - d}{z_0} + \Psi_m(\zeta)\right]^{-1} \left[\ln \frac{z_a - d}{z_T} + \Psi_h(\zeta)\right]^{-1}, \tag{7}$$

$$C_E = k^2 \left[\ln \frac{z_a - d}{z_0} + \Psi_m(\zeta) \right]^{-1} \left[\ln \frac{z_a - d}{z_q} + \Psi_e(\zeta) \right]^{-1}, \tag{8}$$

where

$$\zeta = \frac{z_a - d}{L}.$$

Here d is zero-plane displacement height, z_0, z_T, and z_q are the roughness lengths of the entire canopy-soil system for momentum, sensible heat, and water vapor, respectively, and Ψ_m, Ψ_h, and Ψ_e are the correction terms expressing the thermal stability effect on the profiles of wind speed, air temperature, and specific humidity, respectively (see a review paper by Yaglom, 1977). Also, L is the Monin-Obukhov length

$$L = \frac{\Theta_0 (\tau/\rho)^{3/2}}{kg(H/c_p \rho)}, \tag{9}$$

where, Θ_0 is the reference potential temperature, k is the von Kármán constant, and g is gravitational acceleration. The zero-plane displacement height and the roughness lengths are functions of vegetation height (h), leaf area index (LAI), leaf transfer coefficients (c_d, c_h, and c_e), and roughness lengths of the soil surface (z_{0s}, z_{Ts}, and z_{qs}). These surface parameters are expressed by the following empirical formulae which were determined to obtain the best fit to values calculated by the multilayer canopy model (see Paper A for the evaluation methods using the multilayer model).

The zero-plane displacement height is analytically expressed by

$$\frac{d}{h} = 1 - \frac{1}{A^+} [1 - \exp(-A^+)], \tag{10}$$

where

$$A^+ = \frac{c_d \mathrm{LAI}}{2k^2},$$

The empirical equation for the momentum roughness is

$$\left(\ln \frac{h - d}{z_0} \right)^{-1} = \left[1 - \exp(-A^+) + \left(-\ln \frac{z_{0s}}{h} \right)^{-1/0.45} \exp(-2A^+) \right]^{0.45}. \tag{11}$$

The value of z_0 can be calculated from the combination of Equations (10) and (11). Assigning as $F_\chi = c_h/c_d$ (or c_e/c_d) and $z_{\chi s} = z_{Ts}$ (or z_{qs}), respectively, the scalar roughness length for the case of $F_\chi = 0$ represented by $z_\chi^\dagger$ can be calculated from

$$\left(\ln \frac{h-d}{z_\chi^\dagger}\right)^{-1} = \frac{1}{-\ln(z_{\chi s}/h)} \left[\frac{P_1}{P_1 + A^+ \exp(A^+)}\right]^{P_2}, \tag{12}$$

where

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

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

Using the value of $z_\chi^\dagger$, the scalar roughness length z_χ ($= z_T$ or z_q) for $0 < F_\chi \leqslant 1$ can be calculated from

$$\left(\ln \frac{h-d}{z_0}\right)^{-1} \left(\ln \frac{z-d}{z_\chi}\right)^{-1} = C_\chi^\infty \left[1 - \exp(-P_3 A^+) + \left(\frac{C_\chi^0}{C_\chi^\infty}\right)^{1/0.9} \exp(-P_4 A^+)\right]^{0.9}, \tag{13}$$

where

$$\frac{1}{C_\chi^0} = \ln \frac{h-d}{z_0} \ln \frac{h-d}{z_\chi^\dagger},$$

$$C_\chi^\infty = \frac{-1 + (1 + 8F_\chi)^{1/2}}{2},$$

$$P_3 = [F_\chi + 0.084 \exp(-15F_\chi)]^{0.15},$$

$$P_4 = 2F_\chi^{1.1}.$$

Figures 1 to 3 show the zero-plane displacement height and the roughness lengths calculated for vertically uniform vegetation (the leaf area per unit volume is constant with height). The figures show that these equations give good approximations for the surface parameters calculated by the multilayer canopy model. The small values of z_χ in the figure are expected for water vapor (z_q) for small values of c_e, such as non-transpirating leaves (e.g., in a deciduous forest in late autumn; in crops just before harvest). It should be noted that these equations describe the relationship between the resistances in the Penman–Monteith equation and vegetation properties (LAI, h, c_d, F_χ, z_{0s}, $z_{\chi s}$) since

$$r_a = \frac{1}{C_H u}, \tag{14}$$

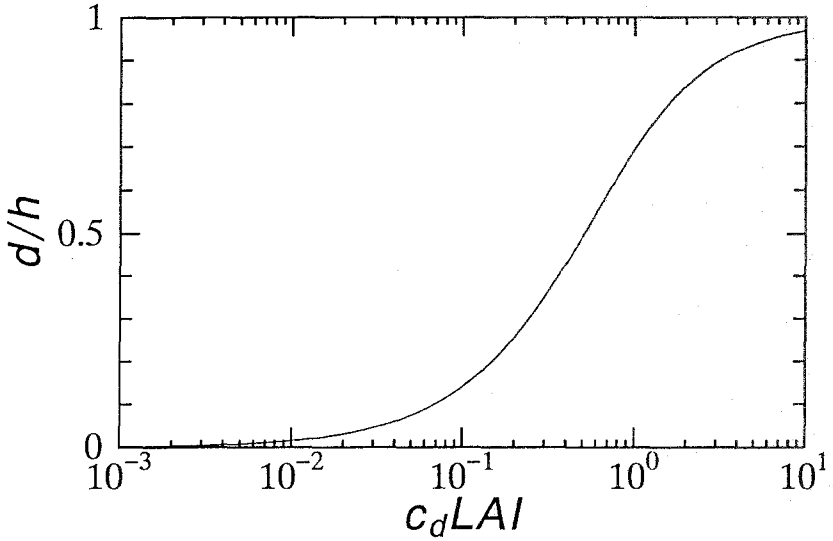

Fig. 1. The zero-plane displacement height d normalized by the vegetation height h calculated for vertically homogeneous vegetation. These values are expressed exactly by Equation (10).

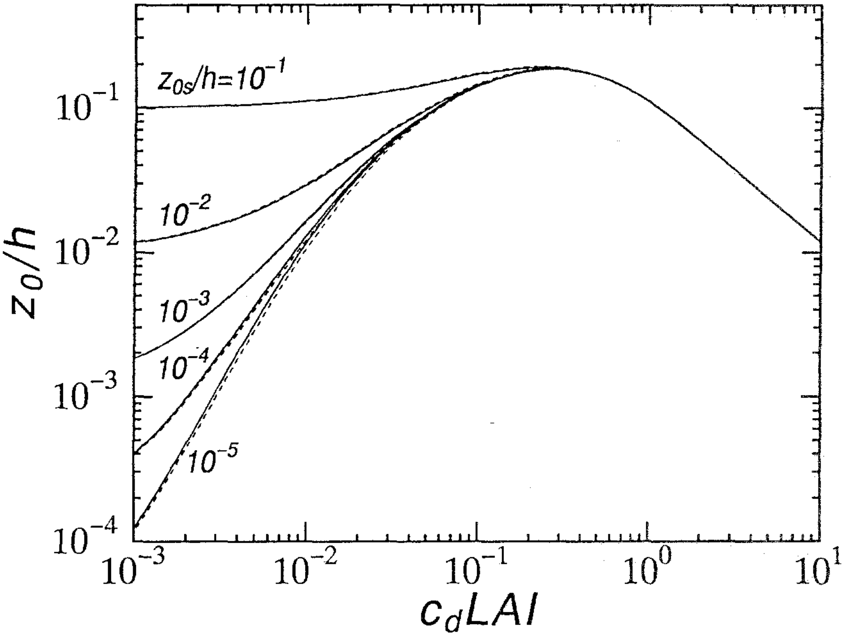

Fig. 2. The momentum roughness-length z_0 normalized by the vegetation height h calculated by the multilayer model (solid lines) and those from Equations (10) and (11) (broken lines) for various values of z_{0s}/h.

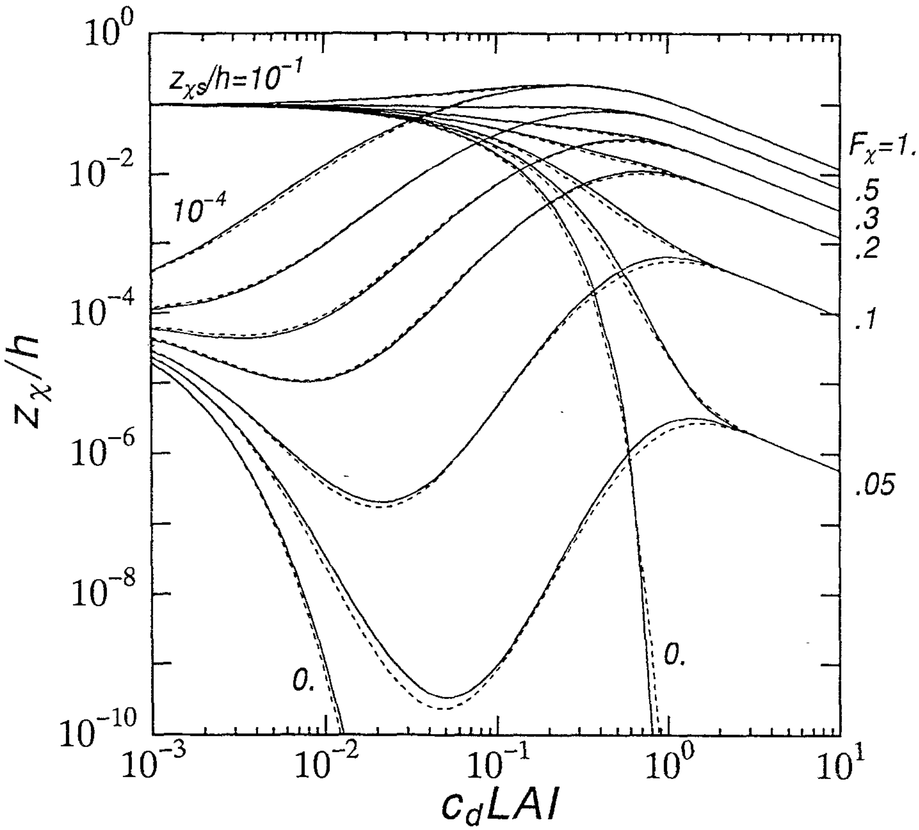

Fig. 3. Same as Figure 2 but for the scalar roughness z_χ ($=z_T$ or z_q) for two cases of $z_{0s}/h = z_{xs}/h = 10^{-1}$ and 10^{-4}, and for various values of F_χ ($=c_h/c_d$ or $=c_e/c_d$ for z_T or z_q, respectively).

$$r_a + r_c = \frac{1}{C_E u}, \tag{15}$$

where r_a is the boundary-layer resistance, and r_c the canopy resistance.

2.3. Double-source model

In the double-source model, both the energy budget of the vegetation canopy and that of the underlying soil surface are considered. However, Equation (3) is also used for describing momentum exchange. The energy budget equation for a soil surface is

$$m_S S_n + m_L L^{\downarrow} + (1 - m_L)\sigma T_c^4 = \sigma T_g^4 + H_g + lE_g + G, \tag{16}$$

and that for a vegetation canopy is

$$(1 - m_S)S_n + (1 - m_L)(L^{\downarrow} + \sigma T_g^4) = 2(1 - m_L)\sigma T_c^4 + H_c + lE_c, \tag{17}$$

where

$$S_n = (1 - ref)S^{\downarrow} , \tag{18}$$

$$m_S = \exp(-F \cdot \mathrm{LAI} \cdot \sec Z) , \tag{19}$$

$$m_L = \exp(- F \cdot \mathrm{LAI} \cdot d_f) . \tag{20}$$

Here m_S and m_L are the transmittances of the vegetation canopy for shortwave and longwave radiation, respectively, S_n is net shortwave radiation at the canopy top, *ref* is the albedo of the entire canopy-soil system, and subscripts *g* and *c* denote soil surface and vegetation canopy, respectively. Also, *F* is a factor representing the mean orientation of leaves (Ross, 1975), *Z* is solar zenith angle, and d_f is the diffusive factor for longwave radiation, which is set to unity in this paper for simplicity.

In previous canopy models, H_g and lE_g were always modeled as proportional to differences in temperature and humidity between soil and air within the vegetation, respectively. In the present double-source model, the soil surface is assumed to exchange directly with the atmosphere above the vegetation canopy. Although this may not represent true field conditions, the model can easily describe the difference in the energy balance between vegetation and the soil surface. The model is also capable of representing the direct interaction between the soil and the atmosphere above a vegetation canopy caused by penetrating wind gusts which are typical in forests. Fluxes are thus written as

$$H_g = c_p \rho C_{Hg} u(T_g - T) , \tag{21}$$

$$lE_g = l\rho C_{Eg} u[q_{\mathrm{sat}}(T_g) - q] , \tag{22}$$

$$H_c = c_p \rho C_{Hc} u(T_c - T) , \tag{23}$$

$$lE_c = l\rho C_{Ec} u[q_{\mathrm{sat}}(T_c) - T] , \tag{24}$$

where C_{Hg} and C_{Eg} are the bulk transfer coefficients for the soil surface, and C_{Hc} and C_{Ec} are those for the vegetation canopy. Total fluxes H, lE and $L^{\uparrow}$ are then expressed as

$$H = H_g + H_c , \tag{25}$$

$$lE = lE_g + lE_c , \tag{26}$$

$$L^{\uparrow} = \sigma T_R^4 = m_L \sigma T_g^4 + (1 - m_L)\sigma T_c^4 , \tag{27}$$

where $L^{\uparrow}$ is upward longwave radiation, and T_R is the radiative surface temperature. For the case of $T_g = T_c = T_{sfc}$, fluxes calculated by the double-source model must be the same as those by the single-source model. Hence the relations

$$C_H = C_{Hg} + C_{Hc} \, , \tag{28}$$

$$C_E = C_{Eg} + C_{Ec} \tag{29}$$

must hold. For the case of very sparse vegetation, $C_H = C_{Hg}$ and $C_E = C_{Eg}$. For very dense vegetation, C_{Hg} and C_{Eg} can be ignored. C_{Hg} and C_{Eg} are calculated as follows,

$$C_{Hg} = k^2 \left[\ln \frac{z_a - d}{z_{0g}} + \Psi_m(\zeta_g) \right]^{-1} \left[\ln \frac{z_a - d}{z_{Tg}} + \Psi_h(\zeta_g) \right]^{-1} , \tag{30}$$

$$C_{Eg} = k^2 \left[\ln \frac{z_a - d}{z_{0g}} + \Psi_m(\zeta_g) \right]^{-1} \left[\ln \frac{z_a - d}{z_{qg}} + \Psi_e(\zeta_g) \right]^{-1} , \tag{31}$$

where

$$\zeta_g = \frac{z_a - d}{L_g} \, .$$

Here z_{0g}, z_{Tg} and z_{qg} are the roughness lengths, which express the effects of the soil surface on the profiles above the canopy, and z_a is the reference level where values of u, T and q are given [the same as those in Equations (6)–(8)]. Also, L_g is the modified Monin-Obukhov length that describes the buoyancy effect on the exchange between the soil surface and the above-canopy atmosphere. In the present study, L_g is assigned as

$$L_g = \frac{\Theta_0 (\tau/\rho)^{3/2}}{kg(H_g/c_p\rho)} \, . \tag{32}$$

Values of z_{0g}, z_{Tg} and z_{qg} are calculated by the multilayer canopy model for the imaginary case that momentum and scalars are passively transferred only to/from the soil surface within the simulated wind field. The calculations are as follows: (i) the wind profile is calculated by the multilayer model with the specified values of c_d and z_{0s}, (ii) fluxes are then calculated in this wind field by setting $c_d = c_h = c_e = 0$, and (iii) values of z_{0g}, z_{Tg} and z_{qg} are finally calculated from the relationships between the fluxes and the differences in wind speed (u), temperature ($T_s - T$), and specific humidity [$q_{\text{sat}}(T_s) - q$], respectively. Here T_s is the soil surface temperature calculated by the multilayer model. Additional details are presented in Paper A. Using these calculations, values of z_0, z_{Ts}, and z_{qs} are small for dense vegetation since turbulent transfer through the canopy layer is suppressed. For very sparse vegetation, values of z_{0g}, z_{Tg} and z_{qg} approach asymptotic values of z_{0s}, z_{Ts} and z_{qs}, respectively. For practical calculations, values of z_{0g} and z_{xg} ($= z_{Tg}$ or z_{qg}) are also empirically expressed by

$$\left(\ln \frac{h-d}{z_{0g}}\right)^2 = \ln \frac{h-d}{z_0} \ln \frac{h-d}{z_0^{\dagger}}, \tag{33}$$

and

$$\ln \frac{h-d}{z_{0g}} \ln \frac{h-d}{z_{\chi g}} = \ln \frac{h-d}{z_0} \ln \frac{h-d}{z_{\chi}^{\dagger}}, \tag{34}$$

where the term $\ln[(h-d)/z_0]$ has been expressed by Equation (11), and the terms $\ln[(h-d)/z_0^{\dagger}]$ and $\ln[(h-d)/z_{\chi}^{\dagger}]$ can be expressed by the same formula as Equation (12) redefining $z_{\chi s} = z_{0s}$ (or z_{Ts} or z_{qs}) for calculating z_{0g} (or z_{Tg} or z_{qg}), respectively. C_{Hc} and C_{Ec} can then be evaluated respectively from Equations (28) and (29) with values of C_{Hg} and C_{Eg} calculated from Equations (30) and (31), and those of C_H and C_E from Equations (7) and (8).

2.4. Vegetation and Soil Parameters

The three models require specification of vegetation and soil parameters, such as LAI, h, c_d, F_{χ}, z_{0s} and $z_{\chi s}$. It is difficult to assign values, especially leaf transfer coefficients and soil roughness lengths, but there are several ways to infer them. Many experimental and theoretical studies have been performed to investigate values of c_d and c_h, which are dependent on leaf shape and the Reynolds number, etc. (e.g., Brutsaert, 1979). However, the value of c_e is controlled by plant physiological processes through stomatal behavior, as well as leaf shape and Reynolds number. On this problem, many formulae have been proposed for various kinds of plants (e.g., Jones, 1983), and they can be incorporated in the present energy-budget models. On the other hand, values of z_{0s} and z_{Ts} depend on micro-geometrical characteristics of the soil surface. Kondo and Yamazawa (1986) showed that values of z_{0s} and z_{Ts} of a snow surface are clearly related to the ruggedness of small-scale surface protrusions (wavelength <0.1 m). A similar relationship must exist for a soil surface. Also, in cultivated fields, these values can be directly observed before planting or after harvesting. The value of z_{qs}, however, is strongly affected by soil moisture. (Here, the saturation specific humidity is adopted as representative of soil surface humidity.) Therefore models that predict soil moisture must be incorporated (e.g., Kondo *et al.*, 1992) if the present models are used in a prognostic way.

In the present study, the vegetation and soil parameters are specified as constants for simplicity. It should be noted that the values adopted in the next section are merely examples used to compare simplified models and the multilayer model, without considering a specific type of vegetation.

3. Model Application

3.1. Diurnal Variation of Surface Energy Balance

For the case where soil heat flux is negligible, surface energy fluxes on a horizontally homogeneous vegetated surface have been previously calculated by the single-source and double-source models (see Paper A). These findings indicate that in general the single-source model gives good estimations, but results may be somewhat erroneous for moderately dense vegetation. The double-source model, however, can correctly represent the fluxes for all cases ranging from sparse to dense vegetation.

To determine model reliability for predicting the time variation of the energy balance, the soil heat flux G is included in all three models by using the force-restore method (e.g., Bhumralkar, 1975), i.e.,

$$G = \left(\frac{c_s \rho_s \lambda_s}{2\omega}\right)^{1/2} \left[\frac{\mathrm{d}T_s}{\mathrm{d}t} + \omega(T_s - T_{s0})\right], \tag{35}$$

where T_s is the soil surface temperature, T_{s0} is the daily mean soil-surface temperature, c_s, ρ_s and λ_s are specific heat, density and heat conductivity of soil, respectively, and ω is the angular frequency of the diurnal oscillation. The assumption $T_s = T_{sfc}$ is used in the single-source model, while $T_s = T_g$ is assumed for the double-source model. For the boundary conditions, solar radiation is given by the empirical formula for a clear day (Kondo and Miura, 1983), and the air temperature is given by

$$T = T_m + \delta T\left[\sin\left(\omega t - \frac{2\pi}{9}\sin \omega t - \frac{7\pi}{9}\right) + 0.211\right], \tag{36}$$

where T_m is daily mean air temperature and δT is the amplitude. Equation (36) was originally given by Takemasa *et al.* (1988) to represent the temperature of the dry soil layer, but is used here for air temperature, which has a minimum value at 6:00 and a maximum value at 14:00 (LST). Also, downward longwave radiation ($L^{\downarrow}$), wind speed (u), and specific air humidity (q) are assumed constant. The reference level is 30 m above the soil surface (20 m above the top of the vegetation). Specified values are as follows:

Solar radiation	For the summer solstice day at latitude 38°
Downward longwave radiation	$L^{\downarrow} = 365\ \mathrm{W\,m^{-2}}$
Wind speed	$u = 3\ \mathrm{m\,s^{-1}}$
Air temperature	$T_m = 20\ ^{\circ}\mathrm{C}$, $\delta T = 5\ ^{\circ}\mathrm{C}$
Specific air humidity	$q = 10\ \mathrm{g\,kg^{-1}}$
Daily mean soil surface temperature	$T_{s0} = 23\ ^{\circ}\mathrm{C}$

Also, values for the soil and vegetation properties are:

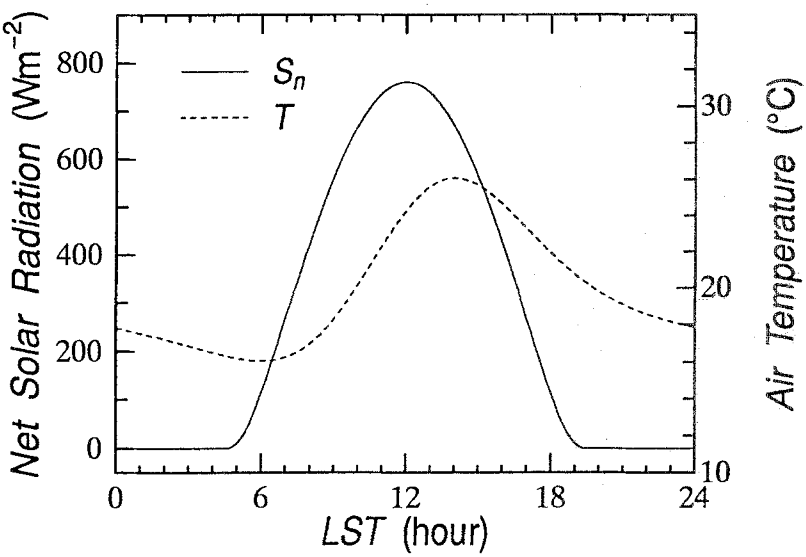

Fig. 4. Diurnal variations of net solar radiation (solid line) and air temperature (dashed line) given by the simulation.

Surface albedo	$ref = 0.1$
Heat capacity of soil	$c_s \rho_s = 2.64 \times 10^6\ \mathrm{J\,m^{-3}\,K^{-1}}$
Heat conductivity of soil	$\lambda_s = 0.47\ \mathrm{W\,m^{-1}\,K^{-1}}$
Roughness length of soil surface	$z_{0s} = z_{Ts} = z_{qs} = 0.001\ \mathrm{m}$
Vegetation height	$h = 10\ \mathrm{m}$
Leaf transfer coefficients	$c_d = 0.2$, $c_h = 0.06$, $c_e = 0.006$
Leaf orientation factor	$F = 0.5$

Diurnal variations of solar radiation and air temperature are shown in Figure 4. The leaf area density is assumed to be vertically uniform, and two cases are considered: sparse vegetation (LAI = 0.5); and dense vegetation (LAI = 5). Atmospheric stability is assumed to be always neutral in these calculations.

Assuming $\mathrm{d}T_s/\mathrm{d}t = 0$ at 6:00 on the first day of calculation, all variables were initialized as balanced solutions, and the calculation was continued for 48 h. Results are shown in Figures 5 and 6 for the second day. In the case of dense vegetation (Figure 5), fluxes of sensible and latent heat calculated by the single-source and double-source models are almost equal to those by the multilayer model. However, the soil surface temperature T_s and the soil heat flux G cannot be predicted by the single-source model. The value of T_s calculated by the single-source model reaches a maximum around 12:00, coinciding with the solar radiation maximum. Values of T_s calculated by the other two models reach their maxima later (near 14:00) when air temperature is highest. The prediction difference of the single-source model is much greater when the soil surface is covered by sparse

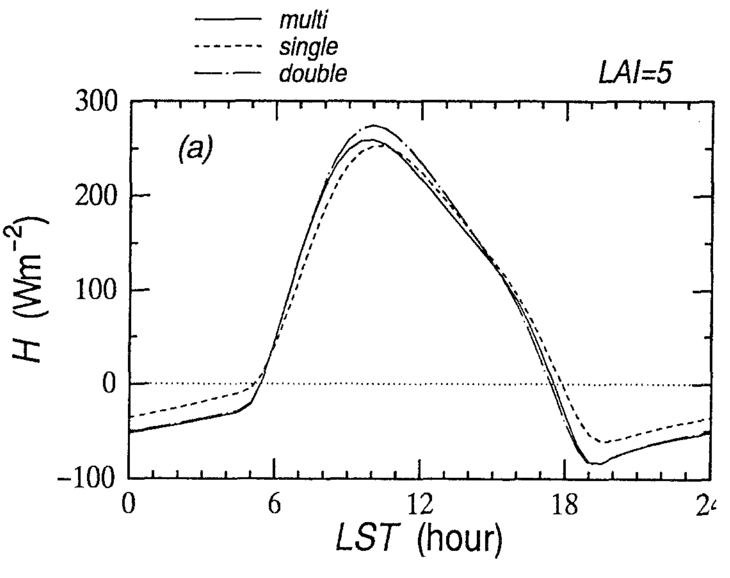

Fig. 5(a).

vegetation (Figure 6); for this case, the double-source model still gives acceptable predictions. These show that the double-source model is reliable for predicting the time variation of the surface energy balance above vegetation.

3.2. Nocturnal drainage flow over vegetated slopes

To estimate the regional energy balance or evapotranspiration, the influence of vegetation on the regional air circulation in complex topography must be considered. Many numerical studies have been conducted dealing with this problem. In particular, Yamada (1981) tried to simulate nocturnal drainage flow observed in complex terrain with a three-dimensional mesoscale atmospheric model in which the surface boundary conditions were given by a single-source model. However, the simulated wind speed near the surface was much larger than observations. Yamada speculated that this discrepancy resulted from smoothing of the measured wind speed profile because of sparse spatial measurements, and from the decrease in measured wind speed by the form drag induced by trees (not sufficiently described by the single-source model).

3.2.1. *Equations*

A one-dimensional model is adopted, considering an ideally homogeneous slope with a constant angle α. The Coriolis terms are neglected for simplicity. The slope is also assumed to be fully covered by vegetation in which the leaf area density a is uniformly distributed (a is constant for $0 \leqslant z \leqslant h$). The x-axis is taken to be parallel to the down-slope direction, and the z-axis is perpendicular to the slope.

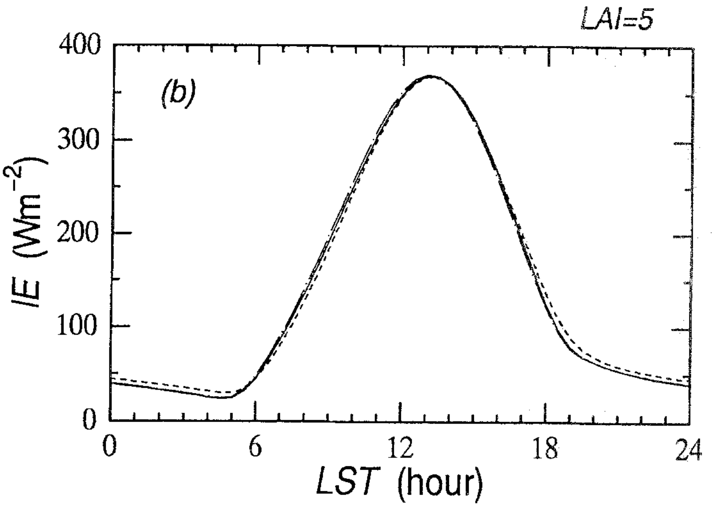

Fig. 5(b).

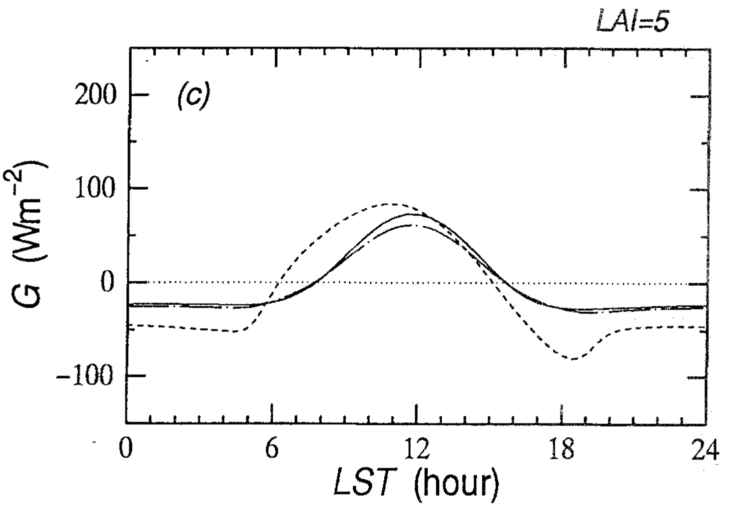

Fig. 5(c).

The governing equations including the multilayer canopy model are

$$\frac{\partial u}{\partial t} = g\,\frac{\Theta - \theta}{\Theta}\sin\alpha + \frac{\partial}{\partial z}\left(K_M\,\frac{\partial u}{\partial z}\right) - c_d a u^2\,, \tag{37}$$

$$\frac{\partial \theta}{\partial t} = \Gamma u \sin\alpha + \frac{\partial}{\partial z}\left(K_H\,\frac{\partial \theta}{\partial z}\right) - c_h a u(\theta - \theta_c)\,, \tag{38}$$

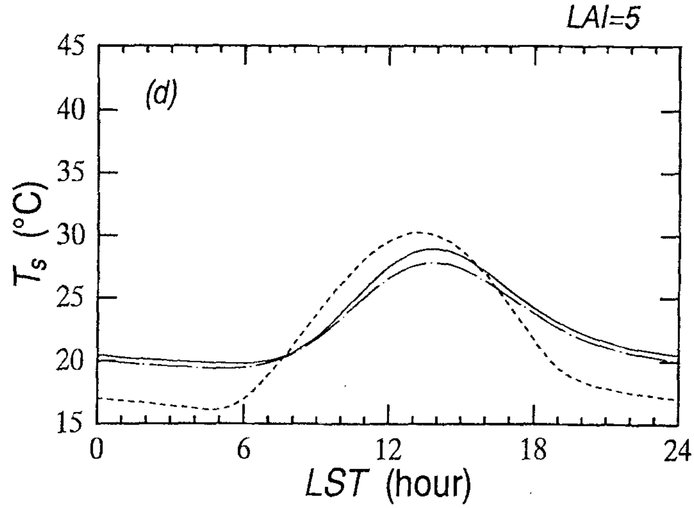

Fig. 5(d).

Fig. 5. Diurnal variations of (a) sensible heat flux, (b) latent heat flux, (c) energy storage rate, and (d) soil surface temperature calculated by the multilayer model (solid line), by the single source model (broken line), and by the double source model (dotted-dashed line). Case of dense vegetation (LAI = 5).

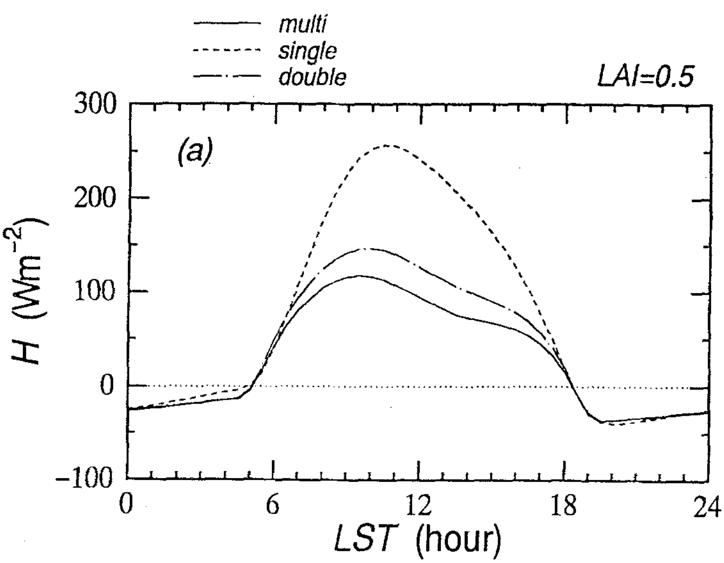

Fig. 6(a).

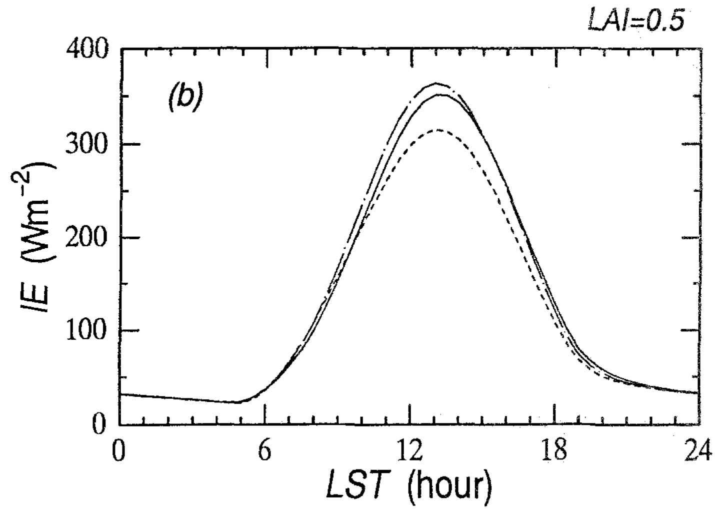

Fig. 6(b).

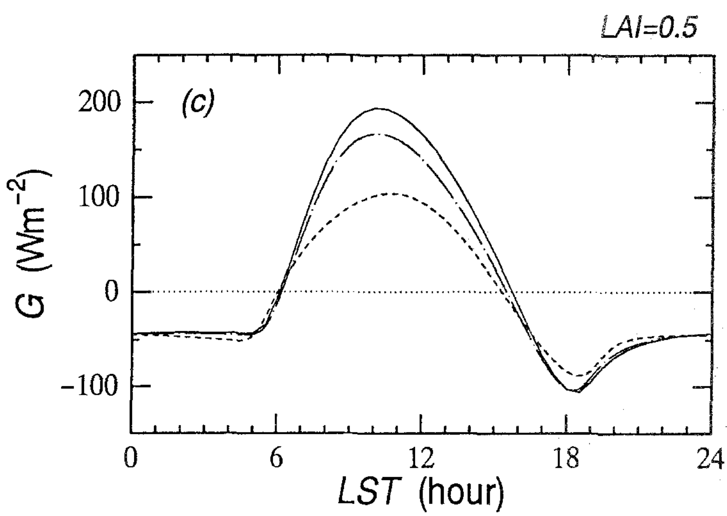

Fig. 6(c).

where u is the down-slope wind speed, θ is potential temperature, θ_c is potential temperature of individual leaves, and K_M and K_H are the turbulent diffusivities for momentum and sensible heat, respectively. Also, Θ is the potential temperature of the reference atmosphere given by

$$\Theta = \Theta_0 + \Gamma z \cos \alpha , \tag{39}$$

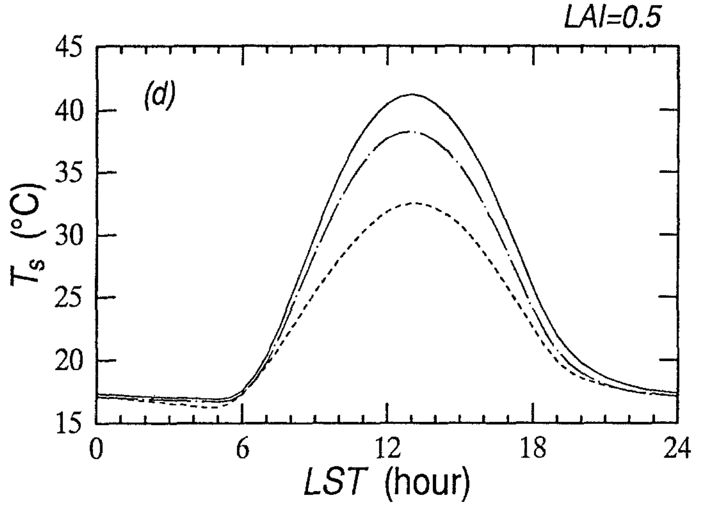

Fig. 6(d).

Fig. 6. Same as Figure 5 but for sparse vegetation (LAI = 0.5).

where Θ_0 is the reference potential temperature and Γ is the vertical potential temperature gradient. Boundary conditions are $u = 0$ and $\theta = \Theta$ at the top boundary level (z_{top}); and $u = 0$ and $\theta = \theta_s$ at $z = 0$, where θ_s is the potential temperature of the soil surface. Turbulent diffusivities are expressed by

$$K_M = \frac{l^2(z)}{\varphi_m^2(\zeta)} \left| \frac{\partial u}{\partial z} \right| , \tag{40}$$

$$K_H = \frac{l^2(z)}{\varphi_m(\zeta)\varphi_h(\zeta)} \left| \frac{\partial u}{\partial z} \right| , \tag{41}$$

where

$$\zeta = \frac{z - d'(z)}{L(z)} . \tag{42}$$

Here φ_m and φ_h are the nondimensional shear functions. To avoid unrealistically small values of K_M and K_H near the levels of maximum and minimum wind speed, the running-mean average over the domain $[z - l, z + l]$ is operated on Equations (40) and (41). Applying the mixing-length model presented by Watanabe and Kondo (1990) to the present case of vertically uniform vegetation, the mixing length l and the zero-plane displacement height d' at each level z can be written by

$$l(z) = k[z - d'(z)] , \tag{43}$$

$$z - d'(z) = \begin{cases} \dfrac{2k^2}{c_d a}\left[1 - \exp\left(-\dfrac{c_d a z}{2k^2}\right)\right], & 0 \leqslant z \leqslant h\,; \\ z - d'(h), & z > h\,. \end{cases} \tag{44}$$

It should be noted that $d'(h)$ corresponds to the conventional zero-plane displacement d, which is expressed by Equation (10). Equations (43) and (44) were derived theoretically regarding $(z - d')$ as the mean free path-length of the turbulent motion within and above vegetation. Equation (44) means that the eddy size in a sparse canopy (when $c_d a$ is small) is controlled by the soil surface as well as leaves, and even in a dense canopy, the limitation on eddy size by the soil surface is still important at lower levels (when z is small).

Values of $\theta_c(z)$ and θ_s are determined as balanced solutions of the energy budget equations for an individual leaf and for the soil surface and equations for longwave radiation listed below.

$$F[L^{\downarrow}(z) + F^{\uparrow}(z)] = 2F\sigma\theta_c^4(z) + c_p \rho c_h u(z)[\theta_c(z) - \theta(z)]\,, \tag{45}$$

$$L^{\downarrow}(0) = \sigma\theta_s^4 + c_p \rho C_{Hs} u(z_1)[\theta_s - \theta(z_1)]\,, \tag{46}$$

$$\frac{\mathrm{d}L^{\downarrow}(z)}{\mathrm{d}z} = FaL^{\downarrow}(z) - Fa\sigma\theta_c^4(z)\,, \tag{47}$$

$$\frac{\mathrm{d}L^{\uparrow}(z)}{\mathrm{d}z} = -FaL^{\uparrow}(z) + Fa\sigma\theta_c^4(z)\,, \tag{48}$$

where z_1 is the level of the lowest calculating node. Boundary conditions for Equations (47) and (48) are $L^{\downarrow} = L_h$ at the canopy top ($z = h$) and $L^{\uparrow} = \sigma\theta_s^4$ at the soil surface ($z = 0$). The bulk transfer coefficient of the soil surface C_{Hs} is calculated from

$$C_{Hs} = \left(\frac{1}{k}\ln\frac{z_1}{z_{0s}}\right)^{-1}\left(\frac{1}{k}\ln\frac{z_1}{z_{Ts}}\right)^{-1}. \tag{49}$$

C_{Hs} should not be confused with C_{Hg} of the double-source model.

For the single-source and double-source model calculations, on the other hand, the governing equations are

$$\frac{\partial u}{\partial t} = g\frac{\Theta - \theta}{\Theta}\sin\alpha + \frac{\partial}{\partial z}\left(K_M\frac{\partial u}{\partial z}\right), \tag{50}$$

$$\frac{\partial \theta}{\partial t} = \Gamma u \sin\alpha + \frac{\partial}{\partial z}\left(K_H\frac{\partial \theta}{\partial z}\right). \tag{51}$$

Boundary conditions for the top boundary are the same as those for the multilayer model calculation. At the lower boundary, the sensible heat flux is calculated by the single-source or double-source model. The influences of vegetation are reflected only in the lower boundary conditions in these calculations.

Values assigned for the calculation are:

Vegetation height	$h = 10$ (m)
Leaf area index	LAI = 0.5, 5
Leaf transfer coefficients	$c_d = 0.2$, $c_h = 0.06$
Leaf orientation factor	$F = 1$
Roughness length of the soil surface	$z_{0s} = z_{Ts} = 0.001$ (m)
Slope angle	$\alpha = 5$ (°)
Reference potential temperature	$\Theta_0 = 293.15$ (K)
Potential temperature gradient	$\Gamma = 0.003$ (K m^{-1})
Atmospheric longwave radiation	$L_h = 320$ (W m^{-2})
Top boundary level	$z_{\text{top}} = 400$ (m)

The top boundary level was defined to be high enough not to affect the drainage flow profiles, after pre-calculations with a higher top boundary level. The functions φ and Ψ presented in the paper by Kondo (1975) were adopted. The soil heat flux was not considered. Calculations were initiated from the radiation balance condition, and continued until stationary solutions were obtained.

Before turning to results, some general comments should be made on the use of K-theory, which cannot handle counter-gradient or non-gradient fluxes observed commonly in vegetation canopies (e.g., Denmead and Bradley, 1985). Some reasons for use of K-theory in such situations are as follows. First, an exact simulation of profiles is not our objective. It is not necessary for performing comparisons between simple models and the multilayer model and for investigating qualitatively the influence of vegetation upon the drainage flow. Second, smoothing of the turbulent diffusivities reduces the difficulty of K-theory. Also, the difference between K-theory and any other theory may not be very large since the turbulent kinetic energy cannot be large enough in stably stratified situations to cause strong counter-gradient flux. Third, K-theory has computational simplicity compared with higher-order closure models including the turbulent diffusion terms in their flux budget equations.

3.2.2. *Simulated results and discussion*

Simulated profiles of wind speed and potential temperature are shown in Figure 7 for dense vegetation (LAI = 5). It can be seen from the results for the multilayer model (solid line) that the flow is well-mixed by the thermally induced convection below the level where the vegetation is most strongly cooled. However, the existence of a minimum in the windspeed profile at about $z = 0.6h$ indicates that the exchange is not so active between the upper and lower portion of the vegetation layer, and only the upper portion can exchange with the atmosphere above. In such a situation, the vegetated surface simply acts as a displaced rough surface.

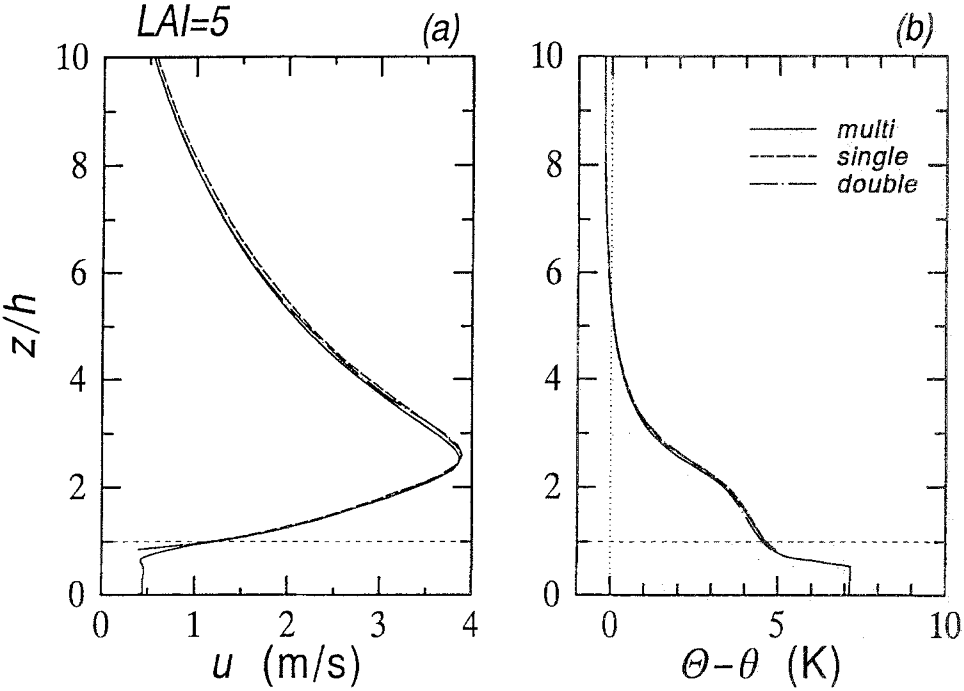

Fig. 7. Calculated profiles of (a) wind speed and (b) potential temperature plotted against the level normalized by vegetation height. Case of nocturnal drainage flow over a densely vegetated slope (LAI = 5).

Hence, both the single-source and the double-source models yield the same wind and temperature profiles as the multilayer model at levels above the vegetation.

Calculated results for a sparse vegetation cover (LAI = 0.5) are shown in Figure 8. The single-source model overestimates the wind speed and the atmospheric cooling as has been indicated by Yamada (1981). Also, it is found from multilayer-model calculations that the wind speed and atmospheric cooling are less over sparse vegetation than over dense vegetation (compare the solid lines in Figures 7 and 8). The roughness length for the sparse vegetation is larger than for the dense vegetation (Figure 2). However, profiles of wind speed and potential temperature calculated by the single-source model are not very different for the sparse and the dense covers (see broken lines in Figures 7 and 8). Clearly the weaker flow over sparse vegetation is not caused only by the larger roughness length. According to the results for the multilayer model in Figure 8, the following explanation may be given for behaviour of drainage flow over sparse vegetation. The soil surface is strongly cooled by the radiative energy emission through the sparse vegetation canopy. This coolness, however, cannot be transferred efficiently to the atmosphere above the vegetation since the turbulence within the vegetation is suppressed by the canopy elements and the stably stratified air. In other words, the

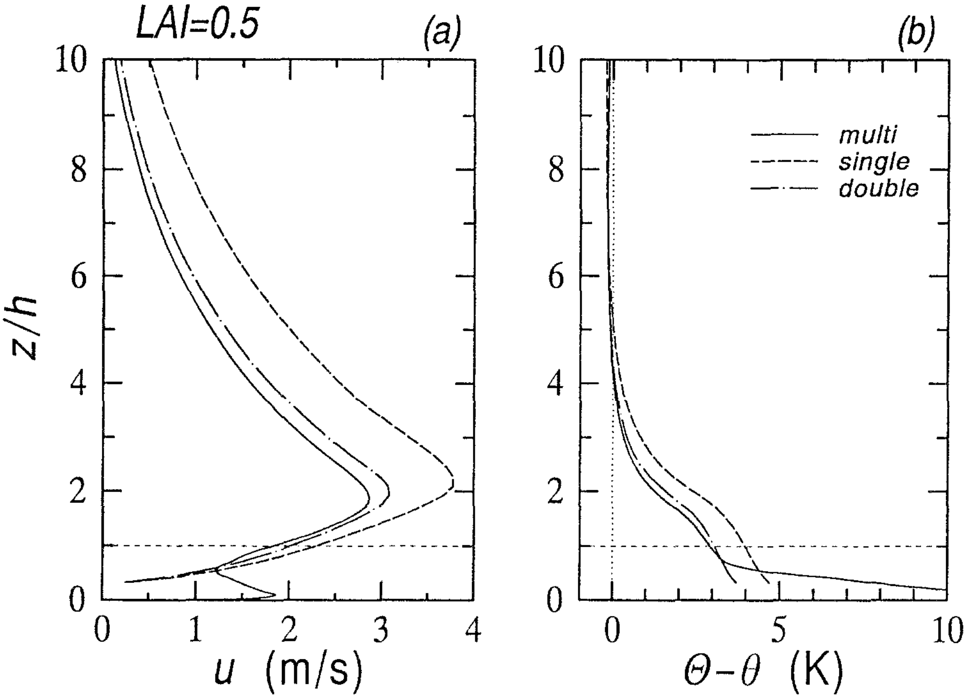

Fig. 8. Same as Figure 7 but for sparse vegetation (LAI = 0.5).

outgoing radiative energy is compensated mainly by cooling of the soil layer, not by atmospheric cooling. This is different from the case of dense vegetation in which only the upper portion of the vegetation is cooled by radiation, inducing atmospheric cooling. Consequently, drainage flow over sparse vegetation cannot develop so fully as over dense vegetation. The single-source model cannot represent this feature and accordingly fails to simulate these profiles. The double-source model, on the other hand, which can express the temperature contrast between the soil surface and the vegetation, is applicable to drainage flow over a sparsely vegetated slope.

In order to clarify the error caused by the single-source model more clearly, another calculation was performed for the case where soil and vegetation temperatures were constant without changing other conditions. Setting the temperature as $\theta_c(z) = \theta_s = 288.15$ K (i.e., $\Theta_0 - \theta_c = 5$ K), profiles were obtained as shown in Figure 9. It should be noted that the double-source model is exactly the same as the single-source model in the case of $\theta_c = \theta_s$ [see Equations (28) and (29)]. In this case, results from the single-source model are almost identical to those from the multilayer model. This indicates that the single-source model can correctly express the form drag induced by trees, and therefore errors in Figure 8 and also

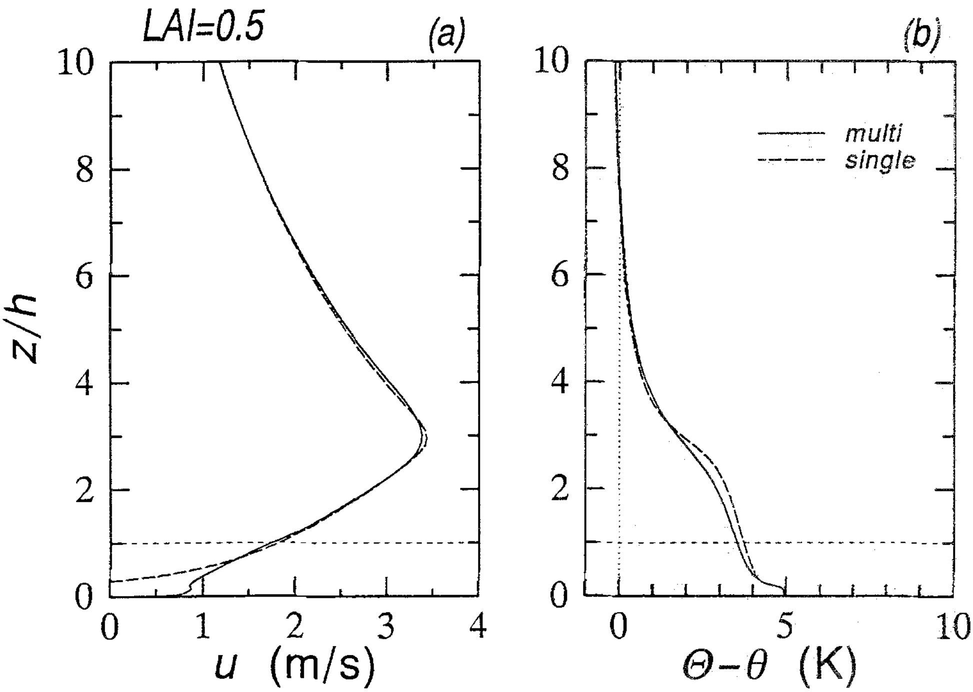

Fig. 9. Same as Figure 8 but for the results from the constant-temperature calculation.

in the paper by Yamada (1981) are due to the temperature contrast between the soil and vegetation, which cannot be represented by the single-source model.

4. Conclusions

Two important atmospheric processes are influenced by vegetation cover. First, vegetation increases the surface roughness through the action of form drag, which can be expressed by the traditional single-source model. Second, the turbulent energy exchange between the underlying soil surface and atmosphere is suppressed by vegetation while radiative energy transfer still exists. The latter process is very important for a sparse vegetation cover, or a less than full canopy. The nocturnal drainage flow over a sparsely vegetated slope has been calculated to be less as a result of this process. Consequently, it is necessary to take into account the difference between the soil surface and the vegetation layer when representing a vegetated surface in atmospheric numerical models. It was also found that the double-source model presented here can satisfactorily express the influence of vegetation in spite of its simplicity, as judged against a multilayer model.

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