Validation and Sensitivity Analysis of a New Atmosphere–Soil–Vegetation Model

HARUYASU NAGAI
Japan Atomic Energy Research Institute, Ibaraki, Japan

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ABSTRACT

This paper describes details, validation, and sensitivity analysis of a new atmosphere–soil–vegetation model. The model consists of one-dimensional multilayer submodels for atmosphere, soil, and vegetation and radiation schemes for the transmission of solar and longwave radiations in canopy. The atmosphere submodel solves prognostic equations for horizontal wind components, potential temperature, specific humidity, fog water, and turbulence statistics by using a second-order closure model. The soil submodel calculates the transport of heat, liquid water, and water vapor. The vegetation submodel evaluates the heat and water budget on leaf surface and the downward liquid water flux. The model performance was tested by using measured data of the Cooperative Atmosphere–Surface Exchange Study (CASES). Calculated ground surface fluxes were mainly compared with observations at a winter wheat field, concerning the diurnal variation and change in 32 days of the first CASES field program in 1997, CASES-97. The measured surface fluxes did not satisfy the energy balance, so sensible and latent heat fluxes obtained by the eddy correlation method were corrected. By using options of the solar radiation scheme, which addresses the effect of the direct solar radiation component, calculated albedo agreed well with the observations. Some sensitivity analyses were also done for model settings. Model calculations of surface fluxes and surface temperature were in good agreement with measurements as a whole.

1. Introduction

The heat and water exchanges between the atmosphere and the ground surface are important processes for environmental studies. Heat energy and water supplied to the ground surface as solar and longwave radiation and precipitation are redistributed to the ground and the atmosphere through the ground-surface processes, becoming the driving force of atmospheric phenomena. Thus, the processes of heat and water exchanges at the ground surface play an important role in determining the meteorological and climatological conditions. Moreover, the processes of heat and water exchanges at the ground surface are essential components for pollutant movement. However, the behavior of heat and water in the atmosphere–soil–vegetation system, which can be found everywhere on the globe, is very complicated and is still poorly understood. Many factors including meteorological, hydrological, and biological factors control the processes of heat and water exchanges in the system. The property of water that has solid, liquid, and vapor phases under the natural environment makes the processes more complicated because phase change of water causes heat exchange.

In recent decades, many models that deal with the interaction between the atmosphere and the ground surface have been developed. Models that represent soil as multilayer and deal with vegetation as one or two canopy layers (Deardorff 1978; Sellers et al. 1986; Dickinson et al. 1993) or as the part of the ground-surface parameterization (Pan and Mahrt 1987; Noilhan and Planton 1989; Pleim and Xiu 1995; Chen et al. 1996) are widely used in atmospheric models such as general circulation, regional, and mesoscale models. The objectives of these models are mainly in the improvement of bottom boundary condition of the atmospheric models by accurately evaluating the ground-surface impact on the atmosphere. While multilayer representation is used for the soil, considering the importance of influences due to soil moisture and heat capacity, rather simple representation is preferred for the vegetation, considering the computational constraint and the difficulty of specifying model parameters of vegetation in practical use. On the other hand, multilayer vegetation models have also been developed (Willson and Shaw 1977; Yamada 1982; Meyers and Paw U 1986, 1987; Raupach 1987; Shaw and Schumann 1992). The focus of these studies is mainly on the turbulence transfer of momentum, heat, and moisture within and above vegetation canopy, and simple representation is used for heat and water exchanges among the atmosphere, soil, and vegetation compared to the former models. Some models (Yamada 1982; Meyers and Paw U 1987) calculate the radiation transfer in canopy and the heat and water budget on leaves with multilayer vegetation representation.
However, few models of this type include soil processes to deal with the heat and water exchanges between the atmosphere and the ground surface.

A new atmosphere–soil–vegetation model named SOLVEG is developed (Yamazawa and Nagai 1997; Nagai and Yamazawa 1999). The model is designed to simulate processes in the atmosphere–soil–vegetation system in a more realistic way than do commonly used models by using multilayer expression for the atmosphere, soil, and vegetation and avoiding uncertain parameterizations as far as possible. The present model is one-dimensional and applicable only to the near-surface layer (about 10 m) of the atmosphere by using surface meteorological observations or the atmospheric model’s outputs as the top boundary conditions. This model is intended for studies to understand the ground-surface processes and to design better parameterizations for the ground-surface boundary conditions of atmospheric models. Furthermore, it is also planned to use the model directly as the ground-surface boundary model by coupling it to three-dimensional atmospheric models.

To test the model’s performance, simulations were carried out for measured data of the Cooperative Atmosphere–Surface Exchange Study (CASES; LeMone et al. 2000). The surface measurements at a winter wheat field were used. It was found that the surface fluxes did not satisfy the energy balance. The cause of this energy imbalance was discussed and the sensible and latent heat fluxes were corrected. Determination of vegetation parameters and the necessity of considering the effect of a direct solar radiation component in radiation transmission are discussed. Sensitivity analyses are also done to evaluate the influence of the model settings on the surface fluxes. The model performance examined in this paper is the predictability of the surface fluxes as the overall response of soil and vegetation canopy.

2. Model description

SOLVEG consists of one-dimensional multilayer submodels for atmosphere, soil, and vegetation, and radiation schemes for calculating the transmission of solar and longwave radiation fluxes in the canopy layer. Physical processes of water and heat exchange considered in the model are schematically shown in Figs. 1 and 2. Basic equations for the submodels and radiation schemes are described in the following sections. Model parameters for the soil and vegetation are also summarized. Since the model deals with soil and vegetation separately, the word “soil surface” is used to represent the interface between soil and the atmosphere, distinguishing it from the ground surface that is generally
used as the mixture of soil surface and vegetation. Definitions of main variables used in the equations are listed in appendix A, and a detailed description of the model equations is in appendix B.

**a. Atmosphere submodel**

The atmosphere submodel calculates atmospheric variables by numerically solving prognostic equations for horizontal wind speed components $u$ and $v$, potential temperature $\theta$, specific humidity $q_a$, turbulence kinetic energy $e$, turbulence length scale multiplied by turbulence kinetic energy $e^\lambda$, and fog water content $w_f$, which is the mass of liquid water per unit mass of moist air. Using $\phi$ for the atmospheric variables, one-dimensional diffusion equations used in the model are described in the same form as

$$\frac{\partial \phi}{\partial t} = \frac{\partial}{\partial z} K_z \frac{\partial \phi}{\partial z} + F_{\phi},$$

where $K_z$ is the vertical turbulence diffusivity for atmospheric variables. The last term $F_{\phi}$ is a forcing term.

**b. Soil submodel**

The soil submodel prognostically calculates soil temperature $T_s$, volumetric soil water content $\eta_w$, and specific humidity of air in soil pore space $q_s$. For the soil water processes, the transports in liquid and gas phases, phase change, and root uptake are considered. Heat conduction through plant root is assumed to be negligible. When soil water exceeds the saturated soil water content, water in excess is assumed to be stored at soil surface and does not run off in this model. This assumption is applicable to flat surfaces. However, the surface runoff process should be included when the model is coupled to three-dimensional atmospheric models. The model does not have processes of water in solid phase.

To calculate soil temperature, a one-dimensional heat conduction equation expressed as

$$\frac{\partial T_s}{\partial t} = \frac{\partial}{\partial z} K_z \frac{\partial T_s}{\partial z} + \frac{H_s}{C_s \rho_s} - \frac{C_w E_w \partial T_s}{C_s \rho_s \partial z},$$

is used. The second and last terms on the right-hand side express the heat source or sink term due to latent heat exchange caused by evaporation or condensation of soil water and the heat transport caused by movement of liquid water in soil, respectively.

The volumetric soil water content $\eta_w$ is calculated by using a prognostic equation based on the Richards equation (Richards 1931).
\[
\frac{\partial \eta_w}{\partial t} = -\frac{1}{\rho_w} \left( \frac{\partial E^s}{\partial z} + E_v + E_n \right).
\] (3)

In addition to the liquid water flux in soil, the diffusion of water vapor in soil pores is also considered in the model. A one-dimensional diffusion equation for water vapor in soil is expressed as

\[
\frac{\partial[(\eta_w - \eta_0)q^s]}{\partial t} = \frac{\partial}{\partial z} \left[ D_{fs}(\eta_w) \frac{\partial q^s}{\partial z} + E^g \right],
\] (4)

where \((\eta_w - \eta_0)\) represents the volumetric fraction of air-filled pores, and \(f(\eta_w)\) is the coefficient for distortion of pore (tortuosity) at volumetric water content \(\eta_w\). Analogous to Jackson et al. (1974), the model uses \(f(\eta_w) = (\eta_{w0} - \eta_w)/1.5\).

c. Vegetation submodel

The vegetation submodel calculates temperature of vegetation \(T_c\), amount of leaf surface water \(w_s\), and vertical liquid water flux in the canopy \(P_r\). The model deals with vegetation canopy as multiple layers consisting of air with heat capacity and imaginary leaves without heat capacity. Sensible and latent heats are exchanged between leaf surface and canopy air so that the heat budget on the leaf surface holds at each layer. The exchanges of water vapor, sensible heat, and momentum by trunks, branches, and stems of vegetation are assumed to be implicitly included in the formulation for leaves because of their smaller surface area compared to that of leaves. The heat conduction by trunks, branches, and stems is neglected. The model does not consider water in solid phase. The fluxes of sensible heat, water vapor, and radiation fluxes are formulated so that their continuity holds at the canopy top.

The temperature of vegetation \(T_c\) is calculated from the heat budget on leaf surface. By neglecting the heat capacity of leaves and the energy consumption due to photosynthesis, a heat budget equation for unit leaf area is expressed as

\[
R_c = H_c + I(a) + L(a),
\] (5)

where the left-hand side is the net radiation and the terms on the right-hand side are the sensible and latent heat fluxes and cooling or warming by precipitation, respectively.

The leaf surface water increases due to interception of precipitation, capture of fog water, and condensation of water vapor, and decreases due to drip and evaporation. The amount of leaf surface water on unit leaf area \(w_s\) is calculated by

\[
\frac{dw_s}{dt} = E_{in} - E_{ev} + E_{cap} - P_r.
\] (6)

The vertical liquid water flux in canopy \(P_r\), which is positive downward, is determined from the water budget in canopy

\[
\frac{dP_r}{dz} = a(E_{in} - P_r) + E_{pr} - E_{col}.
\] (7)

Fog water has no vertical speed macroscopically, and is not included in \(P_r\). Parameter \(P_r\) includes the water flow along tree trunk; \(P_r\) at the outside of canopy is the precipitation intensity.

d. Radiation schemes

The reflection, absorption, and emission by vegetation and liquid water in the atmosphere are considered in the radiation transmission. Although the solar and longwave radiation fluxes are generally not isotropic, they are dealt with as isotropic downward and upward fluxes to construct formulations expressing the radiation transmission in the canopy. No spectral dependency of radiation is assumed. However, it was recognized in the model validation that the model did not simulate diurnal variations in albedo depending on the solar zenith angle since the radiation schemes are based on the assumption of isotropic radiation. Thus, two options are included in the solar radiation scheme to consider the nonisotropic properties of the direct radiation component.

The solar radiation transmission is calculated for downward and upward fluxes \(S^\downarrow\) and \(S^\uparrow\), separately in the model. Assuming the isotropic radiation, the transfer equations for downward and upward solar radiations are expressed as

\[
\frac{dS^\downarrow}{dz} = (aF_r + a'_{w} + A'_w)S^\downarrow - (aF_c + A'_c)S^\uparrow,
\] (8)

and

\[
\frac{dS^\uparrow}{dz} = -(aF_r + a'_{w} + A'_w)S^\downarrow + (aF_c + A'_c)S^\uparrow.
\] (9)

where the reflectivity \(A'_{w}\) and absorptivity \(a'_{w}\) of liquid water in the atmospheric layer are determined using the parameterization by Stephens (1978). Here, \(F_r\) is the projection coefficient of leaves and is calculated by \((B27)\).

Besides reflection, absorption, and transmission, emission from leaf surface and liquid water in the atmosphere is included in the longwave radiation transmission. The transmission equations for the downward longwave radiation flux \(L^\downarrow\) and the upward longwave radiation flux \(L^\uparrow\) are expressed as

\[
\frac{dL^\downarrow}{dz} = aF_L[L^\downarrow - (1 - \varepsilon_L)L^\uparrow - \varepsilon_L \sigma T^\uparrow]\n\]

\[
+ k_Lw_L(L^\downarrow - \sigma T^\uparrow) \quad \text{and} \quad \frac{dL^\uparrow}{dz} = aF_L[-L^\downarrow + (1 - \varepsilon_L)L^\downarrow + \varepsilon_L \sigma T^\uparrow]\n\]

\[
+ k_Lw_L(-L^\downarrow + \sigma T^\uparrow),
\] (10)

where coefficients \(k_l\) and \(w_l\) are determined using the parameterization by Stephens (1978).
To consider the nonisotropic properties of the direct solar radiation component, the model has two options. Option 1 considers the change in path length of the direct solar radiation component depending on the solar zenith angle. In option 1, the downward solar radiation is assumed to have only a direct component and its path length in the canopy is inversely proportional to the cosine of solar zenith angle. With this assumption, \( \delta z / \cos z \) is used instead of \( \delta z \) to calculate \( F_z \) by (B27) for the downward solar radiation flux.

The forward scattering of direct solar radiation component may occur on leaves with large inclination when the solar zenith angle is small. In option 2, its effect is included in the transmission calculation by subtracting the forward scattering term expressed as \( F_{\text{scat}} \) from the right-hand side of (8) and (9). Parameter \( F_{\text{scat}} \) is a function of solar zenith angle and depends on leaf inclination. It has been determined by comparing calculations with observations, as discussed in the validation section.

e. Soil and vegetation parameters

Soil parameters used in the model are the saturated volumetric soil water content \( \eta_{sw} \), the saturated hydraulic conductivity \( K_s \), the exponent in the Clapp and Hornberger equations \( b \), the wilting volumetric soil water content \( \eta_{wil} \), and the dry soil heat capacity \( C_{s,\text{dry}} \). These values were determined according to the U.S. Department of Agriculture (USDA) texture classes, by using data by Clapp and Hornberger (1978), McCumber (1980), and Chen and Dudhia (2001) based on Cosby et al. (1984). Parameter values are summarized as a lookup table for soil type to use in the model calculation.

Vegetation parameters are categorized into two types. One is leaf surface property, and the other is vegetation structure. The former includes the leaf surface reflectivity \( A_r \), the leaf surface emissivity \( e_r \), the maximum leaf surface water \( w_{sl} \), the leaf surface water at which evaporation is saturated \( w_{es} \), the drag coefficient of leaf \( c_D \), the heat exchange coefficient \( c_H \), the water vapor exchange coefficient \( c_F \), and the minimum stomatal resistance \( r_s,\text{min} \). The latter has the distribution of leaf area density and root distribution, which is expressed as the fraction of root uptake of water in each soil layer to total uptake. The model has no lookup table for these parameters, and they are given the values for actual vegetation conditions.

3. Validation with CASES data

The model validation was carried out by using observation data obtained by the first CASES field program in 1997, CASES-97 (LeMone et al. 2000). The focus of the validation here is mainly on the performance of the overall model in predicting the ground surface response, that is, the surface fluxes, to the meteorological input. A gridded dataset based on the CASES-97 observation generated by F. Chen and D. Yates (1999, personal communication) is used to make model inputs, that is, the initial values for variables of the atmosphere and soil submodels and time series of the top boundary values for the atmosphere submodel and external conditions such as precipitation and radiation. Model simulations are carried out for one grid point of the dataset corresponding to one of the surface flux stations of CASES-97. Ground surface heat and radiation fluxes calculated by the model were compared with the observations, concerning the diurnal variation and change in 32 days from 20 April to 21 May 1997.

a. Observation data

There are 12 surface flux stations sited at several types of land cover such as grassland, winter wheat, and bare ground in CASES-97. Of these stations, station 7 at a winter wheat field is the best-equipped station with the National Center for Atmosphere Research (NCAR) Atmosphere—Surface Turbulent Exchange Research (ASTER) facility (Businger et al. 1990) and measurements of downward and upward solar and longwave radiations. Besides the quality of data, a winter wheat field is easy to handle in the model because of its dense canopy and horizontal homogeneity. This site was selected to examine the model performance. Half-hour averages for solar and longwave radiation fluxes, net radiation, sensible and latent heat fluxes, and ground heat conduction were used for comparisons with calculations.

The quality check of surface flux observations was carried out. Both the sensible and latent heat fluxes were obtained by the eddy correlation method using measurements by a sonic anemometer. The ground heat conduction at the soil surface was calculated from the soil heat flux measured by heat flux plate and heat storage within the layer above the sensor. As described in the status report for surface data on the web page for CASES-97 surface station data archive (http://www.atd.ucar.edu/ssfip/projects/cases97/), all surface flux data were corrected for known problems. Since the surface fluxes—net radiation \( RN \), sensible heat flux \( SH \), latent heat flux \( LH \), and ground heat conduction \( GH \)—were measured separately, the surface energy budget \( EB \), expressed by

\[
EB = RN - (SH + LH + GH), \tag{12}
\]

can be used for the quality test of flux measurements (Kanemasu et al. 1992).

Time series of \( EB \) and \( RN \) for the test period are shown in Fig. 3. The energy budget is small during nighttime and large during daytime. Daytime \( EB \) is always positive and, moreover, \( EB \) does not cancel out in diurnal cycles and longer period. This indicates that there were some systematic errors in flux measurements or other factors affecting the surface energy balance.
The canopy heat storage, the energy consumption by photosynthesis, and the horizontal advection of sensible and latent heat are considered to be causes of the surface energy imbalance. The heat energy used to raise canopy temperature can be estimated as 12 W m\(^{-2}\) at largest by assuming a temperature rise of 20 K in a 6-h period in the morning and by assuming that the crop has a heat capacity equivalent to that of 3 kg m\(^{-2}\) of water. However, this heat storage by canopy becomes the energy source and reduces EB in evening and nighttime. According to Jones (1992), the typical rate of photosynthesis is 0.5–2.0 mg CO\(_2\) m\(^{-2}\) s\(^{-1}\) at maximum, which corresponds to a loss of heat energy of 8–32 W m\(^{-2}\).

The total of these energy terms cannot account for the missing energy in the observation during daytime. The influence by advection is difficult to be evaluated but considered to be small because of the large fetch, at least 200 m for the predominate winds as described on the CASES-97 data Web page. Also, no specific correlation is seen between EB and the surface wind data. Blanken et al. (1997) discussed this problem and indicated the possibility that local, large turbulent structures were not properly spatially averaged by measurements at one fixed point. To close the surface energy budget, they recalculated SH and LH by dividing the available energy (PN - GH - canopy storage) using the ratio of measured SH and LH. The same method was used to correct SH and LH in this study, although the canopy storage was neglected here.

### b. Model setting and vegetation parameters

The model grids in this application were set as follows. The atmosphere submodel had 10 layers with boundary heights of 0.1, 0.3, 0.5, 0.7, 1.0, 1.5, 3.0, 5.0, 8.0, and 12.0 m, respectively. The vegetation submodel uses the same layers as the atmosphere submodel except for the top one. Since the model expresses the structure of vegetation in terms of the leaf area density at each model layer, such fine lower layers were used to cope with the growth of winter wheat (0.3–0.7 m) during the simulation period. The influence of the model layer setting for the atmosphere and vegetation will be discussed later in the sensitivity analysis. In this study, wind data measured at 10 m were used as the input. Therefore, the height of top layer was set as 8–12 m and the observation data at 10 m were given as the upper boundary values. The soil submodel had seven layers with boundary depths of 0.02, 0.05, 0.1, 0.2, 0.5, 1.0, and 2.0 m, respectively. The soil temperature at the bottom layer (1–2 m) had a constant value of 10°C, and the initial temperature profile was made by interpolation between the surface and bottom values. The initial soil moisture had a homogeneous distribution of observed soil moisture at 5-cm depth. The influence of uncertainty in the initial soil moisture is also examined in the sensitivity analysis.

In this study, values from literature were used for most of the parameters. The emissivity of leaf surface \(\varepsilon_c\) was 0.98. The maximum leaf surface liquid water \(w_{ds}\) and the leaf surface liquid water at which evaporation is saturated \(w_{ds0}\) were assumed to have the same value of 0.2 kg m\(^{-2}\) (equivalent to 0.2 mm of water on the leaf surface), which was derived from the maximum water content of vegetation canopy by Noilhan and Planton (1989). For the drag coefficient of leaf \(c_D\), 0.2 by Meyers and Paw U (1986) was used. The heat exchange coefficient \(c_E\) and the water vapor exchange coefficient \(c_H\) had the same value of 0.06 given by using the relationship \(c_E/c_D = 0.3\) according to Kondo and Watanabe (1992). The minimum stomatal resistance \(r_{s,min}\) was 200 s m\(^{-1}\) according to Deardorff (1978). The reflectivity of leaf surface \(\rho_A\) was determined by using measurements of solar radiation, and details are described in the following section.

The morphological property of vegetation varies widely by vegetation type, growing stage, site, and so on, and parameters of vegetation structure; that is, leaf area density and root distribution, need to be given values by measuring the target vegetation. However, no such data are available for the winter wheat field of CASES-97 except for the plant height, which was measured occasionally during the observation period. In this study, the leaf area density was assumed to have a vertically homogeneous distribution and a constant value and was given the value by considering the measured height and leaf area index (LAI) from literature. The average height of winter wheat varied from about 0.3 m to the maximum height of about 0.7 m during the
c. Leaf surface reflectivity

The leaf surface reflectivity $A_c$ was determined by comparing the ratio of upward to downward solar radiation fluxes, that is, albedo, between calculations and observations. The measured albedo has a diurnal variation with large amplitude as shown in Fig. 4. The value changes between the maximum after sunrise and before sunset and the minimum at local noon (about 1800 UTC). The minimum value decreases during the period. These diurnal variations of albedo indicate that the transmission of solar radiation in vegetation canopy strongly depends on the solar zenith angle and the assumption of isotropic radiation is not applicable. It is considered that the variation of albedo is caused by the difference of how far the solar radiation penetrates into vegetation canopy depending on the solar zenith angle. When the solar zenith angle is nearly 90°, the direct solar radiation is mostly shielded by the top thin layer of the canopy. In this case the reflection of solar radiation by the whole vegetation canopy becomes large and is considered to close to that by a single leaf. Therefore, it can be said that the $A_c$ is estimated to have the maximum value of observed albedo (0.3) or larger. On the other hand, the direct solar radiation can penetrate further into vegetation canopy and absorbed more by the vegetation or the soil surface when the solar zenith angle is small. Thus, the decrease of minimum value of albedo during the period can be explained by the change of solar zenith angle at noon.

Comparisons of the observed and calculated albedo were carried out. Since the effect by nonisotropic prop-
Fig. 5. Comparison of albedo for the first 10 days of calculation period between observation (open circles) and calculations (solid lines) by (a) the original solar radiation scheme with \( A_s = 0.3 \), (b) using option 1 with \( A_s = 0.2 \), and (c) using both option 1 and option 2 with \( A_s = 0.3 \).

Property of solar radiation was significant as described above, such a large diurnal variation of observed albedo was not calculated by the formulations of solar radiation transmission based on the isotropic radiation as shown in Fig. 5a. However, calculations agree well with the observation on cloudy days (25 and 26 April). It is considered that the solar radiation was almost isotropic on these days. By using option 1, which considers the change in path length of direct solar radiation with incidence angle, and tuning the value of \( A_s \), the solar radiation scheme simulated diurnal variation of observed surface albedo much better than the original scheme as shown in Fig. 5b. However, much smaller value of \( A_s \) (0.2) than that estimated from observation (0.3 or larger) had to be used. Also, the variation range of calculated albedo is still narrower than that of observation. This result shows that the change in path length of direct solar radiation component alone is not enough to explain the wide range of diurnal variation of observed surface albedo, and other factors need to be considered. Since winter wheat has rather vertical leaves, the effect of forward scattering of direct solar radiation component is considered to be large. Thus, option 2, which considers the forward scattering of direct solar radiation, was applied in addition to option 1. Here, the forward scattering coefficient \( F_{sc} \) was assumed to be a function of the solar zenith angle and determined by trial and error. By using \( A_s = 0.3 \) and \( F_{sc} = 0.6 (1 - \sin z_s) \), the scheme simulated diurnal variation of observed surface albedo successfully except for cloudy days in the calculation period as shown in Fig. 5c. This case was used for the performance test.

d. Sensitivity analyses

The influence of the model setting on the calculations was examined concerning the model layers for the atmosphere and vegetation, the initial soil moisture distribution, one of the vegetation parameters, and leaf area density. The surface fluxes calculated using different model setting were mainly compared.

For the atmosphere and vegetation layers, five different settings, which are listed in Table 1, were applied and calculated surface fluxes for the 32 days of calculation period were compared. The average difference (AD) and root mean square difference (rmsd) of the surface fluxes from the results by the original setting are summarized in Table 1. The differences in the surface flux calculations are large for the settings with different resolutions for the canopy layer compared to...
Setting volumetric soil water content decreases with depth by (A) 0.1 and (B) 0.05 and increases by (C) 0.05 and (D) 0.1, m initial soil moisture distributions and one with the homogeneous initial soil moisture. For the assumed initial soil moisture settings, the moisture was examined by assuming four different dis-same resolution or higher than the original setting. That the layer just above the canopy should have the large differences in the surface fluxes. It is concluded in the profiles. These differences in the pro-iles of the wind speed, temperature, and specific humidity change drastically around the canopy top, the different resolution of layers for this part causes slight differences in the profiles. These differences in the profiles of the atmospheric variables combine and make the relatively large differences in the surface fluxes. It is concluded that the layer just above the canopy should have the same resolution or higher than the original setting.

The influence of the uncertainty in the initial soil moisture was examined by assuming four different distributions besides the homogeneous one, which was used in the comparison with observations. In the four assumed settings, the volumetric soil water content decreased or increased with depth by 0.05 and 0.1 m⁻¹. The values at the surface were the same for all the settings. There were rainfalls on days 13, 19, 25, and 29 in calculation. Temporal changes of soil water distributions are shown in Fig. 6. The difference of near-surface soil water content among the settings increases by the influence of large difference of soil moisture at deep layer at first, then decreases with time by the influence of rainfalls. However, the difference remains until the end of calculation. The AD and rmsd of the surface fluxes from the results by the original setting are summarized in Table 2. The differences are relatively large for the sensible and latent heat fluxes and small for the net radiation and the ground heat conduction. Biases of the sensible and latent heat fluxes increase as the difference in the initial soil moisture profile becomes larger, but almost cancel out with each other. It seems that only Bowen ratio is affected by the different setting of the initial soil moisture. On the other hand, negative biases of the net radiation and the ground heat conduction are seen only in the settings with less moisture at deeper layers (settings A and B in Table 2). It is considered that the soil surface albedo increases as the surface soil moisture decreases, causing the decrease of the absorption of solar radiation at the soil surface. Although these influences are recognized, they are not significant as a whole.

### Table 1. AD and rmsd of net radiation (RN), sensible heat flux (SH), latent heat flux (LH), and ground heat conduction (GH) between calculations with different settings for the atmosphere and vegetation layers. The differences from the calculations with the original setting are summarized for five different settings listed below.

<table>
<thead>
<tr>
<th>Setting</th>
<th>RN (W m⁻²)</th>
<th>AD</th>
<th>Rmsd</th>
<th>SH (W m⁻²)</th>
<th>AD</th>
<th>Rmsd</th>
<th>LH (W m⁻²)</th>
<th>AD</th>
<th>Rmsd</th>
<th>GH (W m⁻²)</th>
<th>AD</th>
<th>Rmsd</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>2.6</td>
<td>4.5</td>
<td>4.8</td>
<td>9.8</td>
<td>2.1</td>
<td>5.0</td>
<td>0.1</td>
<td>2.6</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>−3.2</td>
<td>5.1</td>
<td>−3.1</td>
<td>12.8</td>
<td>−1.0</td>
<td>10.2</td>
<td>0.8</td>
<td>5.1</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C</td>
<td>−0.3</td>
<td>1.3</td>
<td>−0.5</td>
<td>3.1</td>
<td>0.2</td>
<td>1.7</td>
<td>0.0</td>
<td>1.1</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>D</td>
<td>−0.3</td>
<td>1.5</td>
<td>0.2</td>
<td>3.9</td>
<td>−0.5</td>
<td>2.3</td>
<td>0.1</td>
<td>1.4</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>E</td>
<td>−1.5</td>
<td>4.9</td>
<td>−1.2</td>
<td>11.1</td>
<td>−0.7</td>
<td>5.9</td>
<td>0.5</td>
<td>4.2</td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>

1 Original setting: canopy layers (0.1, 0.3, 0.5, and 0.7 m) and upper layers (1, 1.5, 3, 5, 8, and 12 m).
2 Setting A: higher resolution for canopy layers (0.1, 0.2, 0.3, 0.4, 0.5, 0.6, and 0.7 m).
3 Setting B: lower resolution for canopy layers (0.3 and 0.7 m).
4 Setting C: higher resolution for upper layers (1, 1.5, 2, 3, 4, 6, 8, and 12 m).
5 Setting D: lower resolution for upper layers (1, 4, 8, and 12 m).
6 Setting E: lower resolution for upper layers (2, 4, 8, and 12 m).

### Table 2. AD and rmsd of net radiation, sensible heat flux, latent heat flux, and ground heat conduction between calculations using assumed initial soil moisture distributions and one with the homogeneous initial soil moisture. For the assumed initial soil moisture settings, the volumetric soil water content decreases with depth by (A) 0.1 and (B) 0.05 and increases by (C) 0.05 and (D) 0.1, m⁻¹.

<table>
<thead>
<tr>
<th>Setting</th>
<th>RN (W m⁻²)</th>
<th>AD</th>
<th>Rmsd</th>
<th>SH (W m⁻²)</th>
<th>AD</th>
<th>Rmsd</th>
<th>LH (W m⁻²)</th>
<th>AD</th>
<th>Rmsd</th>
<th>GH (W m⁻²)</th>
<th>AD</th>
<th>Rmsd</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>−0.7</td>
<td>2.7</td>
<td>2.9</td>
<td>6.1</td>
<td>−2.6</td>
<td>4.4</td>
<td>−1.1</td>
<td>4.7</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>−0.2</td>
<td>1.3</td>
<td>1.7</td>
<td>4.3</td>
<td>−1.6</td>
<td>2.8</td>
<td>−0.4</td>
<td>3.1</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C</td>
<td>0.1</td>
<td>1.1</td>
<td>−1.2</td>
<td>3.4</td>
<td>1.2</td>
<td>2.3</td>
<td>0.0</td>
<td>1.4</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>D</td>
<td>−0.1</td>
<td>1.4</td>
<td>−1.7</td>
<td>4.5</td>
<td>1.6</td>
<td>3.0</td>
<td>0.0</td>
<td>2.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
The influence of the uncertainty in leaf area density on the model calculations was examined. As mentioned before, the leaf area density of winter wheat was determined using LAI from literature and the measured plant height. Dividing LAI (about 5–6) by the plant height (0.7 m), the leaf area density has the value about 7–9 (m² m⁻³). The middle value 8.0 (m² m⁻³) was used for the model calculations and the uncertainty of this value was about 1 (m² m⁻³). Considering uncertainties in the plant height, the uncertainty of leaf area density was larger. Therefore, calculations using four different values of leaf area density (6.0, 7.0, 9.0, and 10.0 m² m⁻³) were carried out and compared with the calculation for the original setting. The AD and rmsd of the surface fluxes from the results by the original setting are summarized in Table 3. The sensible heat flux decreases and the latent heat flux increases almost linearly as the leaf area density increases and these changes cancel out. For the net radiation and the ground heat conduction, a similar tendency as the sensible heat flux is seen but changes are small. Although biases are small, rmsd of the ground heat conduction are the same level as those of the sensible and latent heat fluxes. The small rmsd of the net radiation indicates that the change in leaf area density mainly affects the distribution of the net radiation to the other three heat fluxes and has a small effect on the net radiation. Although these differences are much larger than those caused by the uncertainty in the initial soil moisture condition, they are acceptable if the uncertainty of leaf area density is less than 1 (m² m⁻³).

e. Comparison with observations

Comparisons between observation and calculation are shown in Fig. 7 for time series of net radiation, sensible and latent heat fluxes, and ground heat conduction. The model simulated the net radiation and the ground heat conduction almost perfectly. Although some overestimation and underestimation are seen in the sensible and latent heat fluxes, the model simulated diurnal variations and peak value changes in a several-day cycle satisfactorily. Correlations between observation and calculation for the surface fluxes are shown in Fig. 8. Although the calculations are well correlated with the observations, the overestimation of sensible heat flux and the underestimation of latent heat flux are shown. As described in the sensitivity analysis, the sensible and latent heat fluxes calculated by the model decrease and increase, respectively, if the initial soil moisture at deep layer or the leaf area density increases. Some portion of the overestimation of sensible heat flux and the underestimation of latent heat flux can be reduced with this. Therefore,
TABLE 3. AD and rmsd of RN, SH, LH, and GH between calculation with the original setting of leaf area density, 8.0 (m$^2$ m$^{-3}$) and those using other settings of leaf area density; A is 6.0, B is 7.0, C is 9.0, and D is 10.0 (m$^2$ m$^{-3}$).

<table>
<thead>
<tr>
<th>Setting</th>
<th>RN (W m$^{-2}$) AD</th>
<th>Rmsd</th>
<th>SH (W m$^{-2}$) AD</th>
<th>Rmsd</th>
<th>LH (W m$^{-2}$) AD</th>
<th>Rmsd</th>
<th>GH (W m$^{-2}$) AD</th>
<th>Rmsd</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0.5</td>
<td>3.7</td>
<td>3.0</td>
<td>7.8</td>
<td>-2.9</td>
<td>9.2</td>
<td>0.4</td>
<td>9.4</td>
</tr>
<tr>
<td>B</td>
<td>0.2</td>
<td>2.1</td>
<td>1.5</td>
<td>4.9</td>
<td>-1.5</td>
<td>4.8</td>
<td>0.2</td>
<td>4.7</td>
</tr>
<tr>
<td>C</td>
<td>-0.5</td>
<td>2.1</td>
<td>-1.8</td>
<td>4.6</td>
<td>1.4</td>
<td>4.9</td>
<td>-0.1</td>
<td>4.0</td>
</tr>
<tr>
<td>D</td>
<td>-0.5</td>
<td>2.9</td>
<td>-3.3</td>
<td>6.4</td>
<td>3.2</td>
<td>8.3</td>
<td>-0.4</td>
<td>7.8</td>
</tr>
</tbody>
</table>

there is a possibility that the biases in sensible and latent heat fluxes have been caused by the incorrect model settings for the initial soil moisture or the leaf area density. However, this cannot be verified because of the uncertainty in observation as follows. The observed sensible and latent heat fluxes used in this comparison were corrected to satisfy the energy balance with the other fluxes as discussed before. In this correction both sensible and latent heat fluxes increased by preserving the measured Bowen ratio for most of daytime measurements. This correction was based on the assumption that Bowen ratio was the same within the fetch of the measuring point. Therefore, the corrected values of sensible and latent heat fluxes had the uncertainty depending on the spatial variation in Bowen ratio.

Further comparison between observations and calculations was carried out for the radiometric surface temperature. In the model simulation, the radiometric surface temperature was calculated from the upward longwave radiation flux at the vegetation top. The temporal change of calculation agrees well with the measurement as shown in Fig. 7. The correlation between

Fig. 7. Time series of calculations (solid lines) and observations (open circles) for RN, SH, LH, and GH at the station 7 of CASES-97. Observed SH and LH were corrected to satisfy the energy balance with RN and GH.
the observation and the calculation is also very high as shown in Fig. 10. Since the ground-surface temperature is determined by the surface heat budget, these good results in the comparison of radiometric surface temperature indicate that the model simulated the surface fluxes correctly. However, the model has a tendency to overestimate the radiometric surface temperature within the range about 18–25°C of observation (shaded range in Fig. 10). It is considered that this overestimation of surface temperature indicates the existence of biases in sensible and latent heat fluxes because the underestimation of latent heat flux causes the higher surface temperature and the increase of sensible heat flux. Although these biases were recognized, the model simulated the surface fluxes and the radiometric surface temperature satisfactorily as a whole.

4. Conclusions

The validation and sensitivity analysis of a new atmosphere–soil–vegetation model SOLVEG was presented. The model consists of multilayer submodels for the atmosphere, soil, and vegetation and radiation transmission schemes in the canopy layer. The performance test and sensitivity analysis of the model were carried out using observation data of CASES-97, especially the surface measurements at a winter wheat field.

The surface fluxes measured at the winter wheat field did not satisfy the energy balance. This energy imbalance was not explained by the canopy heat storage, energy consumption by photosynthesis, and horizontal advection of sensible and latent heat, but was considered to be caused by large turbulent structures, which were
The change in leaf area density mainly affected the dis-

 Yanazawa also provided helpful discussions. The study

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 APPENDIX A

 List of Symbols

 $a$ Leaf area density ($m^2 m^{-3}$)
 $a_w$ Absorptivity of liquid water in air
 $A_{s, A}, A_{w}$ Reflectivities of soil surface, leaf surface, and liquid water in air
 $c_D$ Drag coefficient of leaf
 $c_E, c_H$ Water vapor and heat exchange coefficients between leaf and canopy air
 $c_{EO}, c_{H0}, c_{A0}$ Water vapor, heat and momentum exchange coefficients at soil surface
 $C_p^S, C_S, C_w$ Volumetric heat capacities of bulk soil and water ($J m^{-3} K^{-1}$)
 $D$ Soil water diffusivity ($m^2 s^{-1}$)
 $D_w$ Molecular diffusivity for water vapor in air ($m^2 s^{-1}$)
 $e$ Turbulence kinetic energy ($m^2 s^{-2}$)
 $E_b$ Evaporation rate of soil water (kg $m^{-2} s^{-1}$)
 $E_{cap}$ Dissipation rates of fog water concerning capture by leaves (kg $m^{-2} s^{-1}$)
 $E_{col}$ Dissipation rates of fog water concerning capture by rain droplets (kg $m^{-3} s^{-1}$)
 $E_d$ Evaporation rate from liquid water on unit leaf area (kg $m^{-2} s^{-1}$)
 $E_f$ Evaporation rate from fog water in unit volume air (kg $m^{-3} s^{-1}$)
 $E_{int}$ Interception rate of precipitation by unit leaf area (kg $m^{-2} s^{-1}$)
 $E_r$ Evaporation rate from rain droplets in unit volume air (kg $m^{-3} s^{-1}$)
 $E_s$ Transpiration rate from unit leaf area (kg $m^{-2} s^{-1}$)
 $E_t$ Root uptake rate of water (kg $m^{-3} s^{-1}$)


Fig. 10. Comparison between calculation and observation for the radiometric surface temperature at the station 7 of CASES-97. The model has a tendency to overestimate the observation within the shaded range.
\[ E_y \] Liquid water flux in soil (kg m\(^{-2}\) s\(^{-1}\))

\( f \) Coriolis parameter (s\(^{-1}\))

\( F_f, F_p \) Projection coefficient of fog water moving in horizontal direction, interception of precipitation, and shielding of radiation

\( F_{ref} \) Ratio of forward scattering to total reflection on leaf

\( H_b \) Latent heat due to phase change of soil water (J m\(^{-3}\) s\(^{-1}\))

\( H_c \) Sensible heat exchange between unit leaf area and canopy air (J m\(^{-2}\) s\(^{-1}\))

\( H_e \) Latent heat due to evaporation of fog water (J m\(^{-3}\) s\(^{-1}\))

\( H_p \) Latent heat due to evaporation of rain water (J m\(^{-3}\) s\(^{-1}\))

\( H_r \) Heating rate of air due to absorption or emission of radiation by liquid water in unit volume air (J m\(^{-3}\) s\(^{-1}\))

\( K, K_s \) Hydraulic conductivity of soil and its value for saturated soil (m s\(^{-1}\))

\( K_r \) Thermal conductivity of soil (m\(^2\) s\(^{-1}\))

\( l \) Latent heat of vaporization (J kg\(^{-1}\))

\( L^d, L^\uparrow \) Downward and upward longwave radiation fluxes (W m\(^{-2}\))

\( P_d \) Drip from leaves (kg m\(^{-2}\) s\(^{-1}\))

\( P_v \) Vertical liquid water flux in canopy (kg m\(^{-2}\) s\(^{-1}\))

\( q_a, q_s \) Specific humidity of atmosphere and air in soil pore (kg kg\(^{-1}\))

\( r_a \) Resistance for water vapor exchange between leaf and the atmosphere (s m\(^{-1}\))

\( r_b, r_{s,t} \) Resistance for evaporation of soil water and leaf surface water (s m\(^{-1}\))

\( r_s, r_{s,\text{min}} \) Stomatal resistance and its minimum value (s m\(^{-1}\))

\( R_t \) Stagnated water at soil surface (kg m\(^{-2}\))

\( S^\downarrow, S^\uparrow \) Downward and upward solar radiation flux (W m\(^{-2}\))

\( T_a, T_r, T_p, T_e \) Temperatures of air, vegetation, precipitation, and soil (K)

\( u, v \) Horizontal wind speed components (m s\(^{-1}\))

\( u, v \) Geostrophic wind components (m s\(^{-1}\))

\( [u] \) Wind speed (m s\(^{-1}\))

\( w_d, w_d^\text{sat} \) Amount of water on unit leaf area and its maximum value (kg m\(^{-2}\))

\( w_s \) Leaf surface liquid water at which evaporation is saturated (kg m\(^{-2}\))

\( w_f \) Fog water content (kg kg\(^{-1}\))

\( z_s \) Solar zenith angle

\( e_s, e_v, e_w \) Emissivities of soil surface, leaf surface, and liquid water in air

\( \eta_\text{silt} \) Wilting volumetric soil water content (m\(^3\) m\(^{-3}\))

\( \theta \) Potential temperature (K)

\( \lambda \) Turbulence length scale (m)

\( \rho, \rho_r, \rho_s \) Densities of air, bulk soil, and water (kg m\(^{-3}\))

\( \sigma \) Stephan–Boltzmann constant (5.67 × 10\(^{-8}\) W m\(^{-2}\) K\(^{-4}\))

\( \tau_r, \tau_e \) Momentum exchange at soil surface (kg m\(^{-1}\) s\(^{-2}\))

\( \Psi, \Psi_r \) Soil water potential and its value for saturated soil (m)

**APPENDIX B**

**Detailed Description of the Model Equations**

a. **Atmosphere submodel**

The forcing terms for \( u, v, \theta, q_a, e, e_\lambda \), and \( w_f \) in (1) are expressed, respectively as

\[
F_u = f(v - u_y) - ac_{u u}[u]u, \quad \text{(B1)}
\]

\[
F_v = -f(u - u_y) - ac_{u u}[u]v, \quad \text{(B2)}
\]

\[
F_\theta = (H_r + ah + \theta_r + H_{pr})(\rho C_p), \quad \text{(B3)}
\]

\[
F_q = \left[ a(E_d + E_s) + E_r + E_{pr} \right]/\rho, \quad \text{(B4)}
\]

\[
F_r = P_{cb} + P_{ca} - P_{ca} - P_{cb}, \quad \text{(B5)}
\]

\[
F_w = -[E_d + E_{col} + aE_{ap}]/\rho, \quad \text{(B7)}
\]

where \( H_r = -\Delta E_d \) and \( H_{pr} = -\Delta E_{pr} \). The terms on the right-hand side of (B5) and (B6) are shear production, buoyancy production, dissipation by viscosity, and production by leaf surface; expressions by Yamada (1982) are also used. The terms on the right-hand side of (B7) are the production or dissipation of fog water by phase change, accretion with rain droplets, and capture by leaf surface, respectively. It is assumed that the evaporation and condensation of fog water take place within an infinitely small period \( \delta t \) according to equations

\[
E_r \delta t = \rho \min[q_{sat}(T_m) - q_a, w_f] = -C_r \rho(T_m - T_e)l, \quad \text{(B8)}
\]

\[
H_f \delta t = \rho C_p(T_m - T_a), \quad \text{(B9)}
\]

where \( q_{sat}(T) \) represents saturated specific humidity for temperature \( T \) and \( T_m \) is the temperature after the phase change.

b. **Soil submodel**

The boundary condition for (2) at the soil surface is a heat budget equation expressed as

\[
(1 - A_b)S^\downarrow + e_s(L^\downarrow - \sigma T_m^4) = H_r + G + H_{sat}, \quad \text{(B10)}
\]

where the subscript 0 refers to the variable at the soil
surface. This equation has no latent heat term, which is usually included on the right-hand side of the ground surface heat budget equation for bare soil. This is because the model is designed to deal with the water vapor transport in soil explicitly and to calculate the latent heat exchange caused by evaporation of soil water at each layer of soil as volume source or sink term expressed by the second term on the right-hand side of (2). The terms on the right-hand side of (B10) are expressed as

\[ H_b = C_r \rho_c c_{f0} \left| \mathbf{u} \right| (T_{so} - T_r), \]  

\[ G_0 = C_r \rho_c K_r \frac{\partial T}{\partial z} \bigg|_{z=9}, \]  

\[ H_{so} = C_r \rho_c P_v (T_{so} - T_r), \]  

where \( \left| \mathbf{u} \right| \) and \( T_r \) are wind speed and temperature at a reference height of the atmosphere, respectively. The cooling or warming of soil caused by precipitation is also at any surface of soil particles. This means that the condensation of water vapor in soil pores takes place within an infinitesimal period \( \delta t \) according to the equations

\[ E_a \delta t = \rho (\eta_{ss} - \eta_s) (q_{so} - q_s). \]  

\[ H_g \delta t = C_r \rho_c (T_{so} - T_g). \]  

where \( H_g = \rho c_{f0} T \) and \( T_g \) is the soil temperature after the condensation of water vapor in soil pores. Parameter \( H_g \delta t \) is the latent heat released by condensation and the temperature rise of the soil including the air of soil pores.

If the precipitation is strong enough to saturate the surface soil, the water in excess stagnates at soil surface. This stagnated water \( R_s \) is calculated by

\[ \frac{dR_s}{dt} = P_{s0} + E_{w0}, \]  

where \( E_{w0} \) is the direct evaporation from the thin surface layer to the atmosphere, obtained by integrating \( I_{so} \) in the layer (0 to \( \delta z_s \)). Parameter \( E_o \) is water vapor flux from soil surface to the atmosphere expressed by

\[ E_o = \rho c_{f0} \left| \mathbf{u} \right| (q_{so} - q_s). \]  

The sensible and latent heat exchanges between soil surface and the atmosphere are expressed by (B11) and (B21), respectively. The momentum exchange \( \tau = (\tau_r, \tau_z) \) is calculated by

\[ (\tau_r, \tau_z) = \rho c_{f0} \left| \mathbf{u} \right| (u_r, v_z), \]  

c. Vegetation submodel

The terms in (5) are expressed as

\[ R_s(z) = F_s [(1 - A_s) \sigma \left( T_s^4 + T^4 \right)] + e_s \left[ L_s^4 + L_s^4 - 2 a \sigma \left( T_s^4 \right) \right]. \]  

\[ E_s(z) = \rho c_p \left( T(z) - T_s \right), \]  

\[ H_s(z) = C_s F_s P_s (T(z) - T_s). \]  

A coefficient \( F_s \) accounts for that effective leaf area whose shielding radiation flux is smaller than that of
the total leaf area because of the inclination of leaf surface to radiation and the overlap of leaves. \( F_p \) is assumed to have the same value as \( F_r \). Based on the shielding factor of leaves by Kanemasu et al. (1976), the projection coefficient is calculated by

\[
F_r = \frac{1 - \exp(-0.4\Delta A)}{\Delta A}, \quad (B27)
\]

where \( \Delta A \) is an accumulated leaf area per unit area in the canopy layer with thickness of \( \Delta z_c \). The water vapor flux from leaf surface to the atmosphere \( E_c \) is divided into two components; transpiration \( E_t \) and evaporation from leaf surface water \( E_d \), expressed as

\[
E_t = \rho R_{d_s} [q_{sl}(T_c) - q_a], \quad \text{and} \quad (B28)
\]

\[
E_d = \rho R_{r_s} [q_{sl}(T_c) - q_a], \quad (B29)
\]

respectively, where \( R = (r_s r_p + r_s r_d + r_p r_d)^{-1} \). The model uses the formulations for the stomatal resistance \( r_s \) and the resistance for evaporation of leaf surface water \( r_d \) proposed by Deardorff (1978). Parameter \( r_s \) is assumed to depend only on the aerodynamical characteristic of leaf surface, and expressed as \( r_s = (c_{rs} |\mathbf{u}|)^{-1} \). Since sensible heat and water vapor are passive scalars, the same value can be used for exchange coefficients \( c_{rp} \) and \( c_{rc} \). The water vapor fluxes described above are positive for the direction toward the atmosphere, and may have a positive or negative value.

The terms on the right-hand side of (6) are expressed as

\[
E_{sat} = F_p P_r, \quad (B30)
\]

\[
E_{cap} = F_p \left| \mathbf{u} \right| \rho w_t, \quad \text{and} \quad (B31)
\]

\[
P_d = \begin{cases} 0 & w_p < w_{ds} \\ E_{sat} - E_d + E_{cap} & w_p = w_{ds} \end{cases}, \quad (B32)
\]

The amount of fog water captured by unit leaf area \( E_{cap} \) is proportional to speed of fog water (wind speed \( |\mathbf{u}| \)), and weight of fog water in unit weight air. When the amount of water on unit area of leaf surface exceeds the threshold value \( w_{ds} \), the excess becomes the drip from leaf surface \( P_d \). The value \( w_{ds} \) would naturally depend on shape, surface characteristic, angle, and motion (wind speed) of leaves. However, the model uses a constant value for \( w_{ds} \).

The terms on the right-hand side of (7) are expressed as

\[
E_{pr} = \frac{3\rho e_p}{4\rho w} [q_{sl}(T_p) - q_a], \quad \text{and} \quad (B33)
\]

\[
E_{cas} = \frac{3\rho P_w w_t}{4\rho w}. \quad (B34)
\]

These terms are considered to be proportional to the speed of a raindrop relative to the air, and the density of a raindrop cross section, and obtained using the assumptions that the specific humidity at the surface of a raindrop has the saturated value for the temperature of the raindrop, that every raindrop has a spherical shape with the same radius \( r \), and that raindrops capture all the fog water in their paths.

**REFERENCES**


