Description and validation of the atmosphere–land–surface interaction scheme (ALSIS) with HAPEX and Cabauw data

Parviz Irannejad, Yaping Shao

Centre for Advanced Numerical Computation in Engineering and Science, The University of New South Wales, Sydney, NSW, Australia

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Abstract

A new land surface parameterization scheme (ALSIS), with emphasis on soil moisture prediction, is described and validated with observations from HAPEX-MOBILHY and Cabauw. An important feature of the scheme is the inclusion of vertical heterogeneity of soil hydraulic parameters in modelling unsaturated flow. The simulated soil moisture for HAPEX site using a vertically homogeneous soil has a positive bias in the upper soil layers and a negative bias in the deep soil layers. Taking into account the soil vertical heterogeneity greatly eliminates this discrepancy and results in an excellent agreement between annual cycles of modelled and observed soil moisture profiles. The mean annual soil moisture in the top 1.6 m of soil increased from 394 mm for homogeneous case to 433 mm for the heterogeneous case, consistent with 435 mm observed. The improvement in soil moisture simulation resulted in an improved skill in predicting the mean and the diurnal cycles of surface fluxes for the intensive observational period (28 May–3 July). The simulated monthly averages of surface temperature and fluxes follow observations over the year, except for January when the model overestimates the latent heat flux due to its failure in simulating high rates of dew fall. The deviation of modelled monthly mean surface fluxes from observations are well within the estimated observational errors. The simulated mean daily surface temperature, and surface fluxes are generally consistent with observations, except for some times in the winter period. The modelled diurnal cycles of temperature and fluxes are in agreement with those observed. However, the model overestimates the night-time latent heat flux, especially during January.

1. Introduction

The land surface component of an atmospheric model should be able to realistically represent the partitioning of the surface available radiative energy between latent and sensible heat fluxes. Studies show that this partitioning has an important influence on the global climate (e.g., Shukla and Mintz, 1982; Yeh et al., 1984) and on the mesoscale atmospheric circulation and precipitation (Avissar and Pielke, 1989; Blyth et al., 1994). The partitioning of the surface energy is closely linked to surface water balance through high dependency of surface evapotranspiration on the availability of soil moisture. Therefore, to provide a good approximation of
surface energy fluxes, the land surface scheme must adequately compute soil moisture, and other components of the soil water balance such as runoff and drainage.

In the early general circulation models (GCMs), the land surface is typically modelled as a simple bucket (Manabe, 1969) that can be filled by precipitation and emptied by evaporation. Vegetation is not modelled explicitly and there is no distinction between evaporation and transpiration. When the bucket is full the excess precipitation leaves the grid area in the form of instant runoff/drainage. Because of its simplicity and computational efficiency, this type of land surface representation is still used in long-term climate studies (e.g., Delworth and Manabe, 1989; Manabe and Stouffer, 1996).

Studies showed that, due to the direct contribution of the bulk soil moisture, the bucket model overestimates evaporation, does not respond rapidly to surface wetness following precipitation, and dries out the soil earlier and to a greater extent than observations (see Shao et al., 1994). Recently, many advanced land surface schemes with explicit parameterization of the effects of vegetation on surface fluxes and with multiple soil layers have been developed (e.g., Dickinson et al., 1986; Sellers et al., 1986; Noilhan and Planton, 1989; Pitman et al., 1991). These schemes have improved the simulation of land surface fluxes (Sellers, 1992) and their diurnal cycle (Dickinson and Henderson-Sellers, 1988).

Robock et al. (1994) showed that both the bucket model and Simplified SiB (SSiB) model (Xue et al., 1991) have deficiencies in simulating soil moisture. The comparison of soil moisture simulation by 14 land–surface schemes for the Hydrologic Atmospheric Pilot Experiment and Modelisation du Bilan Hydrique (HAPEX-MOBILHY) site in southern France (Shao et al., 1994) showed a range of about 100 mm for a 1.6 m deep soil at the end of the equilibrium year. The differences were larger during the growing season, when almost all of the schemes underestimated soil moisture. Irannejad et al. (1995) showed that the inconsistencies in soil moisture simulation persisted when the schemes were coupled into their host atmospheric models. It was concluded that the calculation of soil moisture was quite unreliable and that there was a need for improved soil moisture treatment in the land surface parameterization schemes.

The atmosphere–land–surface interaction scheme (ALSIS) presented here is a land–surface parameterization scheme with an emphasis on simulating the soil moisture. The one-dimensional Richards’ Equation (Richards, 1931) is solved to predict the soil moisture evolution. ALSIS considers soil as a vertically heterogeneous media for calculating water and heat transfer. The model explicitly accounts for the effects of vegetation on surface roughness, surface fluxes and water uptake from soil. The subgrid scale heterogeneity of the land surface is modelled following the mosaic approach of Koster and Suarez (1992). The scheme, developed as a component of a land use sustainability information system, is currently running interactively in the University of New South Wales high resolution limited area weather prediction model (Leslie and Purser, 1991) over the Australian continent in horizontal resolution of 50 × 50 km and 10 × 10 km. The surface and soil parameters required are compiled in a 0.05° × 0.05° resolution, based on the Atlas of Australian Soils (NATMAP, 1980) and Atlas of Australian Vegetation (AUSLIG, 1990).

A brief description of the scheme is presented in Section 2. Section 3 represents off-line tests of the model for HAPEX-MOBILHY site (Goutorbe and Tarrieu, 1991) and Cabauw site (Monna and Van der Vliet, 1987).

2. Model description

2.1. Soil water

The one-dimensional vertical flow of water in a macroscopically uniform, unsaturated soil is described by the Darcy–Buckingham expression:

\[ q = K \left( 1 - \frac{\partial \psi}{\partial z} \right) \]

where \( \psi \) is the matrix potential of the soil water, \( K (= K (\psi)) \) is soil hydraulic conductivity, and \( z \) is soil depth.
Eq. (1) is sufficient only to describe the soil moisture profile under steady flow. For transient flow, in which the magnitude and possibly the direction of flux and potential gradient vary with time, the introduction of the law of conservation of water, expressed in the equation of continuity, is required. For the one-dimensional vertical flow the equation of conservation of water is:

$$\frac{\partial \theta}{\partial t} = -\frac{\partial q}{\partial z} - S$$

(2)

where \( \theta \) is volumetric soil water content, \( t \) is time, and \( S \) is the sink term for water due to evapotranspiration and horizontal discharge. Substituting Eq. (1) in Eq. (2) provides the Richards’ Equation:

$$\frac{\partial \theta}{\partial t} = -\frac{\partial}{\partial z} \left( K - K \frac{\partial \Psi}{\partial z} \right) - S$$

(3)

The Richard’s Eq. (3) is highly nonlinear, since \( K \) and \( \theta \) are normally highly nonlinear functions of \( \Psi \).

Other forms of the flow equation have been investigated with a view to dealing with its nonlinearity. Haverkamp et al. (1977) re-formulated the Richards’ Equation with Kirchhoff transform, \( U \), as the sole dependent variable. Redinger et al. (1984) and Campbell (1985) applied the Kirchhoff transform to the diffusion term of the Richards’ Equation, so that it becomes:

$$\frac{\partial \theta}{\partial t} = -\frac{\partial}{\partial z} \left( K \frac{\partial U}{\partial z} \right) - S$$

(4)

with:

$$U = \int_{-\infty}^{\Psi} K d\Psi = \int_{0}^{\theta} D d\theta$$

where \( D = K(d\Psi)/(d\theta) \) is soil hydraulic diffusivity. ALSIS uses Eq. (4) for predicting soil moisture. The Kirchhoff transform reduces the high gradients in \( \Psi \) of dry soils to lower gradients in the transformed variable, so that without losing accuracy a coarser node spacing can be used. The other advantage of this transform is that it formally eliminates conductivity from the diffusion term.

The common assumption in land surface schemes is that soil is vertically homogeneous. While uniform soils are a useful start in developing land surface models, they are the exception in nature, rather than the norm. The assumption of uniform soil does not especially hold over the Australian continent, where duplex soils are the most common, only after sandy soils (NATMAP, 1980). Talsma and Flint (1958) report that even in uniform soils hydraulic conductivity decreases with depth. Irannejad and Shao (1996) found improved skill in simulating soil moisture and surface–atmosphere energy and moisture exchanges using a two-horizon soil. ALSIS accounts for the heterogeneity of soil hydraulic properties in the vertical direction.

Campbell (1985) states that the Kirchhoff transform is not applicable for simulating water flow in the layered soils. However, following Ross (1990) and Ross and Bristow (1990) in making appropriate correction due to the change in soil properties with depth, we found no difficulty in using the Kirchhoff transform to solve the Richards’ Equation in layered soils.

Solution of the Richards’ Equation requires closure relationships between hydraulic conductivity, soil water content, and the matrix potential of soil water. Several investigators have proposed models for these relationships. The common model being used in land–surface schemes for atmospheric models is the Brooks–Corey model (Brooks and Corey, 1966; hereafter BC), a simple power law relationship, in its original form (Liang et al., 1994) or its Clapp and Hornberger (1978); hereafter CH modification (Noilhan and Planton, 1989; Dickinson et al., 1993). The Brooks–Corey water retention, \( \Psi(\theta) \), and hydraulic conductivity, \( K(\theta) \), functions are:

$$\Psi(\theta) = \Psi_{r} \theta^{-b}$$

(5a)

$$K(\theta) = K_{r} \theta^{2b+3}$$

(5b)
where $\Psi_s$ is the value of soil water potential at saturation, $K_s$ is saturation hydraulic conductivity, and $b$ is a parameter related to the pore size distribution. $\Theta$ is volumetric water content scaled by its saturation value. The original Brooks–Corey model includes a residual saturation, $\theta_r$, equated to air dry water content, so that:

$$\Theta = \frac{\theta - \theta_r}{\theta_s - \theta_r} \quad (6a)$$

In the Clapp–Hornberger modification the residual term is zero, thus:

$$\Theta = \frac{\theta}{\theta_s} \quad (6b)$$

The problem with the power functions (Eqs. (5a) and (5b)) is that they break down when soil moisture approaches saturation, making them inappropriate for infiltration representation (Wetzel et al., 1996). To avoid this Clapp and Hornberger (1978) modified Eqs. (5a) and (5b) by substituting the power function with a parabolic equation when soil moisture is near saturation.

van Genuchten (1980); hereafter VG presented a model for soil hydraulic functions that is being widely used in the soil physics and hydrology communities. The soil water retention curve proposed by van Genuchten is:

$$\Theta = \left(\frac{1}{1 + (\alpha \Psi)^n}\right)^m \quad (7a)$$

with:

$$m = 1 - \frac{1}{n}$$

where $\alpha$, $n$, and $m$ are model parameters. Using (Eq. (7a)) together with the Mualem (1976) hypothesis, van Genuchten derived the following equation for hydraulic conductivity:

$$K(\theta) = K_s \Theta^2 \left[1 - \left(1 - \Theta^m\right)^{\frac{1}{m}}\right] \quad (7b)$$

By analytically solving the Richards’ Equation, Broadbridge and White (1988; hereafter BW) give a simple functional form for hydraulic conductivity as follows:

$$K(\theta) = K_f + \Delta K \Theta^2 \frac{C - 1}{C - \Theta} \quad (8a)$$

Assuming $K_f \approx 0$, the moisture characteristics, $\Psi(\theta)$, is:

$$\Psi(\theta) = -\lambda_c \left[\frac{1 - \Theta}{\Theta} + C^{-1} \ln \frac{C - \Theta}{\Theta(C - 1)}\right] \quad (8b)$$

where $\lambda_c$ is the capillary length scale and $C$ is a free parameter dependent on the soil pore distribution. In the Broadbridge–White model soil water diffusivity remains finite as the soil becomes either very dry or saturated. Furthermore, hydraulic functions may be scaled across all soils described by the model. These permit guarantee a priori the numerical performance of the finite difference solutions of Richards’ Equation (Short et al., 1993). We may loosely name $b$ in BC and CH, $m$ in VG, and $C$ in BW the shape parameter, and $\Psi_s$ in BC and CH, $\alpha$ in VG, and $\lambda_c$ in BW as scaling parameter.

ALSIS is flexible to use any of models described above. The impact of applying different soil models on soil moisture and the performance of land surface models in general has been discussed in Shao and Irannejad (1997).
2.2. Surface Radiation

The analysis of different components of the surface radiation budget begins with the assumption that solar radiation reaching the soil surface under the canopy ($R_{sc}$) can be calculated using a Beer–Lambert Law relationship:

$$R_{sc} = R_{f} f_{ext}$$  \(\text{(9)}\)

with $f_{ext}$, the radiation fraction incident on the soil surface beneath the canopy, as:

$$f_{ext} = e^{-\kappa \text{LAI}}$$

where $R_{f}$ is shortwave radiation intensity at the top of the canopy, LAI is the integrated leaf area index and $\kappa$ is the extinction coefficient of the vegetation. The radiation absorbed by the canopy is:

$$R_s = R_f (1 - f_{ext}) (1 - \alpha_c)$$  \(\text{(10)}\)

where $\alpha_c$ is vegetation albedo. The absorption of the solar radiation by the soil surface can be written as:

$$R_{ss} = \sigma_{l} \alpha_{s} R_{sc} + (1 - \sigma_{l}) R_{sub}$$  \(\text{(11)}\)

with $R_{sub}$, radiation rate absorbed at the uncovered soil surface as:

$$R_{sub} = R_s (1 - \alpha_s)$$  \(\text{(12)}\)

where $\sigma_l$ is the fraction of the land covered by the canopy and $\alpha_s$ is the soil albedo. Substituting Eqs. (9) and (12) in Eq. (11) gives:

$$R_{ss} = R_s (1 - \alpha_s) \left[1 - \sigma_{l} (1 - f_{ext})\right]$$  \(\text{(13)}\)

Similarly, the sky longwave radiation absorbed by the canopy and by the soil are

$$R_{sc} = \varepsilon_c R_{l} \sigma_{l} (1 - f_{ext})$$  \(\text{(14-1)}\)

$$R_{ss} = \varepsilon_s R_{l} \left[1 - \sigma_{l} (1 - f_{ext})\right]$$  \(\text{(14-2)}\)

where $\varepsilon_c$ and $\varepsilon_s$ are the absorptivities (emissivities) of the vegetation and soil for longwave radiation. The longwave radiation emitted by the surface is calculated using the Stefan–Boltzmann equation:

$$R_{se} = \sigma_l (1 - f_{ext}) \varepsilon_c \sigma T_c^4$$  \(\text{(15-1)}\)

$$R_{ss} = \varepsilon_s \sigma T_s^4$$  \(\text{(15-2)}\)

where $T_c$ and $T_s$ are the canopy and the soil surface temperatures and $\sigma$ is the Stefan–Boltzmann constant ($= 5.67 \times 10^{-8}$ W m$^{-2}$ K$^{-4}$).

The net radiation at the canopy and at the soil surface are calculated as:

$$R_{nc} = R_{sc} + R_{lc} + R_{usu} \sigma_{l} (1 - f_{ext}) - 2 R_{sec}$$  \(\text{(16-1)}\)

and:

$$R_{ns} = R_{ss} + R_{ls} + R_{usc} - R_{sub}$$  \(\text{(16-2)}\)

The land surface net radiation is calculated as the sum of the net radiation at the soil and canopy surfaces:

$$R_n = R_{ns} + R_{nc}$$  \(\text{(17)}\)

2.3. Soil and canopy temperatures

Temperature is calculated separately for the soil and the canopy. Heat can transfer in soils by conduction, by the flow of liquid water, by diffusion of vapour, and by convection. Compared to the two former processes, the
latter two are usually negligible. For dry soils, where water flux is very small, the only significant process for heat flow is conduction. Kirkham and Powers (1972) state that the theory based on heat conduction provides comparable results with observations, although it is not physically plausible. In ALSIS, the vertical flux of heat through the unit cross-sectional area of soil is described by the Fourier law of heat conduction in a multi-layer soil:

\[ G_h = -K_h \frac{\partial T}{\partial z} \]  

(18)

where \( K_h \) is the soil thermal conductivity and \( T \) is the soil temperature. The equation for conservation of energy in soil, neglecting horizontal heat transfer and the heat released/extracted by phase change of water, is

\[ \frac{\partial T}{\partial t} = - \frac{1}{C_h} \frac{\partial G_h}{\partial z} \]  

(19)

with \( C_h \), the volumetric heat capacity, given as:

\[ C_h = \rho_s c_s \]

where \( \rho_s \) and \( c_s \) are soil density and specific heat capacity, respectively. \( \rho_s, c_s \) and \( K_h \) are dependent on the soil porosity (\( \nu \)) and volumetric water content (\( \theta \)). Neglecting the density and heat capacity of the soil air, we obtain:

\[ \rho_s c_s = (1 - \nu) \rho_q c_q + \theta \rho_w c_w \]  

(20-1)

The soil thermal conductivity is calculated following Kowalczyk et al. (1991) as:

\[ K_h = 419(\alpha \theta + b \theta^{0.4}) \]  

(20-2)

where \( \rho_q, c_q, \rho_w, c_w \) are density of water, specific heat of water, density of quartz, and specific heat of quartz, respectively. Constants \( a \) and \( b \) are dependent on the soil type. Substitution of Eq. (18) in Eq. (19) gives the prognostic equation for the soil temperature:

\[ \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} D_h \frac{\partial T}{\partial z} \]  

(21)

where \( D_h = (K_h/C_h) \) is soil thermal diffusivity.

The soil surface temperature is calculated diagnostically by iteratively solving the surface energy balance equation:

\[ R_w(T_s) - \lambda E_c(T_s) - H_c(T_s) - G_h(T_s) = 0 \]  

(22)

where \( G_h(T_s) \) is the heat flux into the soil at the soil surface.

It is assumed that the canopy has no heat capacity. Neglecting the energy used in the photosynthesis and respiration processes, the energy balance equation of the canopy is solved iteratively for canopy temperature, \( T_c \):

\[ R_w(T_c) - \lambda E_c(T_c) - H_c(T_c) = 0 \]  

(23)

where \( E_c \) is the canopy evapotranspiration rate, \( \lambda \) is the latent heat of vaporisation and \( H_c \) is the sensible heat flux from the canopy.

2.4. Prognostic equations for canopy and soil water stores

The prognostic equation for the canopy water store is:

\[ \frac{\partial W_c}{\partial t} = P_c - D_c - E_{vc} \]  

(24)
where $P_s = (\sigma, P_c)$ is the precipitation rate on the canopy, $D_c$ is the rate of water drainage from the canopy and $E_{ev}$ is evaporation rate (negative if condensation on the canopy) from the wet portion of the canopy surface, and $W_c$ is the canopy water content.

The prognostic equation for volumetric water content in different soil layers is:

$$\frac{\partial \theta_i}{\partial t} = \frac{\Delta q_i - f_{ei} E_v - f_{ei} T_i}{\Delta z_i}$$

(25)

where $\Delta q_i = q_i - q_{i-1}$ and $q_0$ and $q_{i-1}$ are the water flux at the upper and lower boundaries of layer $i$, $E_v$ is evaporation from the soil surface, $T_i$ is transpiration rate and $\Delta z_i$ is the thickness of the soil layer. $f_{ei}$ and $f_{ei}$ are the contribution of each soil layer in the total soil evaporation and in the total transpiration, respectively:

$$f_{ei} = \begin{cases} 1, & i = 1 \\ 0, & i > 1 \end{cases}$$

(26a)

and:

$$f_{ei} = \frac{\theta_i - \theta_{wilt}}{\theta_{fc} - \theta_{wilt}} \Delta r_j$$

(26b)

where $\Delta r_j$ is the fraction of roots and $\theta_i$ is the volumetric water content. $\theta_{fc}$ and $\theta_{wilt}$ are volumetric water content at field capacity and permanent wilting point, equated to the water content at the matrix potential of $-1$ m and $-150$ m of water, respectively.

### 2.5. Root distribution

Determining root distribution is important in modelling the uptake of water by plant roots. The greatest concentration of roots generally occurs in the surface soil (Kalisz et al., 1987), where organic carbon content and cation exchange capacity are higher and soil bulk density is lower (Sainju and Good, 1993). To obtain the cumulative distribution of roots usually a nonlinear relationship between the root fraction and the soil depth is used. Gerwitz and Page (1974) proposed the following exponential relationship:

$$R(z) = e^{-fz}$$

(27)

where $f$ is the reciprocal of the depth above which 63% of roots exists. Abramopoulos et al. (1988) used a power function to describe cumulative root distribution with variable coefficients for different vegetation types:

$$R(z) = az^b$$

(28)

The exponential relationship suggested by Verseghy et al. (1993) is:

$$R(z) = a_1 e^{-a_2 z} + a_3$$

(29)

where $R(z)$ is cumulative fraction of root system below depth $z$ and $a_1$, $a_2$ and $a_3$ are constants.

Following Verseghy et al. (1993), ALSIS uses Eq. (29) to estimate plant root distribution. The root distribution for the perennial vegetation is constant. For the annual vegetation we calculate root distribution as a function of time between germination (a function of deep soil temperature) and maturity.
2.6. Drag coefficients

The model applies Monin–Obukhov similarity theory to calculate the bulk transfer coefficients:

\[
C_{Dm} = \frac{\kappa^2}{\left( \ln \frac{z_t - z_d}{z_{0m}} - \Psi_1 \right)^2}
\]

\[
C_{Dh} = \frac{\kappa^2}{\left( \ln \frac{z_t - z_d}{z_{0h}} - \Psi_2 \right)^2}
\]

where \(C_{Dm}\) and \(C_{Dh}\) are bulk transfer coefficients for momentum and heat, \(z_{0m}\) and \(z_{0h}\) are surface roughness lengths and \(\Psi_1\) and \(\Psi_2\) are the stability corrections for momentum and for heat, respectively, \(\kappa (= 0.4)\) is von Karman constant, \(z_t\) is the reference height and \(z_d\) is zero-plane displacement. The bulk transfer coefficient for water vapour is taken equal to that for heat \((C_{Dw} = C_{Dh})\).

Stability corrections are calculated following Paulson (1970):

\[
\Psi_1 = \begin{cases} 
5 \frac{z_t}{L} & \frac{z_t}{L} > 0 \\
\ln \left( \frac{(1 + x_1)^2}{8} \right) - 2 \tan^{-1}(x_1) + \frac{\pi}{2} & \frac{z_t}{L} \leq 0
\end{cases}
\]

\[
\Psi_2 = \begin{cases} 
-5 \frac{z_t}{L} & \frac{z_t}{L} > 0 \\
2 \ln \frac{1 + x_1^2}{2} & \frac{z_t}{L} \leq 0
\end{cases}
\]

with:

\[x_1 = \left( 1 - 16 \frac{z_t}{L} \right) / 2\]

and:

\[L = - \frac{u_s^3 c_p \rho_a T_a}{\kappa g H}\]

where \(L\) is the Monin–Obukhov length and \(u_s = \left( \kappa U_s / \left( \ln \left( z_t - z_d / z_{0m} \right) - \Psi_1 \right) \right)\) is the friction velocity, \(g\) is acceleration due to gravity and \(H\) is sensible heat flux. Roughness length and zero-plane displacement of vegetation are estimated based on Raupach (1994):

\[
z_d = h_c \frac{1 - e^{-x^2}}{x_2}
\]

\[z_{0m} = h_c \left[ \frac{1 - e^{-x_2}}{x_2} e^{(\ln c_s^{-1} + \frac{\kappa}{c_s + c_s A})} - \frac{\kappa}{c_s + c_s A} \right]
\]
with:
\[ x = \sqrt{C_s A} \]

where \( A \) is the canopy (leaf + stem) area index, \( C_s \) and \( C_l \) are drag coefficients of an isolated roughness element and of the substrate surface at canopy height in the absence of roughness elements, respectively, and \( C_d \) is a free parameter. \( h_c \) is the height of the roughness elements (canopy) and \( C_w = (z_w - z_d)/(h_c - z_d) \) with \( z_w \) the upper height limit.

Studies show (e.g., Sugita and Brutsaert, 1991; Garratt, 1992; Betts and Beljaars, 1993) that roughness length for heat is much smaller than that for momentum. Following Garratt (1992) we take:

\[ z_{0h} = \frac{1}{7} z_{0m} \]

2.7. Soil evaporation

Evaporation is calculated using the bulk transfer formulation:

\[ E_v = \frac{q_v - q_a}{r_n} \]

where \( \rho_a \) is the air density, \( q_v \) is the specific humidity of the air at the reference level \((z_r)\), \( q_a \) is the specific humidity at the roughness height for moisture \((z_{0h})\) and \( r_n \) is the aerodynamic resistance to moisture flux from

![Fig. 1. Comparison of simulated annual cycle of total soil water in the top 0.5 m (a) and 1.6 m (b) of homogeneous soil, using Clapp–Hornberger (CH), Brooks–Corey (BC), Broadbridge–White (BW) and van Genuchten (VG) soil hydraulic models, with observations from HAPEX-MOBILHY.](image-url)
\[ q_s = \alpha q_{sat}(T_s) \]  
\[ q_s = \beta q_{sat}(T_s) + (1 - \beta) q_a \]  

where \( q_{sat}(T_s) \) is saturated specific humidity at the soil surface, and \( \alpha \) and \( \beta \) are coefficients related to surface moisture availability. \( \alpha \) shows the relative humidity at the soil surface, but is usually substituted by the relative humidity of the air adjacent to the free water in the soil. Substituting Eqs. (34a) and (34b) in Eq. (33) gives:

\[ E_v = \rho \frac{\alpha q_{sat}(T_s) - q_a}{r_a} \]  
\[ E_v = \rho \beta q_{sat}(T_s) - q_a \]  

Eq. (35a) calculates moisture transfer from \( z_{q0} \) to \( z_r \), while Eq. (35b) gives the moisture transfer directly from the surface of the free water in the soil to \( z_r \). A comparison of different \( \alpha \)-type and \( \beta \)-type methods for

![Fig. 2](https://example.com/fig2.png)

Fig. 2. Same as Fig. 1, but for volumetric water content in five soil layers. The depths of the soil layers are (a) 0.0–0.05 m, (b) 0.05–0.20 m, (c) 0.20–0.50 m, (d) 0.50–1.00 m, and (e) 1.00–1.60 m.
calculating soil evaporation along with the supply-demand approaches (see e.g., Abramopoulos et al., 1988; Dickinson et al., 1993) is presented in Mahfouf and Noilhan (1991). ALSIS uses the $\beta$ formulation for calculating soil evaporation. The parameterization proposed by Lee and Pielke (1992) is used to calculate $\beta$:

$$\beta = \begin{cases} \frac{1}{4} \left( 1 - \cos \left( \frac{\theta_1}{\theta_{fc}} \right) \right)^2, & \theta_1 < \theta_{fc} \\ 1, & \theta_1 \geq \theta_{fc} \end{cases}$$

(36)

where $\theta_1$ is the volumetric water content of the soil surface layer and $\theta_{fc}$ is water content at field capacity.

2.8. Canopy evapotranspiration

Evaporation from the wet canopy ($E_{wc}$) is calculated using the bulk transfer method:

$$E_{wc} = f_{wet} \frac{q_{sat}(T_c) - q_a}{r_a}$$

(37)

where $q_{sat}(T_c)$ is saturation specific humidity at the canopy temperature and $f_{wet}$ is the wet fraction of the canopy.

Fig. 3. Same as Fig. 2, but for a heterogeneous soil.
canopy, parameterized following Deardorff (1978):

\[ f_{\text{wet}} = \left( \frac{W_c}{W_{\text{Max}}} \right)^{2/3} \]  

(38)

where \( W_c \) is the canopy water content and \( W_{\text{Max}} \) is the water holding capacity of the canopy which is a function of leaf area index. Following Dickinson (1983), \( W_{\text{Max}} \) in millimeter is obtained as:

\[ W_{\text{Max}} = \mu \text{LAI} \]  

(39)

where \( \mu \) is a constant with reported values of between 0.05 and 0.2. The upper range of \( \mu \) appears to overestimate the interception loss, at least where the subgrid scale variation of precipitation is not considered. A value of \( \mu = 0.1 \) is used in ALSIS.

Transpiration from the dry portion of the canopy is:

\[ T_e = (1 - f_{\text{wet}}) \rho \frac{q_s(T_c) - q_a}{r_s + r_a} \]  

(40)

Consistent with Eq. (35a), Eq. (40) can be written as:

\[ T_e = (1 - f_{\text{wet}}) \rho \beta_c \frac{q_s(T_c) - q_a}{r_a} \]  

(41)

with:

\[ \beta_c = \frac{r_a}{r_s + r_a} \]

where \( r_s \) is canopy resistance calculated by the modified model of Jarvis (1976) by Noilhan and Planton (1989):

\[ r_s = r_{\text{min}} \text{LAI} \frac{F_1 F_2^{-1} F_3^{-1} F_4^{-1}} {F_1 F_2^{-1} F_3^{-1} F_4^{-1}} \]  

(42)

where \( F_1 \) is the effect of the photosynthetically active radiation, parameterized following Dickinson et al. (1986):

\[ F_1 = \frac{1 + f}{f + \frac{r_{\text{min}}}{r_s}} \]  

(43a)

with:

\[ f = 0.55 \frac{R_G}{R_{\text{Gl}}} \frac{2}{\text{LAI}} \]

where \( R_G \) is solar radiation and \( R_{\text{Gl}} \) is a limiting value of 30 W m\(^{-2}\) for forest and of 100 W m\(^{-2}\) for crops and grassland. The factor \( F_2 \) measures the effect of the water stress on the canopy resistance:

\[ F_2 = \begin{cases} 1, & \theta > \theta_{cr} \\ \frac{\theta - \theta_{\text{wilt}}}{\theta_{cr} - \theta_{\text{wilt}}}, & \theta_{\text{wilt}} < \theta < \theta_{cr} \\ 0, & \theta < \theta_{\text{wilt}} \end{cases} \]  

(43b)

where \( \theta \) is the mean volumetric water in the root zone and \( \theta_{cr} \) is a critical value of 0.75\( \theta_{\text{sat}} \). \( F_3 \) represents the
Table 1
Comparison of mean annual volumetric water content of five soil layers simulated for homogeneous soil ($\overline{\theta}$) and simulated for heterogeneous soil ($\overline{\theta}_h$) with observations from HAPEX-MOBILHY ($\overline{\theta}_{obs}$).

<table>
<thead>
<tr>
<th>Soil layer</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
</tr>
</thead>
<tbody>
<tr>
<td>($\overline{\theta}_{o_h}$)</td>
<td>0.222</td>
<td>0.221</td>
<td>0.224</td>
<td>0.266</td>
<td>0.317</td>
</tr>
<tr>
<td>($\overline{\theta}_h$)</td>
<td>0.229</td>
<td>0.231</td>
<td>0.235</td>
<td>0.248</td>
<td>0.255</td>
</tr>
<tr>
<td>($\overline{\theta}_{o_h}$)</td>
<td>0.219</td>
<td>0.221</td>
<td>0.245</td>
<td>0.258</td>
<td>0.311</td>
</tr>
<tr>
<td>RSME$_1$</td>
<td>5.24e-2</td>
<td>3.38e-2</td>
<td>2.35e-2</td>
<td>2.03e-2</td>
<td>6.21e-2</td>
</tr>
<tr>
<td>RSME$_2$</td>
<td>5.07e-2</td>
<td>2.83e-2</td>
<td>2.65e-2</td>
<td>1.28e-2</td>
<td>2.55e-2</td>
</tr>
</tbody>
</table>

RSME$_1$ and RSME$_2$ are the root mean square error of the model results.
The unit of the variables is m$^3$ m$^{-3}$.

Fig. 4. Comparison of simulated profile of soil water content (open circles) with measurements from HAPEX (solid circles) for year 1987.
The figures in the brackets are the number of the days from the 1st of January 1987.
effect of the water vapour deficit ($\delta q$) and is the same as in Jarvis (1976):

$$F_s = 1 - 0.6\delta q$$  \hspace{1cm} (43c)

$F_s$ is the factor describing the effect of the air temperature on the surface resistance and parameterized as in Dickinson et al. (1986):

$$F_s = 1 - 0.0016(298 - T_s)^2$$  \hspace{1cm} (43d)

3. Model validation

In Section 2, we described the important components of the ALSIS land–surface scheme. In this section the stand-alone simulation of soil moisture and surface water and energy fluxes by ALSIS will be presented and compared with observations. The two data sets used in the present study are from the HAPEX-MOBILHY in southern France (André et al., 1986) and Cabauw in the Netherlands (Beljaars and Bosveld, 1997). These two data sets have been extensively used in Phase 2a and Phase 2b of the Project for Intercomparison of

![Figure 5. Comparison of simulated diurnal cycle of net radiation ($R_{net}$), latent heat ($LH$), sensible heat ($SH$), and ground heat flux ($GH$) for a homogeneous soil with observations from HAPEX for the observational period of 28 May–30 June 1987.](image_url)
Land–Surface Parameterization Schemes (PILPS: Henderson-Sellers et al., 1995). The information about Phase 2a and Phase 2b PILPS experiments can be found in Chen et al. (1997) and Shao et al. (1994), respectively.

3.1. HAPEX-MOBILHY

The data were mostly obtained from HAPEX-MOBILHY at Caumont (SAMER No. 3, 43°41’N, 0°6’W and a mean altitude of 113 m: Goutorbe and Tarrieu 1991). In the case of missing atmospheric data, measurements from neighbouring meteorological stations were used. Therefore, the forcing data may not be fully consistent with the validation measurements for short and intermittent time periods. We will mention some of the inconsistencies in Section 3.1.1. However, this small inconsistency in the data should not have significant effect on the validation of land–surface schemes, particularly if time step values are not of concern (J.-F. Mahfouf and J. Noilhan, personal communication).

The chosen location is a soya crop field that germinated in May and is harvested at the end of September. The forcing data consist of the measured values of wind speed, specific humidity and air temperature at the screen height (2 m) and precipitation, solar radiation, longwave downward radiation and surface atmospheric pressure at half hour intervals. The validation data consists of annual changes of soil moisture and 38 days of the surface energy components. The measurements of soil moisture is conducted weekly at every 0.1 m from the

![Fig. 6. Same as Fig. 5, but for a heterogeneous soil.](image-url)
surface down to 1.6 m using neutron sounding probes. Net radiation, sensible heat flux and ground heat flux are measured every 15 min during the Intensive Observation Period (28 May–3 July) at Caumont. Latent heat flux is calculated as the residual term to close the surface energy balance. Goutorbe and Tarrieu (1991) reported that on a 15 min basis, the measurement error is 12% for sensible heat and around 25% for latent heat flux. An error of 10% is typical for the soil moisture measurements (J.-F. Mahfouf and J. Noilhan, personal communication).

3.1.1. Soil moisture

Fig. 1 compares the simulated annual cycle of total water content in the top 0.5 m and 1.6 m of soil with observations for HAPEX-MOBILHY. In addition to the Clapp and Hornberger (1978) model, for which soil hydraulic parameters are provided in Shao et al. (1994), simulation using Brooks and Corey (1966), van Genuchten (1980), and Broadbridge and White (1988) models for a loamy soil are presented. For all hydraulic models the same values of saturation hydraulic conductivity and saturation water content are used. The shape and scaling parameters for a loamy soil were derived from Rawls and Brakensiek (1982), Carsel and Parrish (1988), and Shao et al. (1997) for BC, VG and BW models, respectively. The air dry water content for different models is specified so that similar values for wilting point (water content at $\psi = -150$ m) are achieved using any hydraulic model.

![Fig. 7. Comparison of cumulative surface fluxes during IOP as simulated for a homogeneous and a heterogeneous soil with observations from HAPEX.](image-url)
Fig. 1 shows that ALSIS correctly predicts the annual cycle of soil moisture: frequent rainfall and low evaporation keep soil moisture close to the field capacity for the first four months of the year; as precipitation decreases and available energy for evaporation increases, soil water begins to deplete at the beginning of the growing season (early May) and reaches a minimum during August to October. For both of the soil layers and using different soil hydraulic models, ALSIS simulates soil moisture comparable with observations during the cold wet season. However, the summer time simulated soil moisture is higher in the top 0.5 m and lower in the top 1.6 m compared to observations. To show the deviation of simulated soil moisture profile from observations, Fig. 2 compares the modelled and observed volumetric water content of the five soil layers. Fig. 2 clearly shows that the summer time soil moisture is overestimated down to 0.5 m soil depth, is comparable with observations for 0.5 m to 1.0 m soil depth, and is underestimated for the lowest 0.6 m of soil. Since the disagreement between simulated and observed soil moisture occurs mainly during the growing season, one may relate this discrepancy to wrong specification of vertical distribution of plant roots (Liang et al., 1996). However, because the soil moisture is underestimated greatly in the lowest soil layer, where root fraction is set to zero for simulations, this conclusion may not be correct.

The underestimation of the deep soil moisture seems to be due to inappropriate choice of the soil hydraulic parameters. This may cause excessive drainage rate, resulting in a drier soil. For the results presented here the lower boundary condition of soil is set to gravitational drainage, in which the water flux becomes zero only when soil moisture at the boundary drops to its air dry value. Several ad hoc methods can be applied to prevent the decrease in deep soil moisture content. One method is to set a soil moisture threshold, for instance field capacity, below which no drainage occurs. Another method is to reduce the drainage rate by introducing a drainage coefficient. Finally, it is possible to push the lower boundary downward by increasing the depth of the

![Graphs](image_url)
soil column, so that for the soil layers of interest the influence of the unknown lower boundary conditions becomes insignificant. We believe that the first solution is not physically meaningful. The second approach is appropriate only if the soil column is underlaid by a low or non-permeable layer. The third way is theoretically sound, but in this way we may face a crucially higher computation expense and longer spin up time, due to erratic initialization.

Overestimation of minimum moisture in the upper layer, where soil moisture approaches its air dry value, and its underestimation in the deepest layer implies that the assumption of a vertically homogeneous soil for HAPEX may be incorrect. In a cultivated land, such as the HAPEX site, apart from natural vertical heterogeneity in texture, the top (20–30 cm) soil is ploughed, causing a substantial change in the soil structure. This may increase the top soil macroporosity, resulting in an increased permeability, a decreased capillarity, and a decreased water holding capacity. Therefore, applying the mean values of hydraulic parameters to total soil layer may cause deviations from observations with different signs for the top and deep soil layers.

The impact of accounting for vertical variation of soil hydraulic parameters on the prediction of soil moisture is shown in Fig. 3. For this numerical experiment, we divided the soil into three vertical regions (horizons), the upper three layers (0.5 m) being coarser sandy loam than, the 4th layer (next 0.5 m) being the same as, and the rest of the soil being finer clayey loam than the loamy soil used in the previous test. Since the annual cycle of soil moisture simulated using different models showed similar patterns, in Fig. 3 the results of two of them (BW and CH models) are shown.

The results presented in Figs. 2 and 3 reveal the positive impact of accounting for vertical heterogeneity of soil hydraulic properties in soil moisture simulation. In general, the annual trend of simulated soil moisture in different layers show a much better agreement with observations in Fig. 3. Table 1 compares the mean

![Fig. 9. Comparison of simulated cumulative surface fluxes with observations from Cabauw.](image-url)
simulated and observed volumetric water content of different soil layers for the 40 observational times. To be consistent with most land surface parameterization schemes, the simulations using the Clapp and Hornberger model are presented in Table 1. From the table it is clear that the inclusion of vertical heterogeneity not only improves the annual cycle, but also results in a long term mean soil water content more consistent with observations. The mean root square error of simulated soil moisture for heterogeneous soil, presented in Table 1, is generally lower than that for homogeneous soil.

Fig. 4 compares the simulated soil moisture profile with observations for observational days. The overall agreement between the simulated and measured soil moisture profile can be more clearly seen on this figure. However, for some of the times the deviations are considerable. Apart from uncertainty in model simulations, this discrepancy may be in part due to observational errors and/or inconsistencies between the measured soil moisture and atmospheric forcing (precipitation). Examples of the latter are days 7, 25, 324 and 331, when the observed profiles show a wet surface layer, typical for rain events, but for a few days prior to these observational days, precipitation was not reported in the atmospheric forcing data. The measured soil moisture increase from day 317 to day 324 is not consistent with measured precipitation, even if we assume zero evaporation and runoff during the period.

3.1.2. Surface fluxes

In Fig. 5, simulated and observed diurnal cycles of surface net radiation and the non-radiative heat fluxes for the Intensive Observational Period (IOP) at HAPEX-MOILHY are compared. The results presented in Fig. 5
for a homogeneous soil shows that ALSIS simulates the surface net radiation quite well (Fig. 5a). The scheme correctly predicts the diurnal trend of sensible heat flux (Fig. 5c). However, the simulated sensible flux has larger diurnal amplitude than that observed. This is in association with the model’s underestimation of evaporation during the day and its overestimation at nights (Fig. 5b).

The simulated surface energy fluxes for HAPEX using the heterogeneous soil are shown in Fig. 6. The inclusion of soil heterogeneity improved the maximum day-time evapotranspiration. However, the lower diurnal amplitude of latent heat and the higher amplitudes of sensible heat and ground heat fluxes remain unchanged. The model’s failure in correctly simulating night-time dew may be attributed to the nature of the $\beta$ formulation for moisture stress term in evaporation routine (see Mahfouf and Noilhan, 1991).

To compare the impact of soil heterogeneity on surface flux simulation, the results presented in Figs. 5 and 6 are shown in Fig. 7 in cumulative term. The unit used to express the surface fluxes is the energy required to evaporate 1 mm of water. The simulated annual evapotranspiration for cases 1 and 2 are 598 and 603 mm compared to 580 mm reported for HAPEX (see Shao and Henderson-Sellers, 1996). On the other hand, the scheme predicts 104 mm evapotranspiration during the IOP for the homogeneous case that is less than 126 mm observed as the residual of the surface energy balance (Mahfouf et al., 1996). Using the heterogeneous soil for HAPEX not only improves the simulation of soil moisture profile, but also results in total IOP evaporation of 118 mm, more consistent with that observed. The improved estimation of total evaporation is associated with a decreased total sensible heat flux from 57 mm in case 1 to 46 mm in case 2, while the total sensible heat flux measured during the IOP is 36 mm. The total measured flux of heat into the soil is 0.2 mm, while the model simulates about 2.2 mm ground heat flux during the IOP.

![Fig. 11. Comparison of diurnal cycle of surface fluxes and surface temperature with observations from Cabauw for the Julian days 253 to 262.](image-url)
3.2. Cabauw

The Cabauw data have been collected at a 213 m meteorological mast in Cabauw, the Netherlands (51°58' N, 4°55' E). The site is located in a flat terrain consisting mainly of meadow interrupted by narrow ditches. The measuring site itself has grass that is kept at about 8 cm height by frequent mowing. Up to a distance of 200 m from the mast, there is no obstacle or perturbation of any importance. Further away there are some scattered trees and houses. The climate of the area is characterized as moderate maritime with prevailing westerly circulation. The measuring program that was operational in 1987 is described in more detail by Monna and Van der Vliet (1987). The soil consists of a 1 m deep layer clay on top of a 10 m deep layer of peat saturated with water. Cabauw data set consists of half hour averages of forcing data (wind, temperature, specific humidity at 20 m height, downward solar and thermal radiation and precipitation) and validation data (net radiation, sensible heat flux, latent heat flux, ground heat flux and soil temperature). For more information about Cabauw site, soil and vegetation characteristics, and forcing and validation data, see Beljaars and Bosveld (1997).

3.2.1. The Cabauw results

The monthly means of surface heat fluxes and surface temperature simulated by ALSIS are compared with observations from Cabauw in Fig. 8. ALSIS captures the annual trend of observed fields reasonably well. However, the simulated net radiation Fig. 8a and sensible heat flux Fig. 8c show a systematic overestimation almost all over the year. The higher simulated net radiation, despite higher rate of longwave radiation emitted from the surface (not shown, but implied in the higher simulated surface temperature Fig. 8e) shows that surface

Fig. 12. Same as Fig. 11, but for Julian days 6 to 13.
absorbs too much radiative energy. This may be due to wrong specification of surface albedo. Observations show two maxima in May and July for sensible heat from the surface. However, sensible heat flux simulated by ALSIS shows a maximum in May and decreases monotonically to December.

The simulated monthly latent heat (Fig. 8b) is consistent with observations, with exceptions in January and May. The higher January simulated latent heat can be related again to the inefficiency of evaporation routine (β formulation) in calculating strong downward moisture flux. During the first 4 months of the year, dew formation is not uncommon in Cabauw. Observations show the mean daily net dew formation of up to 1.7 mm for some days, and instantaneous latent heat flux of down to −80 W m⁻² for many time steps. For the same time period, the simulated latent heat flux seldom drops to −20 W m⁻². The model overestimates the annual range of ground heat flux. This can be related to the fact that a layer of dead vegetation with a low heat conductivity may exist naturally over the soil, while the simulations do not take the effect of this semi-isolating layer into account.

In Fig. 9, the cumulative annual surface fluxes are presented. The mean annual simulated net radiation and sensible heat are 1373.9 MJ m⁻² and 83.3 MJ m⁻², higher than observations 1283.1 MJ m⁻² and 31.0 MJ m⁻², respectively. These correspond to simulated and observed mean annual net radiation of 43.6 W m⁻² and 40.7 W m⁻² and sensible heat flux of 2.6 W m⁻² and 1.0 W m⁻², respectively. ALSIS simulates annual latent heat flux 1291.1 MJ m⁻² (mean 40.9 W m⁻²), similar to 1290.5 MJ m⁻² (mean 40.9 W m⁻²) observed. Observations show that for the study year (1987), about 17.3 MJ m⁻² energy lost from the surface originated from storage. Because the model results are reported for an equilibrium year, the net simulated ground heat flux is much smaller (0.01 W m⁻²). The mean annual temperature is 281.2 K which is slightly higher than observation of 280.6.

In Fig. 10, the mean daily simulated and observed surface fluxes and surface temperature are compared. The daily mean simulations show reasonable agreement with observations. However, ALSIS overestimates the day to day changes in ground heat flux, and fails to represent large negative fluxes of net radiation and latent heat in the early days of the year.

Fig. 13. Scatter diagrams of simulated vs. observed net radiation (1), latent heat flux (2), sensible heat flux (3), ground heat flux (4) and surface temperature (5) at half-hourly (a), daily (b), and monthly (c) time scales for Cabauw.
The simulated diurnal fluctuation of surface fluxes and temperature for the period of Julian days 253 to 262, as required by PILPS Phase 2c experiment, are compared with observations in Fig. 11. Fig. 11 shows that ALSIS can capture short time observations quite well, for the period shown. However, for some periods in the early time of the year, the diurnal simulation is not in agreement with measurements. As an example of such cases the results for days 6 to 13 are shown in Fig. 12. The overall comparison of the results with observations in different time scales are shown in the scatter diagrams of Fig. 13. The top panels (a) have a half-hourly time scale, and the middle (b) and bottom (c) panels are for daily and monthly averages, respectively. Columns 1–5 of each row represent net radiation, latent heat, sensible heat, ground heat, and surface temperature, respectively. Fig. 13 shows the good agreement between simulations and observations at different time scales. Except for the ground heat flux, the slope of regression lines are not very different from 1. The coefficients of correlation vary between 0.990 to 0.996 for net radiation, 0.986 to 0.999 for surface temperature, 0.928 to 0.992 for net radiation, 0.986 to 0.999 for surface temperature, 0.928 to 0.992 for latent heat, and 0.890 to 0.929 for sensible heat flux in different time scales. The ground heat flux shows not only the steepest regression coefficient, but also has largest deviations from the regression line. The correlation coefficients between simulated and observed ground heat varies between 0.770 to 0.929 from half-hourly to monthly time scales. The standard error of coefficients are usually small.

4. Conclusion

A new atmosphere–land–surface interaction scheme (ALSIS) was described. The scheme accounts for the effects of different vegetation types on the surface roughness, on the exchanges of heat and moisture between the surface and the atmospheric boundary layer, and on water uptake from the soil. The evolution of soil moisture is simulated in multilayer vertically heterogeneous soil using the Kirchhoff transform to the diffusion term of the Richard’s Equation. The soil hydraulic models of Brooks and Corey (1966), Clapp and Hornberger (1978), van Genuchten (1980) and Broadbridge and White (1988) were used to describe the relationships among soil hydraulic properties. The surface layer similarity theory is used to calculate the bulk transfer coefficient. The empirical relationship presented by Raupach (1994) is applied to calculate the surface roughness and zero-plane displacement over the vegetation. The scheme applies an exponential model to describe root distribution in different soil layers. Canopy temperature is calculated by solving the canopy energy balance equation assuming zero heat storage for the canopy. Soil temperature is calculated prognostically using the Fourier law of heat transfer. Surface soil temperature is diagnosed by solving the soil surface energy balance equation.

The simulated soil moisture for HAPEX site using a vertically homogenous soil has a positive bias in the upper soil layers and a negative bias in the deep soil layers. Taking into account the soil vertical heterogeneity eliminates this discrepancy and results in an excellent agreement between annual cycles of modelled and observed soil moisture profile. The mean annual soil moisture in the top 1.6 m of soil increased from 394 mm for homogeneous case to 433 mm for the heterogeneous case, consistent with 435 mm observed.

ALSIS showed reasonable skill in predicting surface net radiation and surface heat fluxes for HAPEX. The inclusion of soil heterogeneity not only improved soil moisture simulation, but also resulted in improved skill in predicting the mean and the diurnal cycles of surface fluxes for the intensive observational period (28 May–3 July). For instance, the total evapotranspiration simulated for the IOP is 104 mm using a homogeneous soil. The inclusion of soil vertical heterogeneity increases total evapotranspiration to 188 mm, more consistent with 126 mm observed.

The simulated annual cycles of surface temperature and surface fluxes for Cabauw are consistent with observations. In the annual scale, the modelled mean latent flux of 40.9 W m$^{-2}$ is similar to that observed. Compared to observations, the scheme has a positive bias of 2.9 W m$^{-2}$ in net radiation, 1.6 W m$^{-2}$ in sensible heat, and 0.6 C° in surface temperature. The monthly averages of surface temperature and net radiation show a
systematic overestimation for Cabauw. However, the deviation of modelled monthly mean surface fluxes from observation are well within the estimated observational errors (10 W m$^{-2}$, 10 W m$^{-2}$, and 5 W m$^{-2}$ for net radiation, latent heat flux, and sensible heat flux, respectively). The modelled diurnal cycles of temperature and fluxes are in agreement with those observed. However, the model overestimates the night-time latent heat flux, especially during January. This may be attributed to the $\beta$-type formulation of evaporation used in ALSIS. The impact of including different parameterizations of soil moisture constraint ($\alpha$ and $\beta$) on improving this feature of the scheme is yet to be studied.

The results presented here show that ALSIS is capable in simulating the annual cycle of soil moisture profile. The diurnal and annual cycles of surface temperature and partitioning of surface energy between sensible and latent heat fluxes can be simulated with reasonable accuracy. The simulation of night-time evaporation, especially in cold conditions, needs further considerations.

Acknowledgements

The valuable comments of two anonymous reviewers are greatly appreciated.

Appendix A. Appendix A: mathematical symbols

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>$b$</td>
<td>pore size distribution parameter for the Brooks–Corey-type and models</td>
<td></td>
</tr>
<tr>
<td>$C$</td>
<td>pore size distribution parameter for the Broadbridge–White and model</td>
<td></td>
</tr>
<tr>
<td>$C_p$</td>
<td>bulk transfer coefficient for heat and vapor</td>
<td>$-1$</td>
</tr>
<tr>
<td>$C_{ph}$</td>
<td>bulk transfer coefficient for momentum</td>
<td></td>
</tr>
<tr>
<td>$C_h$</td>
<td>volumetric heat capacity of soil</td>
<td>$J m^{-3} K^{-1}$</td>
</tr>
<tr>
<td>$C_c$</td>
<td>drag coefficient of an isolated roughness element</td>
<td>$-$</td>
</tr>
<tr>
<td>$C_s$</td>
<td>drag coefficient of the substrate surface at canopy height in the absence of roughness elements</td>
<td>$-$</td>
</tr>
<tr>
<td>$c_p$</td>
<td>specific heat of air at constant pressure</td>
<td>$J kg^{-1} K^{-1}$</td>
</tr>
<tr>
<td>$c_q$</td>
<td>specific heat capacity of quartz</td>
<td>$J kg^{-1} K^{-1}$</td>
</tr>
<tr>
<td>$c_s$</td>
<td>specific heat capacity of soil</td>
<td>$J kg^{-1} K^{-1}$</td>
</tr>
<tr>
<td>$e_w$</td>
<td>specific heat capacity of water</td>
<td>$J kg^{-1} K^{-1}$</td>
</tr>
<tr>
<td>$D$</td>
<td>soil water diffusivity</td>
<td>$m^2 s^{-1}$</td>
</tr>
<tr>
<td>$D_w$</td>
<td>water drainage from the canopy</td>
<td>$kg m^{-2} s^{-1}$</td>
</tr>
<tr>
<td>$D_h$</td>
<td>soil heat diffusivity</td>
<td>$m^2 s^{-1}$</td>
</tr>
<tr>
<td>$E_c$</td>
<td>canopy evapotranspiration</td>
<td>$kg m^{-2} s^{-1}$</td>
</tr>
<tr>
<td>$E_v$</td>
<td>soil evaporation</td>
<td>$kg m^{-2} s^{-1}$</td>
</tr>
<tr>
<td>$E_{wc}$</td>
<td>evaporation from the wet canopy</td>
<td>$kg m^{-2} s^{-1}$</td>
</tr>
<tr>
<td>$f_{ext}$</td>
<td>radiation fraction on the soil surface</td>
<td>$-$</td>
</tr>
<tr>
<td>$f_{wet}$</td>
<td>wet fraction of the canopy</td>
<td>$-$</td>
</tr>
<tr>
<td>$g$</td>
<td>acceleration due to gravity</td>
<td>$m s^{-2}$</td>
</tr>
<tr>
<td>$G_h$</td>
<td>soil heat flux</td>
<td>$W m^{-2}$</td>
</tr>
<tr>
<td>$h_c$</td>
<td>canopy height</td>
<td>$m$</td>
</tr>
<tr>
<td>$H_c$</td>
<td>canopy sensible heat flux</td>
<td>$W m^{-2}$</td>
</tr>
<tr>
<td>$H_s$</td>
<td>soil surface sensible heat flux</td>
<td>$W m^{-2}$</td>
</tr>
<tr>
<td>$K$</td>
<td>soil hydraulic conductivity</td>
<td>$m s^{-1}$</td>
</tr>
</tbody>
</table>
\[ K_h \] soil thermal conductivity \[ \text{W} \text{m}^{-1} \text{K}^{-1} \]
\[ K_s \] saturation hydraulic conductivity \[ \text{m} \text{s}^{-1} \]
\[ L \] Monin–Obukhov length \[ \text{m} \]
\[ \text{LAI} \] leaf area index \[ - \]
\[ m, n \] soil parameters of the van Genuchten model \[ - \]
\[ P_c \] precipitation rate on the canopy \[ \text{mm} \text{s}^{-1} \]
\[ q \] soil water flux \[ \text{m} \text{s}^{-1} \]
\[ q_a \] specific humidity of air at reference height \[ \text{kg} \text{kg}^{-1} \]
\[ q_s \] specific humidity at the roughness height \[ \text{kg} \text{kg}^{-1} \]
\[ q_{sat} \] saturation specific humidity at the soil surface temperature \[ \text{kg} \text{kg}^{-1} \]
\[ r_a \] aerodynamic resistance for moisture and heat fluxes \[ \text{s} \text{m}^{-1} \]
\[ r_s \] canopy resistance \[ \text{s} \text{m}^{-1} \]
\[ r_{\text{amin}} \] minimum stomatal resistance \[ \text{s} \text{m}^{-1} \]
\[ r_{\text{amax}} \] maximum stomatal resistance \[ \text{s} \text{m}^{-1} \]
\[ R_d \] downward longwave radiation \[ \text{W} \text{m}^{-2} \]
\[ R_{dc} \] longwave radiation absorbed by the canopy \[ \text{W} \text{m}^{-2} \]
\[ R_{ds} \] longwave radiation absorbed by the soil surface \[ \text{W} \text{m}^{-2} \]
\[ R_{sc} \] longwave radiation emitted by the canopy \[ \text{W} \text{m}^{-2} \]
\[ R_{ss} \] longwave radiation emitted by the soil \[ \text{W} \text{m}^{-2} \]
\[ R_n \] land surface net radiation \[ \text{W} \text{m}^{-2} \]
\[ R_{nc} \] canopy net radiation \[ \text{W} \text{m}^{-2} \]
\[ R_{ns} \] soil surface net radiation \[ \text{W} \text{m}^{-2} \]
\[ R_s \] shortwave radiation intensity at the top of the canopy \[ \text{W} \text{m}^{-2} \]
\[ R_{sc} \] radiation absorbed by canopy \[ \text{W} \text{m}^{-2} \]
\[ R_{ss} \] radiation absorbed at the soil surface \[ \text{W} \text{m}^{-2} \]
\[ R_{s,b} \] incident solar radiation on the bare soil \[ \text{W} \text{m}^{-2} \]
\[ R_{s,c} \] and incident solar radiation on the soil under the canopy \[ \text{W} \text{m}^{-2} \]
\[ R(z) \] cumulative root below depth \[ z \] \[ - \]
\[ T_a \] air temperature \[ \text{K} \]
\[ T_c \] canopy temperature \[ \text{K} \]
\[ T_s \] transpiration rate \[ \text{kg} \text{m}^{-2} \text{s}^{-1} \]
\[ T_z \] soil surface temperature \[ \text{K} \]
\[ U \] Kirchhoff transform \[ \text{m}^2 \text{s}^{-1} \]
\[ U_p \] wind velocity \[ \text{m} \text{s}^{-1} \]
\[ u_* \] friction velocity \[ \text{m} \text{s}^{-1} \]
\[ W_c \] canopy water storage \[ \text{kg} \text{m}^{-2} \]
\[ W_{\text{c,Max}} \] water holding capacity of the canopy \[ \text{kg} \text{m}^{-2} \]
\[ z \] soil depth \[ \text{m} \]
\[ z_{oh} \] surface roughness length for heat \[ \text{m} \]
\[ z_{om} \] surface roughness length for momentum \[ \text{m} \]
\[ z_d \] zero-plane displacement \[ \text{m} \]
\[ z_r \] atmospheric reference height \[ \text{m} \]
\[ z_w \] upper height limit of the roughness sublayer \[ \text{m} \]
\[ \alpha \] soil parameter of the van Genuchten model \[ \text{m}^{-1} \]
\[ \alpha_c \] canopy albedo \[ - \]
\[ \alpha_s \] soil surface albedo \[ - \]
\[ \epsilon_c \] canopy emissivity \[ - \]
\[ \epsilon_s \] soil surface emissivity \[ - \]
The references include a variety of studies on land surface processes and climate models. Some of the key studies mentioned are:


