ELSEVIER

Global and Planetary Change 13 (1996) 135–143

GLOBAL AND PLANETARY CHANGE

Brief description of bare essentials of surface transfer and results from simulations with the HAPEX–MOBILHY data

A.J. Pitman *, C.E. Desborough

School of Earth Sciences, Macquarie University, North Ryde, 2109 NSW, Australia

Received 5 May 1995; accepted 28 August 1995

Abstract

The Bare Essentials of Surface Transfer (BEST) land surface scheme is briefly described and the key physical parameterisations discussed. Results are then presented to illustrate how the model performs, with forcing data for HAPEX–MOBILHY, compared to a series of other schemes in the simulation of evaporation and sensible heat. The implications of the models performance, and some indications of the future development of the scheme are provided. The basic version of BEST was found to overestimate evaporation for the HAPEX–MOBILHY data, simulating 816 mm yr^{-1} compared to a range of 550 to 816 mm yr^{-1} for all models. A modification to the β parameterisation reduced the evaporation to 759 mm yr^{-1} which, although an improvement, is still probably too high.

1. Introduction

This paper describes the physics of the land surface scheme ''Bare Essentials of Surface Transfer'' (BEST). BEST was originally developed by Pitman (1988), improved and generalised by Cogley et al. (1990) and implemented into the Australian Bureau of Meteorology Research Centre (BMRC) Atmospheric General Circulation Model (AGCM) by Pitman et al. (1991). Results reported by, for instance Yang et al. (1995), indicate that the model performed reasonably within the BMRC AGCM.

BEST is a relatively complex land surface scheme and requires the input of incident shortwave (solar) and longwave (atmospheric) irradiance, precipitation, wind vector components, surface pressure, air temperature and specific humidity. These can be provided from either an offline data set or from a host model. The model outputs include the albedo, temperature, wetness and roughness length of the surface which may be expressed as fluxes of radiation, sensible and latent heat, heat conducted into the subsurface, momentum and runoff.

The parameterisation of the surface energy balance and the fluxes between the surface and the atmosphere was the basic reason behind incorporating the land surface into AGCMs. An AGCM requires relatively little from the land surface model (principally the momentum flux, infrared radiation, sensible and latent heat). In order to predict these quantities, the surface roughness length is needed for the momentum flux. The surface temperature and emissivity are needed for the infrared radiation which requires a soil temperature and perhaps a canopy temperature to be predicted. In order to simulate the sensible and latent heat fluxes, the surface tempera-

* Corresponding author. E-mail: apitman@penman.es.mq.edu.au; fax: +61 2 850 8428; phone: +61 2 850 8425.

0921-8181/96/$15.00 Published by Elsevier Science B.V.
SSDI 0921-8181(95)00042-9

ture and wetness are needed. This requires interception, infiltration, soil moisture distribution and runoff to be simulated.

In this paper, a brief description of how BEST parameterises these quantities is presented, followed by a description of the models performance against the HAPEX-MOBILHY data. These experiments formed part of the Workshop held at Macquarie University in November 1994 and discussed elsewhere in this volume (e.g. Shao and Henderson-Sellers, 1996-this issue).

2. Physical structure of best

BEST splits the grid square into two basic components (see Fig. 1). A fraction of the grid square (A_v) is covered by the canopy air space. The canopy air space can be further divided into a fraction (also A_v) which is directly shaded by vegetation and a bare ground fraction ($1-A_v$). The fraction of the grid square which is actually shaded by vegetation is thus A_v^2. Both the canopy and bare ground within the canopy air space interact with the atmosphere only through the canopy air space. This is one of the ways that BEST allows for sub-grid scale heterogeneity.

The second fraction of the grid square ($1-A_v$) is bare soil and interacts directly with the atmosphere and does not interact with the canopy air space.

For energy balance purposes, snow may cover a fraction (A_n) of the grid square. It is considered to be evenly distributed between vegetation and non-vegetation portions of the ground. For below ground temperature and soil moisture calculations the snow is considered to be uniformly distributed across the surface. The ground is treated as a snow–soil composite and depths are measured from the top of the snow pack rather than from the top of the soil.

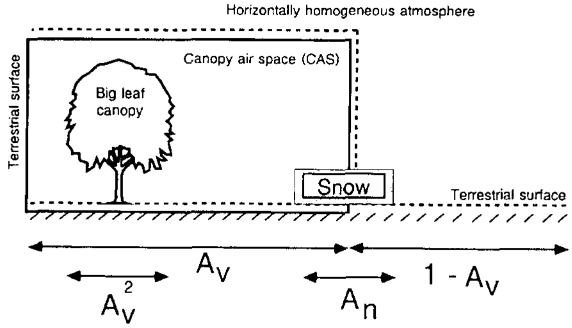

Fig. 1. Diagram of the relationship between the fractional extents of bare soil ($1-A_v$), the canopy air space (A_v), actual vegetation (A_v^2) and snow (A_n) in BEST.

3. Input parameters

In addition to the atmospheric forcing, BEST requires values for a series of parameters which describe the characteristics of the vegetation (Table 1) and soil (Table 2). The values for these parameters can be taken from any available source, but BEST usually uses data of Wilson and Henderson-Sellers (1985).

4. Derived parameters

4.1. Albedo

Henderson-Sellers and Wilson (1983) showed that the albedo of many surfaces is a strong function of wavelength. The albedo (α_i) in BEST is therefore calculated in two wavelength bands (visible and near-infrared) split at 0.7 μm for each of the canopy, soil and snow such that:

$$\alpha_i = 0.5(\alpha_{SW,i} + \alpha_{LW,i}) \quad (1)$$

where α_{SW} is the albedo for the visible wavelength region (≤ 0.7 μm), α_{LW} is the albedo for the infrared wavelength region (> 0.7 μm). Canopy albedos are prescribed (Table 1) while the soil albedos depend on soil colour and soil wetness and the snow albedos depend on temperature and snow age.

This parameterisation of albedo is intentionally simple. There is no parameterisation of the effects of the sun's zenith angle on the albedo (in contrast to BATS, Dickinson et al., 1986) or radiative transfer in the canopy. However, although the formulation described here for BEST is simple, it does differentiate between soil, snow and vegetation albedo and is computationally inexpensive.

4.2. Surface roughness length

Simulations with AGCMs have shown that the climate is sensitive to the surface roughness parame-

Table 1
List of all canopy parameters needed in BEST

Parameter	Definition
Parameters dependent on canopy type	
A_{vmax}	Maximum fractional vegetation
S_c	Seasonal variability of A_{vmax}
L_{AImax}	Maximum leaf area index
S_l	Seasonal variability in leaf area
S_{AI}	Stem area index
d_u	Upper soil layer depth (0.1 m)
d_l	Lower soil layer depth (m)
f_{rootu}	Fraction of total roots in top soil
z_{0c}	Roughness length of canopy (m)
α_{SWc}	Shortwave canopy albedo
α_{LWc}	Infrared albedo of canopy (> 0.7 μm)
Parameters independent of canopy type	
D_{max}	Canopy interception capacity (m)
E_{tr0}	Minimum transpiration rate (m s^{-1})
C_f	Reference leaf conductance (s m^{-1})
S_f	Leaf dimension (for aerodynamic resistance, m)
r_{smin}	Minimum stomatal resistance (s m^{-1})
r_{smax}	Maximum stomatal resistance (s m^{-1})
z_{0n}	Roughness length of snow surface (m)
z_{0u}	Roughness length of the soil surface (m)

terisation (e.g. Sud et al., 1988). However, there is no rigorous method for dealing with heterogeneous areas within a grid element to derive a single appropriate estimate of roughness although Mason (1988), Andre and Blondin (1986) and Taylor (1987) have suggested possible approaches.

A surface roughness length must be defined for each grid element to describe the efficiency at which surface–atmosphere transfers of energy and momentum take place. We define an effective roughness length which takes account of subgrid scale variability following Taylor (1987), by assuming that the local frictional velocity, u^*, is the same above a patchwork of areas of different roughness lengths. Thus the effective roughness length of the grid element (z_{0t}) is

$$\ln z_{0t} = (1 - A_v) \ln z_{0u} + A_v \ln z_{0c} \qquad (2)$$

where $z_{0u} = 0.01$ m is the bare soil roughness length (Oke, 1978) and z_{0c} is the canopy roughness length which is the function of the specified ecotype.

5. Canopy model

While there have been attempts to parameterise the characteristics of vegetation into AGCMs without explicitly modelling the canopy (e.g. Hansen et al., 1983), BEST incorporates an explicit and physically based model of the canopy. This section describes the parameterisation of vegetation incorporated by BEST.

In BEST, the parameterisation of the atmospheric resistance, r_a, follows Dickinson (1984), Deardorff (1978), Brutsaert (1982) and Gates (1980). A simple multiplicative approach, following Jarvis (1976), is included to represent stomatal resistance. The stomatal resistance is a function of canopy temperature, the amount of photosynthetically active radiation absorbed by the canopy and seasonal factors. Transpiration is calculated as the minimum of atmo-

Table 2
List of all soil types and soil parameters needed in BEST. K_{H0} is the saturated hydraulic conductivity B is the Clapp and Hornberger parameter X_v is the porosity and ψ_0 is the soil water suction at saturation

Soil type	K_{H0} (mm s^{-1})	B	c_{vm} (J m^{-3} K^{-1})	X_v	ψ_0 (m)
Clay	0.001	10.0	2.38×10^6	0.60	−0.2
Sandy clay	0.0015	9.2	2.38×10^6	0.57	−0.2
Clay loam	0.0021	8.4	2.38×10^6	0.54	−0.2
Silt clay loam	0.0032	7.6	2.38×10^6	0.51	−0.2
Sandy clay loam	0.0045	6.8	2.38×10^6	0.48	−0.2
Loam	0.006	6.0	2.38×10^6	0.45	−0.2
Loamy silt	0.0090	5.5	2.38×10^6	0.42	−0.2
Silty loam	0.0150	5.0	2.38×10^6	0.39	−0.2
Sandy loam	0.0400	4.5	2.38×10^6	0.36	−0.2
Sand	0.100	4.0	2.38×10^6	0.33	−0.2

spheric demand (which includes stomatal resistance) and a supply rate incorporating the influences of soil moisture and temperature. A root distribution depending on vegetation type is used to divide transpiration between the soil layers.

The turbulent energy fluxes at the top of the canopy are composited from fluxes from the canopy and from the underlying ground (Deardorff, 1978; Dickinson, 1984). Modelling the canopy heat balance and temperature is extremely complex and in BEST we calculate a full canopy energy balance. Assuming the canopy heat storage is zero, the canopy energy balance can be solved iteratively for the canopy temperature by using the Newton-Raphson method.

The parameterisation of interception and the subsequent redistribution of water to drip or re-evaporation is one of the principal reasons for incorporation canopies into AGCMs. Intercepted water which fails to drip to the surface re-evaporates rapidly, often within a few hours, due to the large roughness of canopies, high ventilation and large surface area in contact with the atmosphere. In contrast, precipitation which reaches and infiltrates the soil tends to remain in the soil for much longer periods.

The water storage on the surfaces of the canopy is determined from the balance of intercepted precipitation and the evaporation of the retained water. In BEST the approach is similar to that of Deardorff (1978) and Dickinson et al. (1986). The retained water covers a fraction of the canopy from which transpiration is suppressed. Transpiration is permitted to continue from the rest of the canopy.

5.1. Turbulent energy fluxes

As with most land surface schemes, BEST calculates both the evaporation and the sensible heat fluxes explicitly. Fig. 2a,b shows the routes taken by sensible and latent heat fluxes. There are three important evaporation fluxes: evaporation from the bare soil fraction (E_{us}), evaporation from water stored on the canopy surface and evaporation of water extracted by roots from the upper and lower soil layers. These latter two fluxes are composited into a single flux (E_{ca}). When E_{ca} is combined with the evaporation flux from the fraction of soil *within* the fraction A_v, the total flux from the entire canopy fraction is

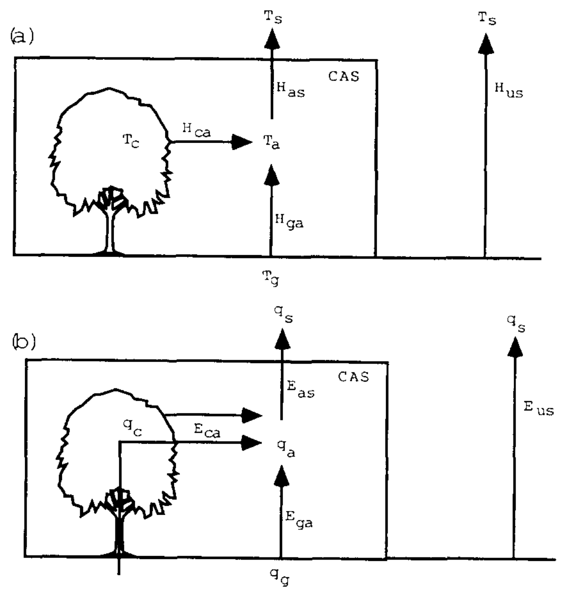

Fig. 2. Diagram showing the pathways taken by (a) sensible and (b) latent heat fluxes in BEST.

produced (E_{as}). The total evaporative flux from the entire land surface to the atmosphere is therefore the sum of E_{as} and E_{us} (Fig. 2b). A similar, although marginally simpler system describes the sensible heat sub-model (Fig. 2a). The general form of the two turbulent energy fluxes between surfaces denoted 1 and 2 are:

$$E = \rho_s U_m C_{Dh} \beta(q_1^* - q_2) \tag{3}$$

$$H = c_p \rho_s U_m C_{Dh}(T_1 - T_2) \tag{4}$$

where ρ_s is the air density, U_m is the wind speed, C_{Dh} is the drag coefficient for heat and moisture, β is the wetness of a surface, q is the vapour pressure and * refers to the saturated value, c_p is the specific heat of air and T is the temperature. The calculation of the evaporation from within the soil, via the canopy to the atmosphere includes the soil–root–xylem conductances following Dickinson et al. (1986). The actual transpiration rate is also calculated taking into account the atmospheric demand and the ability of the canopy–soil system to meet the demand., i.e. the supply and demand method. An effective β parameter is then calculated from estimates of potential and actual evaporation. β_c, the wetness of the canopy, is required in order to calculate the suppression of transpiration and the enhancement of the re-evaporation of intercepted precipitation (defined following Deardorff, 1978).

Historically β_u, the wetness of the soil surface, has been defined as a ratio of total available soil moisture to some maximum possible soil moisture (e.g. "field capacity") which is not a particularly realistic approach since it can not simulate the drying of a soil crust which inhibits evaporation. Dickinson et al. (1986) noted that the top few mm of soil often dry out inhibiting evaporation, irrespective of how wet the soil is beneath and parameterised this by calculating a potential soil evaporation flux according to the mechanical rate at which soil can diffuse towards the surface. BEST takes a similar approach and parameterises the ground wetness, β_u, as

$$\beta_u = \min\left(1, \frac{E_{usmax}}{\rho_s c_u (q_u^* - q_s)}\right) \tag{5}$$

where c_u is a aerodynamic conductance for moisture and E_{usmax} is the maximum exfiltration rate due to the mechanical process of soil moisture uptake within the soil (see Eagleson, 1970). E_{usmax} is:

$$E_{usmax} = K_{HD} W_{L,u}^{0.5B+2} - K_{HO} W_{L,u}^{2B+3} \tag{6}$$

where B is a soil diffusivity parameter (Clapp and Hornberger, 1978), $W_{L,u}$ is the volumetric soil moisture in the upper soil layer (0.1 m) and K_{HO} is the hydraulic conductivity at saturation. Both B and K_{HO} are functions of soil type (see Table 2). K_{HD} is a rate of diffusion from a surface "pond" into dry soil defined as

$$K_{HD} = \left(\frac{-4K_{HO} B \psi_0 \rho_w X_v}{\pi \partial t}\right)^{1/2} \tag{7}$$

where ρ_w is the density of water and ψ_0 is the soil water suction at saturation (Table 2). β_u is therefore an effective β parameter based on the supply and demand philosophy.

6. Soil moisture calculations

BEST adopts a simplified and extended version of the Philip-De Vries theory (Philip and De Vries, 1957; De Vries, 1958) for heat and water transfer within a two soil layer moisture model. The main simplification is that the water vapour flux is set to zero except at the ground surface. A vadose zone at "field capacity" is used as a simple boundary condition at the base of the soil through which excess water can drain to a water table at indefinite depth. Water reaching the ground surface from above is divided between infiltration and runoff. The roots extract water from both soil layers and the canopy releases this water into the atmopshere through transpiration (Fig. 3 shows the basic structure of the hydrological component of BEST). BEST simulates soil moisture in both liquid and frozen phases and accounts for the energy redistributed during phase changes.

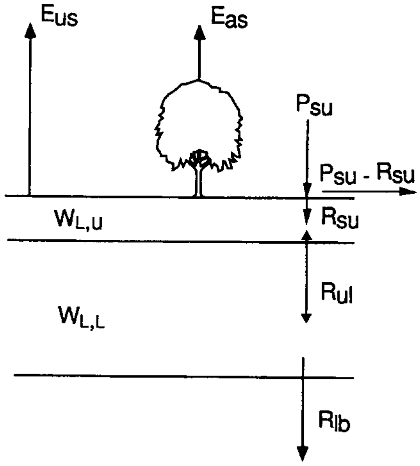

Fig. 3. Diagram showing the principal flows and stored of water in BEST. All symbols are defined in the text.

The capillary rise of soil water, R_{ul}, is parameterised according to Darcy's law, despite the limitations of Darcy's Law at scales above the microscale. We assume that the microscale heterogeneities which limit the value of Darcy's Law at, say 100 m^2, do not dominate moisture redistribution. We therefore write R_{ul} as

$$R_{ul} = K_{Hu}\left[1 - \left(\frac{\Delta\psi}{\Delta z}\right)_u\right] \tag{8}$$

where ψ is the total moisture potential, z is depth and the hydraulic conductivity in the upper soil layer (K_{Hu}) is expressed following Clapp and Hornberger (1978) as

$$K_{Hu} = K_{HO}\left(\frac{W_{L,u} + W_{L,l}}{2}\right)^{2B+3} \tag{9}$$

where $W_{L,l}$ is the volumetric soil moisture in the lower soil layer. The total moisture potential (ψ) combines gravity and water pressure forces. The water pressure force has a variable force in the vertical and under saturated soil with free drainage with steady infiltration at the surface, water flows downward at a positive velocity K_H. The soil properties (K_{HO}, B and X_v) are a function of soil type (Table 2), but the variation in ψ is assumed to be independent of soil type. Drainage from the bottom of the second layer, (R_{lb}), is calculated in a similar fashion but with the underlying soil kept at field capacity (i.e. assumed to be the water table).

Surface runoff is of considerable importance since water which forms overland flow is lost (instantaneously) and cannot be evaporated. Precipitation, leaf drip and snow melt is divided between surface runoff and infiltration. Following Eagleson (1970) a maximum infiltration rate is defined by integrating over a time step (∂t) and averaging

$$\frac{1}{\rho_w} R_{su} = K_{HO} - 2X_v(W_{L,u} - 1)\sqrt{1\frac{D_{HO}}{(\pi \partial t)}} \quad (10)$$

where all terms have been previously defined except the saturation hydraulic diffusivity (D_{HO}) which is

$$D_{HO} = \frac{-B\Psi_0 K_{HO}}{X_v} \quad (11)$$

Eq. (10) rests on a solution of the diffusion equation for a semi-infinite medium forced by a steady flow at the soil surface. The solution is reasonable if infiltrating water travels a short distance in the time span ∂t, compared to the depth of the soil layer. The definition of R_{su} in BEST also includes a simple method of accounting for sub grid scale soil moisture heterogeneity, where a fraction of the soil (A_{wet}) is considered to be saturated preventing infiltration:

$$A_{wet} = \min\left(\frac{W_{L,u}}{W_{fc}}, 1\right) \quad (12)$$

Once R_{su} is defined, the surface runoff rate, R_{us}, can be defined simply as

$$R_{us} = P_{su} - R_{su} \quad (13)$$

where P_{su} is the rainfall rate incident at the surface.

7. Soil temperature

BEST uses an extended version of the force-restore method for heat conduction (Deardorff, 1977, 1978) and also allows for the heat carried by infiltrating rain and for the energy involved in melting and freezing of soil water. The force restore model used in BEST differs from the standard force restore by calculating soil temperatures at three depths instead of the more usual two. This permits a more realistic representation of the diurnal, seasonal and inter-annual temperature waves. The surface soil layer temperature is subject to diurnally conductive forcing, the second soil temperature is forced by about a third of that of the conductive forcing for the ground temperature and the bottom level temperature is influenced by the seasonal forcing.

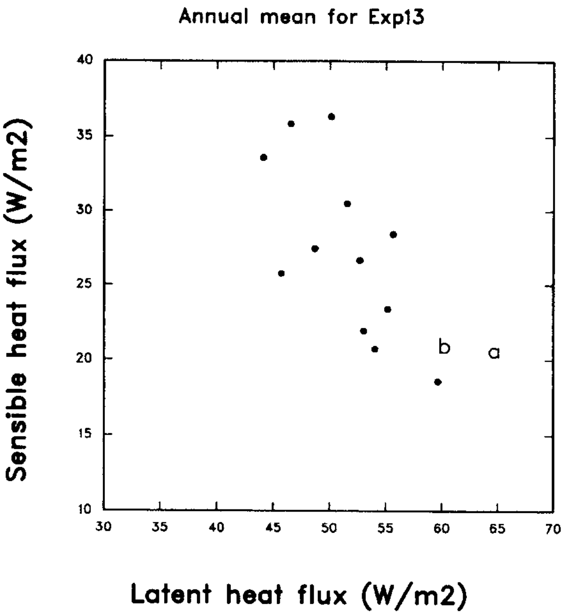

Fig. 4. Mean annual simulation of sensible and latent heat flux simulated by land surface schemes participating in the workshop (●) and for the two simulations with BEST [(a) refers to BEST incorporating the Eagleson (1970) β parameterisation and (b) refers to BEST using Eq. (14)].

8. HAPEX–MOBILHY experiments

As part of the workshop held at Macquarie University, a series of simulations were performed using the HAPEX–MOBILHY data (see Shao and Henderson-Sellers, 1996-this issue). In this section we briefly review the key characteristics of BEST's performance in these experiments and describe how this was subsequently improved by a simple parameterisation modification.

Although there was a wide range in the simulated evaporation for HAPEX–MOBILHY between models (550–816 mm yr^{-1}), the control simulation by BEST, in common with the results from some earlier experiments, was characterised by the simulation of excess evaporation (816 mm yr^{-1}, Fig. 4). BEST

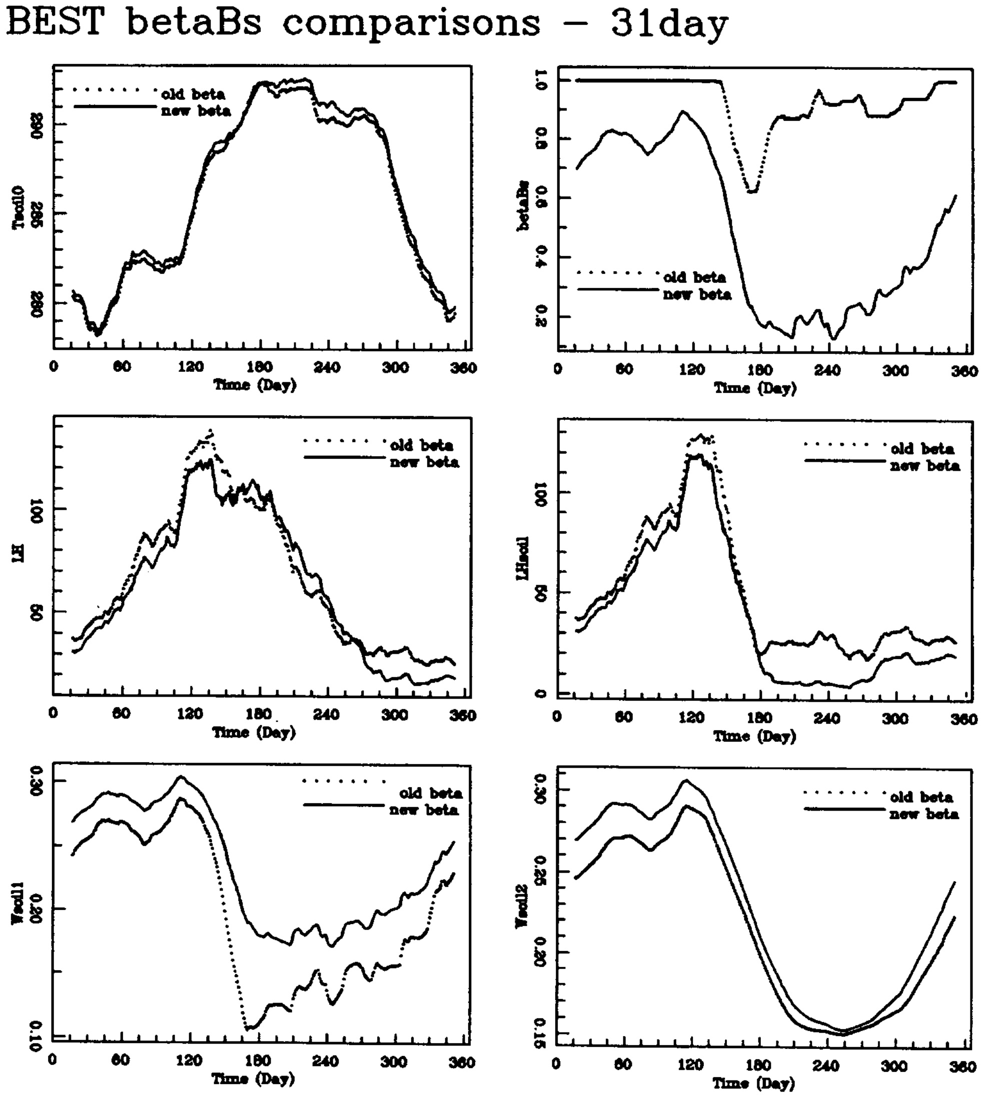

Fig. 5. Effect of changing the parameterisation of β in BEST from a complex method based on Eagleson (1970) ("old beta") to a simpler method ("new beta"). (a) soil temperature (K). (b) Soil β value. (c) Latent heat flux (W m^{-2}). (d) Bare soil latent heat flux (W m^{-2}). (e) Soil moisture concentration in the top 0.1 m and (f) soil moisture concentration in the layer 0.1–1.6 m (both soil moistures are expressed as fractions of saturation).

also simulated a relatively low surface temperature and a relatively low soil moisture.

The sensitivity of BEST to the parameterisation of soil wetness (the β parameterisation) was examined. The complex method incorporated in BEST (Eq. 10) based on Eagleson (1970) was replaced by a much simpler parameterisation where β_u decreases linearly from field capacity (W_{fc}) to wilting point (W_{wilt}):

$$\beta_u = \frac{W_{L,u} - W_{wilt}}{W_{fc} - W_{wilt}} \tag{14}$$

The impact of this change was to reduce the total evaporation from 816 mm yr^{-1} to 759 mm yr^{-1}, Fig. 4) which gave results more in line with the simulations by the other models. The annual bare soil evaporation was reduced from 618 mm yr^{-1} to 478 mm yr^{-1}. The impact of the change in the β parameterisation is more clearly shown in Fig. 5 which illustrates the seasonal patterns in a number of key variables. The β parameter is substantially lower (Fig. 5b), reaching a minimum of 0.2 compared to 0.7. This leads to a substantial reduction in bare soil evaporation (Fig. 5d) throughout the year, and particularly during those periods when the soil is most dry (May to September). The soil moisture is significantly higher than when the control version of the β parameter was used (Figs. 5e and 4f). The cold bias in BEST's soil temperature is also reduced (Fig. 5a).

The results presented here should not be used to reflect on the suitability of Eq. (14) in the prediction of soil wetness, but should instead be interpreted as an indication of the sensitivity of BEST to the formulation of bare soil evaporation. These results imply that at least some of the reasons for BEST's overestimate of evaporation is due to the use of Eq. (5) to simulate the soil moisture control on evaporation. The more complex method does not appear to lead to an improved simulation, although more comparisons for a wider variety of surface types will need to be conducted before this can be firmly concluded.

9. Summary and conclusions

BEST, a complex land surface scheme which includes two soil layers and an explicit canopy parameterisation has been briefly outlined. In a series of simulations conducted as part of a workshop held at Macquarie University the performance of the model, in comparison with a series of other land surface schemes was determined. It was noted that the basic version of the model tended to overestimate evaporation and underestimate temperature. This problem was explored and the method used to simulate the surface control on soil evaporation (the β parameter) was identified as a primary reason for the overestimation of evaporation compared to the other models. By replacing the complex parameterisation based on Eagleson (1970) by a simple linear formulation, BEST's prediction of evaporation became more in agreement with the other schemes. However, since these experiments were only conducted for a single surface type we do not conclude that the Eagleson (1970) methodology is flawed, rather we believe it requires more detailed investigation since BATS, which also includes an analogous method for calculating β did not simulate excessive evaporation. In addition, we note that the lack of suitable data sets to validate the bare soil component of land surface schemes (bare soil evaporation was not measured as part of HAPEX–MOBILHY) limits our ability to determine whether BEST's bare soil evaporation was high with respect to observations, or simply high compared to other land surface models.

Acknowledgements

AJP wishes to thank the Australian Research Council and the Macquarie University Research Grants Scheme for financial support for this project. CED wishes to acknowledge the support of an Australian Postgraduate Research Award. We would like to thank an anonymous reviewer for their very helpful comments.

References

Andre, J.C. and Blondin, C., 1986. On the effective roughness length for use in numerical three-dimensional models. Bound. Layer Meteorol., 35: 231–245.

Brutsaert, W., 1982. Evaporation into the Atmosphere. Theory History and Application. Reidel, Dordrecht, 299 pp.

Clapp, R.B. and Hornberger, G.M., 1978. Empirical equations for some soil hydraulic properties. Water Resour. Res., 14: 601–604.

Cogley, J.G., Pitman, A.J. and Henderson-Sellers, A., 1990. A land surface for large scale climate models. Trent Univ. Tech. Note, 90-1.

Deardorff, J.W., 1977. A parameterization of ground-surface moisture content for use in atmosphere prediction models. J. Appl. Meteorol., 16: 1182–1185.

Deardorff, J.W., 1978. Efficient prediction of ground surface temperature and moisture with inclusion of a layer of vegetation. J. Geophys. Res., 83: 1889–1903.

De Vries, D.A., 1958. Simultaneous transfer of heat and moisture in porous media. Trans. Am. Geophys. Union, 39: 909–916.

Dickinson, R.E., 1984. Modelling evapotranspiration for three dimensional global climate models. In: J.E. Hansen and T. Takahashi (Editors), Climate Processes and Climate Sensitivity. (Geophys. Monogr., 29.) Am. Geophys. Union, Washington, 5: 58–72.

Dickinson, R.E., Henderson-Sellers, A., Kennedy, P.J. and Wilson, M.F., 1986. Biosphere Atmosphere Transfer Scheme (BATS) for the NCAR Community Climate Model. NCAR Tech. Note TN275 + STR, 69 pp.

Eagleson, P.S., 1970. Dynamic Hydrology. McGraw-Hill, New York, 462 pp.

Gates, D.M., 1980. Biophysical Ecology. Springer, New York, 611 pp.

Hansen, J.E., Russell, G., Rind, D., Stone, P.H., Lacis, A.A., Lebedeff, S., Ruedy, R. and Travis, L., 1983. Efficient Three Dimensional Global Models for Climate Studies: Models I and II. Mon. Weather Rev., 111: 609–662.

Henderson-Sellers, A. and Wilson, M.F., 1983. Surface albedo data for climatic modeling. Rev. Geophys. Space Phys., 21: 1743–1778.

Jarvis, P.G., 1976. The interpretation of the variations in leaf water potential and stomatal conductance found in canopies in the field. Philos. Trans. R. Soc. London, Ser. B, 273: 593–610.

Mason, P.J., 1988. The formation of areally-averaged roughness lengths. Q. J. R. Meteorol. Soc., 114: 399–420.

Oke, T.R., 1978. Boundary Layer Climates. Wiley, New York, 372 pp.

Philip, J.R. and De Vries, D.A., 1957. Moisture movement in porous materials under temperature gradients. Trans. Am. Geophys. Union, 38: 222–232.

Pitman, A.J., 1988. The development and implementation of a new land surface scheme for use in general circulation models. Thesis, Liverpool Univ., 481 pp. (unpublished).

Pitman, A.J., Yang, Z.-L., Cogley, J.G. and Henderson-Sellers, A., 1991. Description of bare essentials of surface transfer for the Bureau of Meteorology Research Centre AGCM. BMRC Res. Rep., 32, 117 pp.

Shao, Y. and Henderson-Sellers, A., 1996. Validation of soil moisture simulation in land surface parameterisation schemes with HAPEX data. Global Planet. Change, 13: 11–46.

Sud, Y.C., Shukla, J. and Mintz, Y., 1988. Influence of land surface roughness on atmospheric circulation and precipitation: A sensitivity study with a general circulation model. J. Appl. Meteorol., 27: 1036–1054.

Taylor, P.A., 1987. Comments and further analysis on effective roughness lengths for use in numerical three-dimensional models. Bound. Layer Meteorol., 39: 403–418.

Wilson, M.F. and Henderson-Sellers, A., 1985. A global archive of land cover and soil data sets for use in general circulation climate models. J. Climatol., 5: 119–143.

Yang, Z.-L., Pitman, A.J., McAvaney, B. and Henderson-Sellers, A., 1995. The impact of implementing the Bare Essentials of Surface Transfer (BEST) land surface scheme into the BMRC GCM. Climate Dyn., 11: 279–297.