ELSEVIER

Global and Planetary Change 13 (1996) 47–56

GLOBAL AND PLANETARY CHANGE

Analysis of the relationship between bare soil evaporation and soil moisture simulated by 13 land surface schemes for a simple non-vegetated site

C.E. Desborough [a,*], A.J. Pitman [a], P. Irannejad [b]

[a] School of Earth Science, Macquarie University, North Ryde, 2109, NSW, Australia
[b] Climatic Impacts Centre, School of Earth Science, Macquarie University, North Ryde, 2109, NSW, Australia

Received 5 May 1995; accepted 28 August 1995

Abstract

Atmospheric and land surface data collected from the HAPEX–MOBILHY field experiment were used to compare the bare soil evaporation simulations of 13 land surface schemes and to examine the relationship between differences in evaporation and differences in soil moisture. For a 120-day period in which there was no vegetation present, computed total evaporation ranged between 100 and 250 mm. This large range in evaporation was not related to soil moisture differences. Prescribing surface soil moisture and temperature did not reduce the range in evaporation and instead the range was increased. The models' predictions of evaporation were in closer agreement with each other when they were allowed to select their own surface conditions than when they were forced to use the same conditions. The bare soil evaporation formulations used by the land-surface schemes are not consistent with each other and these inconsistencies produce widely-varying bare soil evaporation rates. The range in bare soil evaporation is unlikely to be reduced by improving the simulation of soil moisture and instead an assessment of why the bare soil evaporation formulations are inconsistent is required.

1. Introduction

The simulation of land surface processes in GCMs began with the simulation of bare soil processes (Manabe, 1969). These simple models were extended to include variable albedo and roughness, and sometimes included an attempt to parameterise the effects of vegetation tapping soil moisture via roots (e.g., Hansen et al., 1983). The trend in land surface modelling has been towards increasing complexity both above and below the ground–air interface, resulting in schemes such as the Biosphere–Atmosphere Transfer Scheme (BATS; Dickinson et al., 1986, Dickinson et al., 1993) and the Simple Biosphere (SiB; Sellers et al., 1986) which include an explicit canopy parameterisation. Simple bare soil models (e.g. Robock et al., 1994) are still used today and have been shown to perform consistently with the more advanced schemes on an annual average (Pitman et al., 1993).

Regardless of their complexity, all land surface

* Corresponding author. Phone: +61 2 850 8433, fax +61 2 850 8428, e-mail carl@cic.mq.edu.au.

SSDI 0921-8181(95)00036-4

schemes can simulate evaporation from barren regions and, in addition, most land surface schemes permit evaporation from the soil beneath the canopy or from a fraction of the host-model grid square which is not vegetated. The ability of land surface models to simulate bare soil processes is therefore important. As bare soil processes are inherently simpler to simulate than the coupled bare soil–canopy processes, it might be expected that a set of land surface schemes, carefully initialised and parameterised, should be able to agree on bare soil evaporation.

This paper draws on results from the RICE and PILPS soil moisture workshop held at Macquarie University in November 1994 (Shao et al., 1995). Data from the HAPEX–MOBILHY field experiment at Caumont (Goutorbe and Tarrieu, 1992) were used to compare the simulation of soil moisture processes by 13 land surface schemes for a soya crop. The workshop aimed to investigate the differences between the schemes' simulations of soil moisture, evaporation, transpiration, runoff and drainage. In this paper, we make use of results from a 120-day period in which there was no vegetation to quantify differences between the simulations of bare soil evaporation and to examine the relationships between those differences and the corresponding differences in soil moisture.

2. Modelling bare soil evaporation

The 13 land surface schemes used in this study all have the same basic design in that they model how the land surface evolves in time given a description of the land surface and atmospheric forcing data. They are however, designed for a variety of applications and consequently, they are of varying levels of complexity and emphasise different aspects of the land surface. In this section, we describe the bare soil evaporation formulations of the models. More detailed descriptions of the models can be found in Shao et al. (1995) and in the following papers: BATS, BEST (Pitman et al., 1991; Pitman and Desborough, 1996), BGC (Running and Hunt, 1993), BIOME2 (Haxeltine et al., 1994), BUCKET (Manabe, 1969), CENTURY (Parton et al., 1993), CLASS (Verseghy et al., 1993), CSIRO9 (Kowalczyk et al., 1991), ISBA (Noihlan and Planton, 1989), PLACE (Wetzel and Boone, 1995), SECHIBA2 (Ducoudre et al., 1993), SSiB (Xue et al., 1991) and VIC (Liang et al., 1994).

2.1. Potential evaporation

Three potential evaporation formulations are used by the models discussed in this paper. BATS, BEST, BUCKET, ISBA, CLASS, PLACE, SSiB and SECHIBA2 use the aerodynamic formulation; CSIRO9, BGC, VIC and CENTURY use the Penman-Monteith formulation (Monteith, 1965); the Priestley-Taylor formulation (Priestley and Taylor, 1972) is used by BIOME2. All three methods are combination methods (Milly, 1991) in that evaporation is calculated by combining aerodynamic expressions for heat and moisture fluxes with the surface energy balance equation:

$$R_n = H + LE + G \tag{1}$$

where R_n is net radiation (W m^{-2}), H is the sensible heat flux (W m^{-2}), L is the latent heat of vaporisation (J kg^{-1}), E is the evaporation rate (kg m^{-2} s^{-1}) and G is the ground heat flux (W m^{-2}). The aerodynamic expressions for potential evaporation (E_p) and sensible heat (H) are:

$$E_p = \rho\left(q^*(T_g) - q_a\right)/r_{aV} \tag{2}$$

$$H = \rho c_p(T_g - T_a)/r_{aH} \tag{3}$$

where q is specific humidity (with * indicating the saturated value) and T temperature (the subscripts g and a refer to the ground surface and the air respectively); r_{aH} and r_{aV} are the aerodynamic resistances to heat and water vapour respectively; ρ is the density of air and c_p the specific heat of air at constant pressure. The potential evaporation rate predicted by Eq. (2) is the rate that would occur if water were freely available at the soil surface (it should be noted that there is some dispute over whether T_g is the appropriate temperature for calculation of potential evaporation: see Milly, 1992). There are several empirical methods which can be used to reduce potential evaporation to actual evaporation and these are discussed later in this section.

The non-linear dependence of q^* on temperature means that the system of Eqs. (1), (2) and (3) (with some modification to Eq. (2) so that it gives actual evaporation) can only be solved either by iteration (the aerodynamic approach) or by including an additional equation to describe the temperature-dependence of q^*. Equations based on the second approach are called combination equations (Milly, 1991). Penman (1948) produced a simple combination equation by approximating the humidity gradient in Eq. (2) by a linear expression involving the gradient (s) of the tangent to the saturation vapour pressure curve evaluated at the air temperature (T_a). Monteith (1965) added a surface resistance (r_s) to account for the physiological control exerted by plants over evaporation to produce a combination equation that could be applied to vegetated surfaces (the Penman-Monteith equation):

$$E_p^* = \frac{s(R_n - G) + \rho c_p d_a / r_{aV}}{s + \gamma(1 + r_s/r_{aV})} \tag{4}$$

where d_a is the vapour pressure deficit of the air above the surface and γ is the psychrometeric constant. It is assumed that the aerodynamic resistances to heat (r_{aH}) and water vapour (r_{aV}) are identical. The * was added to E_p to indicate that it is not, strictly speaking, a potential evaporation in that it includes a surface resistance. In the simple non-vegetated case considered in this paper, r_s is equal to zero and Eq. (4) predicts potential evaporation. The Priestley-Taylor formulation is a simple combination equation in which equilibrium evaporation is multiplied by a factor of 1.26 (α);

$$E_p^* = \alpha \frac{s(R_n - G)}{s + \gamma(1 + r_s/r_{aV})} \tag{5}$$

The value of α has been shown experimentally to be extremely variable (Monteith and Unsworth, 1990).

Milly (1991) points out that equations such as Penman-Monteith and Priestley-Taylor which use Penman's linear approximation of the humidity gradient consistently underestimate evaporation, sometimes by a substantial amount. He presents a series of simple higher-order combination equations which could be used to obtain a more accurate estimate of the evaporation rate. None of the land surface schemes in this paper used the equations of Milly or the comparable equations of Paw and Gao (1988).

As well as using different basic formulations for potential evaporation, the models also use a variety of different methods to reduce the potential evaporation rate to the actual rate.

2.2. Parameterising the dependence of bare soil evaporation on soil moisture

Water in the soil is not freely available for evaporation and the amount of water which can be lost through the soil surface is dependent on the atmospheric demand (potential evaporation) and the hydraulic transfer within the soil. In order to simulate the highly non-linear nature of the soil hydraulic properties (see Mahrt and Pan, 1984) the vertical resolution of the land surface scheme must be very high, with depths of order 1 mm near the surface. This is not possible in land surface schemes designed for General Circulation Models (GCMs) as such a thin soil layer would require a very short timestep and this is computationally expensive. The need for computational efficiency and the lack of detailed soil information on a global scale, limits the number of soil layers to between one and three layers in most cases.

As discussed by Kondo and Saigusa (1990) and Mahfouf and Noilhan (1991), simple methods for relating bare soil evaporation to soil moisture are required. There are three methods used in land surface models at present: the α, β and γ methods. The α and β methods use a parameter to represent the availability of soil moisture for evaporation, while the γ method limits evaporation by the maximum rate at which soil moisture can be supplied to the evaporating surface. Kondo and Saigusa (1990) point out that while there are many ways to represent this surface moisture availability, it has not been established which methodology is generally superior. The methods used by the models discussed in this paper are summarised in Table 1.

In the α formulation, surface specific humidity is assumed to be proportional to saturation specific humidity and bare soil evaporation (E_g) is parameterised as:

$$E_g = \rho\left(\alpha q^*(T_g) - q_a\right)/r_{aV} \tag{6}$$

Table 1
Bare soil evaporation formulations used by the models showing the method used for calculating potential evaporation and the method used to incorporate the dependence of evaporation on soil moisture stress

Model	Timestep	Potential evaporation	Moisture stress
BATS	30 min	Aerodynamic	β and γ
BEST	30 min	Aerodynamic	β
BGC	1 day	Penman-Monteith	–
BIOME2	1 day	Priestley-Taylor	γ
BUCKET	30 min	Aerodynamic	β
CENTURY	1 day	Penman-Monteith	β
CLASS	30 min	Aerodynamic	α
CSIRO9	30 min	Penman-Monteith	β
ISBA	30 min	Aerodynamic	α
LAPS	30 min	Aerodynamic	α
PLACE	30 min	Aerodynamic	γ
SECHIBA2	30 min	Aerodynamic	β
SSiB	30 min	Aerodynamic	α
VIC	60 min	Penman-Monteith	β

Kondo and Saigusa (1990) state that α expresses the land surface moisture availability as the air relative humidity at the humidity roughness height but also point out that, rather than using the relative humidity at the humidity roughness height, the relative humidity of the air in the soil pore space is often used (calculated following Philip, 1957). In the β formulation, evaporation is parameterised as:

$$E_g = \rho\beta\left(q^*(T_g) - q_a\right)/r_{aV} \quad (7)$$

Mahfouf and Noilhan (1991) discuss the methods used to calculate the α and β parameters (see their table 1).

In the third type of approach, the γ method (supply and demand), evaporation occurs at the potential rate until the soil is no longer able to diffuse water to the evaporating surface at a high enough rate to supply the atmospheric demand. Evaporation is thus limited by either supply or demand and occurs at whichever rate is the smallest. The major drawback of this approach is the difficulty in parameterising the rate of moisture diffusion within the soil and specifying the soil properties upon which the diffusivity depends.

Given that three basic methods are used to calculate the potential soil evaporation (Penman-Monteith, Aerodynamic and Priestley-Taylor) and three methods are then used to reduce the potential soil evaporation to the actual evaporation (α, β and γ), it becomes increasingly likely that the actual evaporation rate will differ between models. The first 120 days of the simulations from the soil moisture workshop (Shao et al., 1995) were used to investigate the degree of consistency between the model simulations of bare soil evaporation. There was no vegetation cover during this period and thus the entire evaporation flux was through the soil–atmosphere interface.

3. Experimental design

This section describes the relevant experiments (Experiments 13 and 2a1) from the soil moisture workshop. A more detailed description of the workshop experiments is provided by Shao et al. (1995) and in the other papers contained in this special volume.

A number of control experiments were conducted before and during the soil moisture workshop with soil and vegetation properties being specified differently for each experiment. One of the control experiments, Experiment 13, was designated as the preferred control and it is used as the control in this paper as well. Appropriate soil and vegetation properties were carefully selected so as to ensure that differences between the simulations of the models were not due to differences in these properties. The models all started from the same initial conditions and were forced with the same observed atmospheric data, which was available at a 30 min interval for one continuous year. The models were run continuously against the forcing data for as many years as required to equilibrate with it and the equilibrium simulations were compared. In this paper we compare the models' simulations of surface soil moisture and bare soil evaporation for the first 120 days of an equilibrium year.

A further experiment was conducted (Experiment 2a1) to examine how the treatment of bare soil evaporation varied between the models. The models were run to equilibrium and then restarted from the beginning of an equilibrium year. They were run for 120 days with soil moisture and temperature in the top 10 cm of the soil prescribed from the control simulation of BATS.

4. Results

4.1. Results from Experiment 13

The equilibrium evaporation for Experiment 13 varies substantially between the models such that the range is 10–40 W m^{-2} in January and 50–110 W m^{-2} at the end of April (Fig. 1a). The increases in the mean evaporation and the range between the models follow the increase in net radiation (not shown here). The relative positions of the models within the range of evaporation rates remains quite constant throughout the period (if a model predicts evaporation at the top of the range in January, it remains at the top of the range in April). The total evaporation for the four months varies between 100 and 230 mm (Fig. 1b) which correspond to average fluxes of 25 and 55 W m^{-2}, respectively.

There are two possible explanations for why the models produce such widely varying evaporation

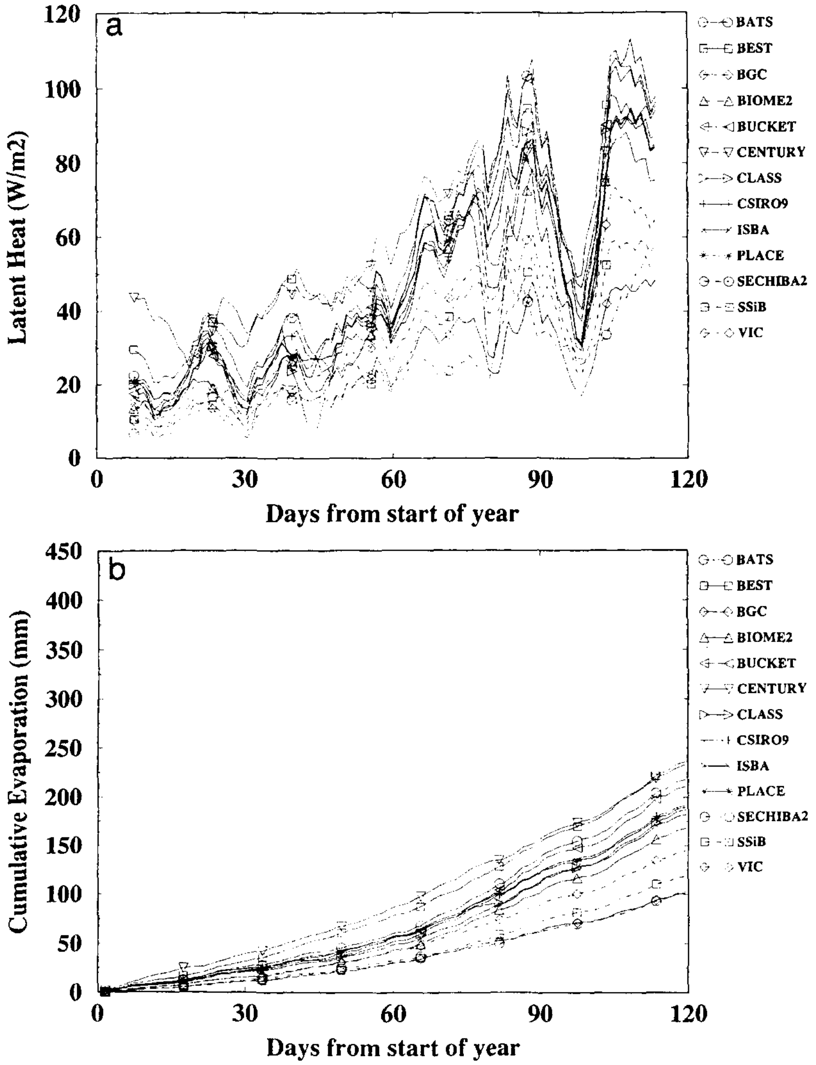

Fig. 1. Evaporation for the first 120 days of an equilibrium year in the control experiment (Experiment 13). (a) 7-day running mean; (b) Cumulative.

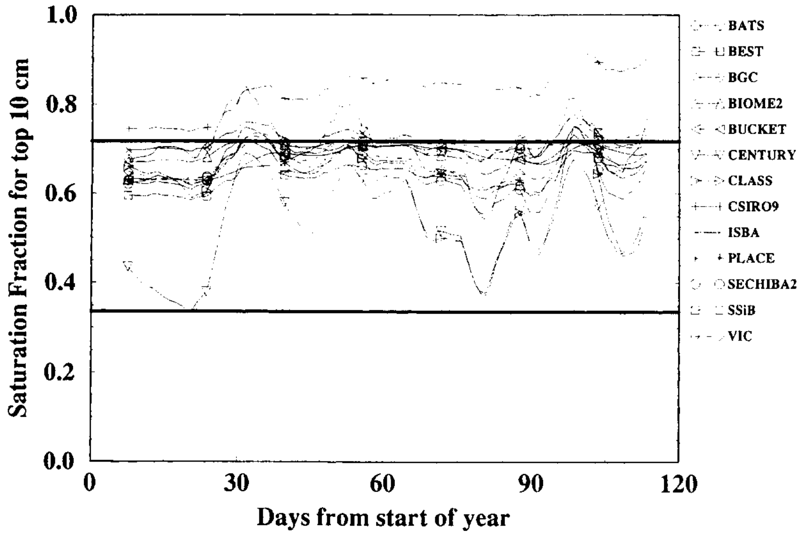

Fig. 2. Soil moisture (saturation fraction) in the top 10 cm of the soil for the first 120 days of an equilibrium year in the control experiment (Experiment 13).

rates given the same land surface characteristics and the same atmospheric forcing. The differences could result from the use of inconsistent bare soil evaporation methodologies or they could result from differences in soil moisture simulated by the models. The models are run from the start of an equilibrium year, so they will have different initial soil moistures. Even given the same initial soil moistures, differing amounts of runoff and drainage during the period could lead to the models having different soil moistures. Thus, the differences could be due to differences in other parts of the models rather than being due to differences in their treatment of bare soil evaporation.

Fig. 2 shows how soil moisture in the top 10 cm of the soil column varies over the 120 day period, where soil moisture is measured in terms of saturation fraction (the ratio of soil moisture to pore space). The thick lines represent wilting point (0.33) and field capacity (0.72). The field capacity is the

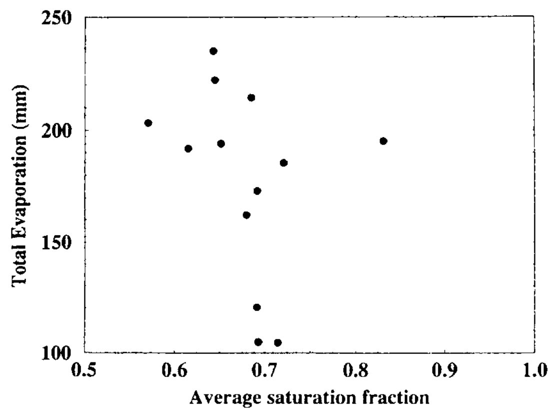

Fig. 3. Dependence of total evaporation on average soil moisture for the first 120 days of the control experiment (Experiment 13) for each model.

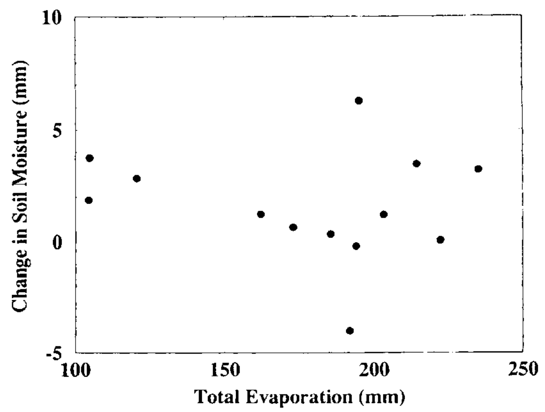

Fig. 4. Dependence of change in soil moisture on total evaporation for the first 120 days of the control experiment (Experiment 13) for each model.

maximum amount of moisture that the soil can hold after free drainage has occurred. The wilting point has no real significance for bare soil evaporation, but is included as a guide. The majority of the models start the equilibrium year with soil moisture between 0.6 and 0.7 and remain in that range for most of the 120 day period. There are three exceptions: CSIRO9 starts with soil moisture just above field capacity and increases throughout the period; CENTURY begins with a much lower value of soil moisture and varies between wilting point and field capacity; CLASS simulates evaporation within the consensus range for the first 60 days and follows the CENTURY values for the remainder of the period.

Fig. 2 shows that the majority of the models have surface soil moistures lying in a narrow range near field capacity. Fig. 3 shows that there is no relationship between total evaporation and average surface soil moisture in the 120 day period. The large differences in the evaporation rates simulated by the models (Fig. 1) cannot be explained in terms of soil moisture differences. This implies that the differences are due to difference in the bare soil evaporation formulations of the models.

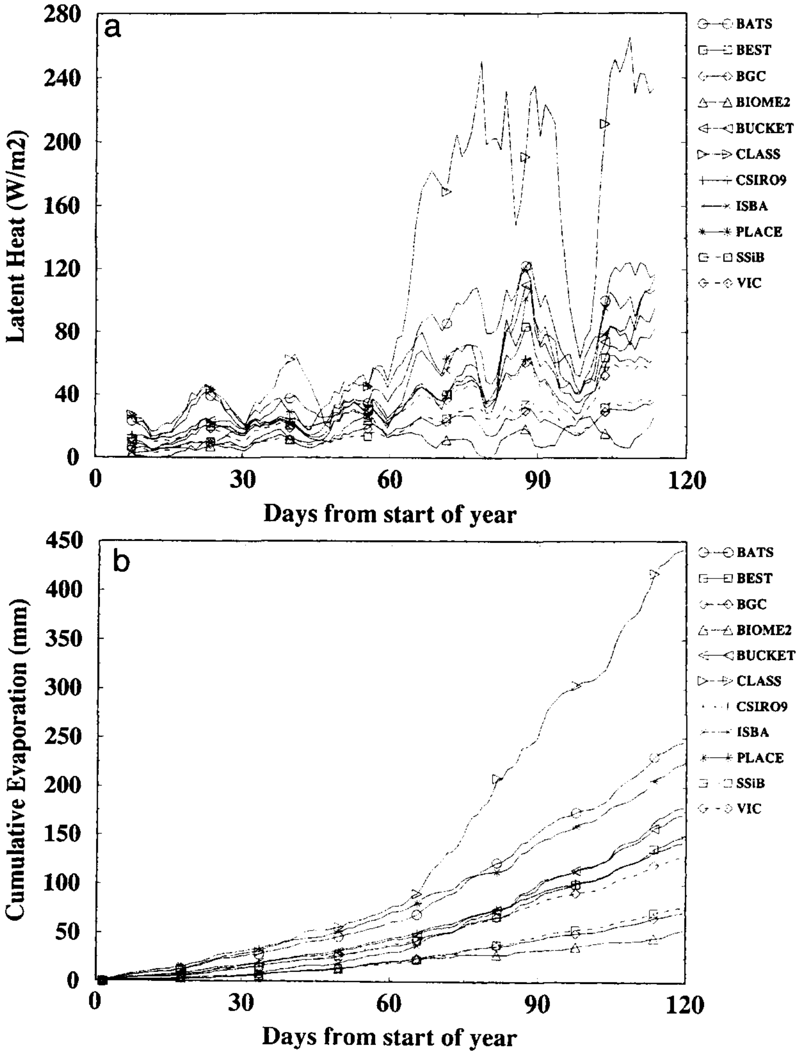

Fig. 5. As Fig. 1 but from Experiment 2a1.

The large differences in evaporation have little effect on soil moisture. The total evaporation for the 120 days varies from 100 to 230 mm, while corresponding changes in soil moisture are negligible for most models. There is no relationship between the change in soil moisture and the total evaporation and the maximum change is only 7 mm (Fig. 4). Given that the precipitation is prescribed, this means that the sum of surface runoff and drainage to deeper soil must also vary by 130 mm. This is not a surprising result, given that the models have soil moistures near field capacity.

4.2. Results from Experiment 2a1

The findings from Experiment 13 are supported by results from Experiment 2a1 in which the surface conditions (soil moisture and temperature in the top 10 cm) were prescribed. If the differences in evaporation (Fig. 1) were due to differences in soil moisture, specifying the surface conditions should decrease the range in evaporation. This decrease was not observed in Experiment 2a1 (Fig. 5). This supports the claim that differences in evaporation cannot be explained by differences in soil moisture.

Furthermore, the range of evaporation in Experiment 2a1 (Fig. 5) is actually larger than the range in Experiment 13 (Fig. 1). Excluding CLASS (which produces 450 mm of evaporation), the total evaporation varies from 50 to 250 mm for Experiment 2a1 (Fig. 5b) compared to a range of 100 to 230 mm for Experiment 13 (Fig. 1b). The level of agreement between the models is higher when they are allowed to predict their own surface conditions than when they are forced to use prescribed conditions. This implies that if all the land surface schemes were forced with the *observed* soil moistures and temperatures, the range in answers would be larger than when permitted to use values which match the schemes evaporation formulations.

These findings are most clearly illustrated by CLASS. When it is allowed to calculate soil moisture internally (Experiment 13), CLASS simulates evaporation which is well within the range predicted by the other models (Fig. 1). When forced to use the BATS soil moisture and temperature, CLASS simulates almost double the next highest evaporation in April and May. This does not reflect badly on CLASS, but it does demonstrate the danger of interfering with the internal consistency of land surface schemes.

5. Summary and implications

These conclusions only apply to the situation where soil moisture is in plentiful supply and further exploration is required to determine the behaviour of the schemes under moisture stress. However, we conclude from the experiments conducted here that:

1. 13 schemes produced very different bare soil evaporation rates, with total evaporation for the 120 days ranging from 100 to 230 mm;
2. differences between the models simulation of bare soil evaporation cannot be explained by variations in soil moisture and are instead due to differences in the methodologies used to calculate bare soil evaporation;
3. models perform better (the range of evaporations is smaller) when allowed to equilibrate to their own surface conditions (soil moisture and temperature) than when the surface conditions are prescribed;
4. change in soil moisture over the 120 days is largely independent of the evaporation rate.

The implications of these results are:

1. Inconsistencies between the bare soil evaporation formulations used by land surface schemes lead them to produce very different evaporation rates for the same conditions. More work is required to determine why 13 land surface schemes produce such a large range in bare soil evaporation for a site with no vegetation, prescribed soil properties and identical atmospheric forcing.
2. The correct simulation of evaporation and runoff requires changes in soil moisture to be correct and does not necessarily require the absolute values of soil moisture to be correct. Specifying soil moisture correctly may not provide improved simulations of fluxes such as evaporation and runoff. This implies that we require direct validation data for these fluxes rather than validation data for soil moisture.
3. Evaporation simulated by some models (e.g. CLASS) with prescribed surface conditions was

worse than the evaporation simulated by the models when the surface conditions were predicted internally. This demonstrates the danger of interfering with the internal consistency of land surface schemes and indicates that it would be extremely difficult (and unpredictable) to interchange sub-models between different land surface schemes.

Acknowledgements

The authors thank Yaping Shao and Ann Henderson-Sellers for organising the soil moisture workshop and the other workshop participants for their part in producing the results analysed in this paper; CED acknowledges the support of an Australian Postgraduate Research Award; AJP wishes to thank the Australian Research Council and the Macquarie University Research Grants Scheme for financial support for this project.

References

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.

Dickinson, R.E., Henderson-Sellers, A. and Kennedy, P.J., 1993. Biosphere–Atmosphere Transfer Scheme (BATS) Version 1e as coupled to the NCAR Community Climate Model. NCAR Tech. Note, TN383 + STR, 72 pp.

Ducoudre, N.I., Laval, K. and Perrier, A., 1993. SECHIBA, a new set of parameterizations of the hydrologic exchanges at the land–atmosphere interface within the LMD atmospheric general circulation model. J. Climate, 6: 248–273.

Goutorbe, J.P. and Tarrieu, C., 1992. HAPEX–MOBILHY data base. In: T. Schmugge and J.C. Andre (Editors), Land Surface Evaporation. Springer, Berlin, pp. 403–410.

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.

Haxeltine, A., Prentice, I.C. and Creswell, I.D., 1994. Use of a coupled carbon and water flux model to predict vegetation structure: An Australian case study (manuscript available from the author, Dep. Plant Ecol., Univ. Lund, Lund, Sweden).

Kondo, J. and Saigusa, N., 1990. A parameterization of evaporation from bare soil surfaces. J. Appl. Meteorol., 29: 385–398.

Kowalczyk, E.A., Garratt, J.R. and Krummel, P.B., 1991. A soil–canopy scheme for use in a numerical model of the atmosphere—1D stand-alone model. CSIRO, DAR, Tech. Pap., 23, 56 pp.

Liang, X., Lettenmaier, D.P., Wood, E.F. and Burges, S.J., 1994. A simple hydrologically based model of land surface water and energy fluxes for GCMs. J. Geophys. Res., 99(D7): 14,415–14,428.

Mahfouf, J.-F. and Noilhan, J., 1991. Comparative study of various formulations of evaporation from bare soil using in situ data. J. Appl. Meteorol., 30: 1354–1365.

Mahrt, L. and Pan, H., 1984. A two-layer model of soil hydrology. Bound. Layer Meteorol., 29: 1–20.

Manabe, S., 1969. Climate and the ocean circulation: 1. The atmospheric circulation and the hydrology of the earths' surface. Mon. Weather Rev., 97: 739–805.

Milly, P.C.D., 1991. A refinement of the combination equations for evaporation. Surv. Geophys., 12: 145–154.

Milly, P.C.D., 1992. Potential evaporation and soil moisture in general circulation models. J. Climate, 5: 209–226.

Monteith, J.L., 1965. Evaporation and environment, state and movement of water in living organisms. In: 19th Symp. Soc. Exp. Biol., pp. 205–234.

Monteith, J.L. and Unsworth, M.H., 1990. Principals of Environmental Physics. Edward Arnold, London, 2nd ed., 241 pp.

Noihlan, J. and Planton, S., 1989. A simple parameterisation of land surface processes for meteorological models. Mon. Weather Rev., 117: 536–549.

Parton, W.J., Scurlock, J.M.O., Ojima, D.S., Gilmanov, T.G., Scholes, R.J., Schimel, D.S., Kirchner, T., Menaut, J.-C., Seastedt, T., Garcia Moya, E., Kamnalrut, A. and Kinyamario, J.I., 1993. Observations and modelling of biomass and soil organic matter dynamics for the grassland biome worldwide. Global Biogeochem. Cycles, 7: 785–809.

Paw, U. and Gao, W., 1988. Applications of solutions to non-linear energy budget equations. Agric. Forest Meteorol., 43: 121–145.

Penman, H.L., 1948. Natural evaporation from open water, bare soil and grass. Proc. R. Soc. London, Ser. A, 193: 120–145.

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.

Pitman, A.J. and Desborough, C.E., 1996. Brief description of bare essentials of surface transfer and results from simulations with the HAPEX–MOBILHY data. Global Planet. Change, 13: 139–147.

Philip, J.R., 1957. Evaporation, and moisture and heat fields in the soil. J. Meteorol., 14: 354–366.

Priestley, C.H.B. and Taylor, R.J., 1972. On the assessment of surface heat flux and evaporation using large scale parameters. Mon. Weather Rev., 100: 81–92.

Robock, A., Vinnikov, K.Ya, Schlosser, C.A., Speranskaya, N.A. and Xue, Y., 1994. Use of Russian soil moisture and meteorological observations to validate soil moisture simulations with biosphere and bucket models. J. Climate, 8: 15–35.

Running, S.W. and Hunt Jr., E.R., 1993. Generalizations of a forest ecosystem process model for other biomes, BIOME-BGC, and an application for global-scale models. In: J.R.

Ehrlinger and C. Fields (Editors), Scaling Physiological Processes: Leaf to Globe. Academic Press, New York.

Sellers, P.J., Mintz, Y., Sud, Y.C. and Dalcher, A., 1986. A Simple Biosphere model (SiB) for use within general circulation models. J. Atmos. Sci., 43: 505–531.

Shao, Y., Anne, R.D., Henderson-Sellers, A., Irannejad, P., Thornton, P., Liang, X., Chen, T.H., Ciret, C., Desborough, C.E., Balachova, O., Haxeltine, A. and Ducharne, A., 1995. Soil moisture simulation—a report on the RICE and PILPS workshop. GEWEX Tech. Note, 14, 179 pp.

Verseghy, D.L., Mcfarlane, N.A. and Lazare, M., 1993. CLASS—a Canadian land surface scheme for GCMs, II. Vegetation and coupled runs. Int. J. Climatol., 13: 347–370.

Wetzel, P.J. and Boone, A., 1995. A parameterization for land–atmosphere–cloud exchange (PLACE): Documentation and testing of a detailed process model of the partly cloudy boundary layer over heterogeneous land. J. Climate, 8: 1810–1837.

Xue, Y., Sellers, P.J., Kinter, L.J. and Shukla, J., 1991. A simplified biosphere model for global climate studies. J. Climate, 4: 345–364.