Soil–Vegetation–Atmosphere Transfer Schemes and Large-Scale Hydrological Models (Proceedings of a symposium held during the Sixth IAHS Scientific Assembly at Maastricht, The Netherlands, July 2001). IAHS Publ. no. 270. 2001.

Relative influence of vertical and horizontal processes in large-scale water and energy balance modelling

ELEANOR BLYTH

Centre for Ecology and Hydrology, Wallingford, Oxfordshire OX10 8BB, UK

e-mail: emb@ceh.ac.uk

Abstract The role of large-scale land-atmosphere transfer schemes is to partition the massive radiative energy from the sun into evaporation and sensible heat. In order to do so, the models need to trace where the radiation energy is being absorbed and how much water is available at the point of absorption. Therefore, great emphasis has been placed on the vertical structure of the land surface and on the vertical movement of water through the soil and the vegetation system. As part of this tracking of the vertical movement of water, there is a necessity to include some representation of the horizontal flow of water. If this is ignored, the land is assumed to be flat and homogeneous, and no surface runoff will be generated, resulting in modelled soil moisture, and hence evaporation, that is too high. This is partly why hydrological models are more commonly being embedded in meteorological land surface schemes. The other reason is that the runoff itself is now required for the purposes of predicting water resources, freshwater flows into the oceans and for model validation using river flow. There is a need to assess how complex these hydrological models need to be. On the one hand, they need to be complex enough to model the processes accurately. On the other hand, the accuracy of the model needs to be weighed against the overhead of obtaining parameters on a global scale. This paper attempts to address this issue by comparing the sensitivity of the desired output (long-term water and energy balance) to the parameters in the horizontal and vertical models. The land surface scheme of the Met Office (Met Office Surface Energy Scheme—MOSES) is used with a standard rainfall–runoff model, Probability Distribution Model (PDM) embedded in the top soil layer. The parameters of the vertical process model are varied from a coarse soil to a fine soil. The parameter for the horizontal process model is varied from a flat homogeneous landscape to one with high variability. It is shown that the long-term water balance in this model is equally sensitive to parameters that describe the horizontal flow of water and the vertical flow of water.

Key words land surface schemes; water balance; hydrological model; soil parameters

INTRODUCTION

Land surface schemes in atmospheric models have been designed to reproduce the diurnal variation in fluxes of heat and water vapour to the atmosphere to represent the meteorological forcing from the surface. Many land surface schemes have multi-layered soil models (for a review, see Garratt, 1993), which give good results in terms of the hourly fluctuations of surface conditions. However, as the atmospheric models aim at longer-term forecasts and at predicting future climates, land surface schemes now also need to model the longer time-scale evolution of soil moisture (Beljaars *et al.*, 1996). In addition to the need for long-term estimates of evaporation, more realistic representations of the surface water balance are needed to meet the goal of including runoff in atmospheric models.

The water balance is still very inaccurate in many atmospheric model land surface schemes (Shao & Henderson-Sellers, 1996). It has been argued that the multi-layered model structure of most atmospheric model land surface schemes is not conducive to accurate water balance modelling. Typically, a water balance model would concentrate model detail on the horizontal heterogeneity rather than on the vertical processes (see Schultz *et al.* (1995) for a review of these issues). Statistical distributions of the spatial heterogeneity of runoff-generation properties of the landscape are usually used in large-scale hydrological models. For instance, Zhao *et al.* (1980) found that the cumulative density function, F, of the soil water storage capacity, C, could be described as follows (called the Xinanjiang model):

$$F = 1 - \left(1 - \frac{C}{C_m}\right)^B \tag{1}$$

where C_m is the maximum storage capacity and B is a tunable parameter. Several examples of the use of this distribution exist in the literature. Firstly Moore (1985) used the Xinanjiang distribution in the Probability Distribution Model (PDM). Then Wood *et al.* (1992) used it in the Variable Infiltration Capacity model (VIC) and Todini (1996) used it in the Arno model. The Arno model has been successfully applied in a general circulation model (GCM) (Dumenil & Todini, 1992) and the PDM is used routinely to do impact studies (e.g. Arnell & Reynard, 1993).

TOPMODEL (Beven & Kirkby, 1979) also applies a variation in soil moisture storage capacity. The variation in this case is a function of the topography. Famiglietti & Wood (1991) successfully applied this in a GCM. There are similarities between the two approaches. For instance, Sivapalan *et al.* (1997) has fitted a cumulative density function curve to several TOPMODEL distributions. Lohmann *et al.* (1998) studied the effect of using these different rainfall–runoff generation sub-models in land surface schemes on the overall water balance.

MODEL

The land surface scheme in the Unified Model (UM) of the Met Office (referred to as MOSES: Met Office Surface Energy Scheme) is described by Cox *et al.* (1998). Some details are included in this paper.

Soil moisture

MOSES has four soil layers, with layer depths of 0.1, 0.25, 0.65 and 2.0 m. The flow between the layers is calculated according to the Darcy-Richards equations:

$$Q = K\frac{\mathrm{d}}{\mathrm{d}z}[\psi + z] \tag{2a}$$

with

$$K = K_s\left(\frac{\theta}{\theta_s}\right)^{2b+3} \tag{2b}$$

$$\psi = \psi_s\left(\frac{\theta}{\theta_s}\right)^{-b} \tag{2c}$$

where K is the hydraulic conductivity (kg m^{-2} s^{-1}), ψ is the suction (m), θ is the volumetric soil water concentration, z is depth of soil (m), Q is the flow of water (kg m^{-2} s^{-1}) and the subscript s indicates the value at saturation.

Evaporation

Evaporation is controlled by the Penman-Monteith equation with a surface resistance that depends on photosynthetic rate. The root-depth averaged soil moisture, θ, reduces the evaporation by a factor, β, which varies linearly from the critical point, θ_c, down to the wilting point, θ_w, according to the following equation:

$$\beta = \left(\frac{\theta - \theta_w}{\theta_c - \theta_w}\right) \tag{3a}$$

$$\theta = \sum_{i=1}^{n} \mathrm{d}z_i\theta_i \tag{3b}$$

where n is number of layers over which roots lie. This implies a uniform root distribution down through the soil. The summation parameter, n, depends on the vegetation type: 3 for grass and crops, 4 for trees.

Runoff

Surface runoff is generated if the rainfall intensity is greater than the infiltration capacity of the soil, which is assumed to be equal to K_s, the saturated hydraulic conductivity. This rarely occurs in temperate zones. Saturation excess runoff is not modelled. If any layer is supersaturated, the water is routed to the layer below, or to drainage if it is the bottom layer. Subsurface flow is generated by gravity drainage at the bottom of the soil column according to the following equation:

$$Q_d = K_s\left(\frac{\theta_4}{\theta_s}\right)^{2b+3} \tag{4}$$

where Q_d is the subsurface flow (kg m^{-2} s^{-1}), and θ_4 is the soil moisture in the bottom layer (m^3 m^{-3}).

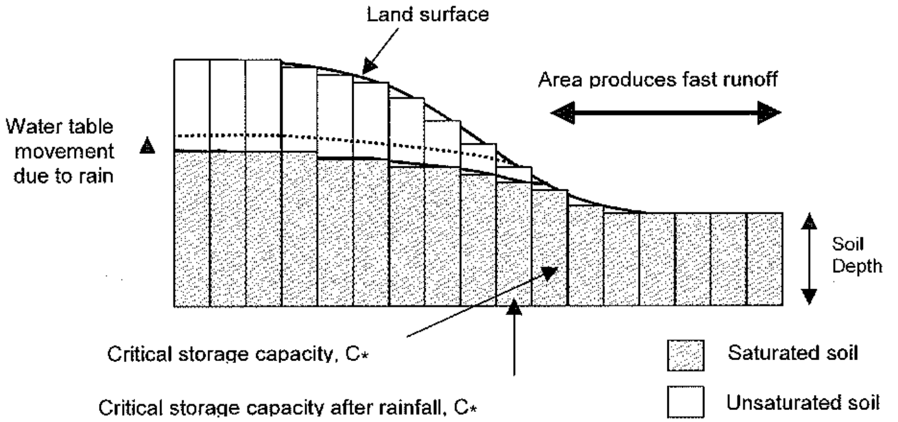

Fig. 1 Schematic diagram of the PDM model.

Rainfall–runoff model

The soil moisture distribution equation from the Probability Distribution Model (PDM) is coded into the top soil layer of MOSES. Figure 1 is a schematic representation of the method, which is used as follows. First, redefine the distribution in terms of the fractional area, A, that has a point storage capacity less than C:

$$C = C_m\left[1-(1-A)^{1/B}\right] \tag{5}$$

Also define the critical storage capacity C_* below which all stores are full.

$$C_* = C_m\left[1-\left(1-\frac{\theta}{\theta_s}\right)^{1/B+1}\right] \tag{6}$$

Integrating equation (1) from $C = 0$ to $C = C_m$ gives the relationship between C_m and θ_s as follows:

$$C_m = \theta_s(1+B) \tag{7}$$

If precipitation P is added to the area, the critical point rises by that amount as follows:

$$C'_* = C_* + \frac{P}{dz_1} \tag{8}$$

By inverting equation (6), it is possible to calculate the area storage as a function of the critical point. The area storage before and after rainfall can be quantified. Any excess between the rainfall and the change in storage must be due to runoff.

$$Q_S = \left\{\frac{P}{dz_1} - \theta_s\left[\left(1-\frac{C_*}{C_m}\right)^{B+1} - \left(1-\frac{C_*'}{C_m}\right)^{B+1}\right]\right\}\frac{\rho_w dz_1}{\Delta t} \tag{9}$$

where Q_S is the surface runoff (kg m^{-2} s^{-1}), ρ_w is the density of water (kg m^{-3}) and Δt is the time step (s).

PARAMETERS

Soil parameters

Three representative soil types are used in MOSES: fine, medium and coarse. To obtain the parameters for these soil types, representative percentages of sand, silt and clay were found for the three soil types. Then the method used by Cosby *et al.* (1984) was applied to obtain the values of b, K_s, ψ_s and θ_s from the percentage of sand, silt and clay. The values of θ_w and θ_c were obtained by solving equation (2(c)) at suctions of –153 m and –3.37 m, respectively. These two values are typical low and high soil moisture suctions. The resulting soil hydraulic and evaporation parameters for the coarse, medium and fine soils are given in Table 1.

Table 1 Soil hydraulic parameters.

	Coarse	Medium	Fine
θ_{sat} (m^3 m^{-3})	0.382	0.458	0.456
θ_{crit} (m^3 m^{-3})	0.096	0.242	0.310
θ_{wilt} (m^3 m^{-3})	0.033	0.136	0.221
K_{sat} (kg m^{-2} s^{-1})	0.011	0.00472	0.00363
ψ_{sat} (m)	0.022	0.0495	0.0453
b	3.60	6.63	11.2

PDM parameters

All the parameters for the PDM model can be calculated from the soil parameters in the MOSES model, except for the value of B found in equation (7). The value of B theoretically depends on the degree of heterogeneity of soil moisture within the area being modelled: a value of zero implies a flat homogeneous terrain, and the equations collapse back to MOSES without the PDM. A study in the UK where B was calibrated in 25 catchments (Arnell & Reynard, 1993) showed values from 0 to 2. Often a constant value of 1 is used in modelling studies in Europe. In this study, values of 0, 1 and 2 were used to represent the likely range of this parameter.

APPLICATION

Driving data

The land surface scheme was run using observed meteorological driving data from a site in southern Britain (see Harding *et al.*, 2000). The model is run on an hourly basis and requires incoming long- and short-wave radiation, rainfall, temperature, humidity and wind speed at every time step. These were provided by a Hydra system and an automatic weather station. Default parameters for the land use type "grass" were used in the land surface scheme (see Cox *et al.* (1998) for list of parameters). These include a root depth of 1 m, an unstressed surface resistances of 70 s m^{-1} and a canopy capacity of 0.69 mm.

Initialization

The data as shown in Harding *et al.* (2000) show that 1995 was close to having an annual water balance in this area. It was therefore decided to use 1995 as a model spin up until the soil moisture store was equal at the beginning and the end of the year. Results from the final equilibrium run are presented below.

RESULTS

Figure 2 shows the soil moisture content in the top 1 m of soil for four runs selected to represent the extremes (Sand and Clay, $B = 0$ and 2), while Table 2 summarizes the results of all the simulations in terms of the annual water balance. The biggest soil moisture deficits are developed by the sandy soils with winter to summer deficits of 163–178 mm as compared to deficits of 113–143 mm for the clay soils. This difference ties in with observations (Naden *et al.*, 2000). The sandy soils have higher evaporation

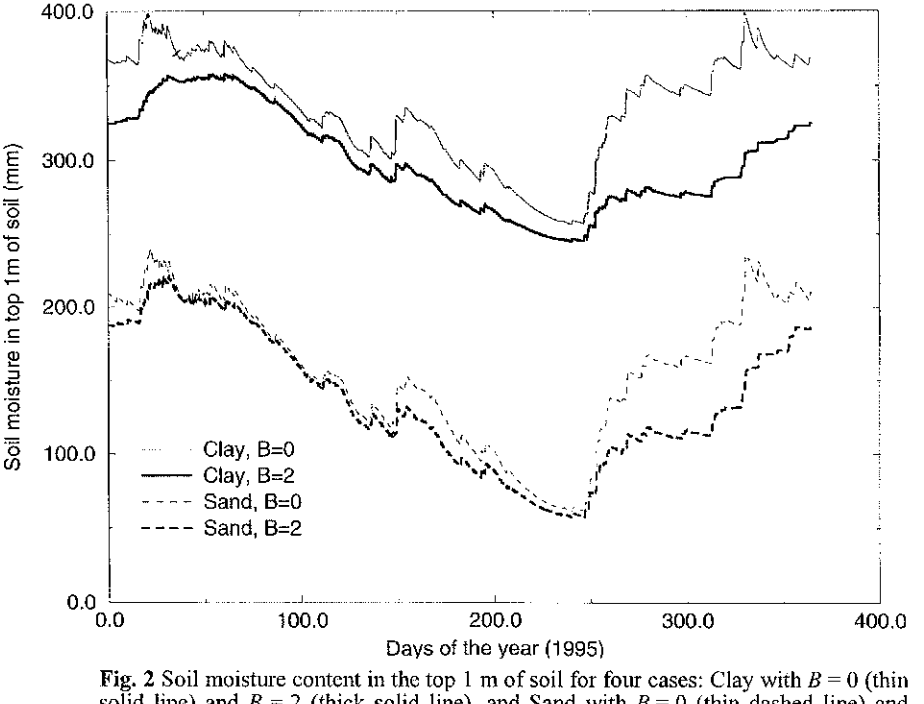

Fig. 2 Soil moisture content in the top 1 m of soil for four cases: Clay with $B = 0$ (thin solid line) and $B = 2$ (thick solid line), and Sand with $B = 0$ (thin dashed line) and $B = 2$ (thick dashed line).

Table 2 Summary of annual water balance for simulations.

	Soil moisture ($mm\ m^{-1}$)		% of annual rainfall	
	Max–min	Evaporation	Drainage	Surface runoff
Clay $B = 0$	143	47	43	10
Clay $B = 1$	125	41	8	51
Clay $B = 2$	113	39	4	57
Loam $B = 0$	172	50	43	7
Loam $B = 1$	155	45	13	42
Loam $B = 2$	144	44	8	48
Sand $B = 0$	178	50	45	5
Sand $B = 1$	168	46	22	32
Sand $B = 2$	163	46	17	37

rates (46–50%) than the clay soils (39–47%). They also, of course, have higher sub-surface runoff (17–45%) than the clay soils (4–43%) and lower surface runoff (5–37%) than the clay soils (10–57%). This trend in the variation of the ratio of sub-surface to surface runoff with soil type is well documented (Boorman *et al.*, 1995).

The range of results and their dependence on the value of B is as follows. A value of $B = 0$ gives the highest winter to summer deficits of 143–178 mm, as opposed to using $B = 2$ which gives 113–163 mm. A value of $B = 0$ gives higher evaporation rates (47–50%) than a value of $B = 2$ (39–46%). It also gives very high sub-surface runoff (43–45%) compared to using a value of $B = 2$ (4–17%), and $B = 0$ gives low surface runoff (5–10%) compared to a value of $B = 2$ (37–57%). This is a quantification of the truism that flat areas will generate less surface runoff than hilly areas.

CONCLUSIONS

From the results described above, it is possible to see that the water balance of the land surface scheme is significantly altered by both the soil parameters in the model, and the parameters used in the hydrological part of the model. It is possible to see that the results are more responsive to the hydrology if the soil is clayey. This is because, in the sandy soils, the water flows down faster through the soil column, and there is less surface water to be affected by the rainfall–runoff model parameters. It is interesting to note that the sandy soils, which allow more water into their matrix, have higher evaporation rates and larger soil moisture deficits, compared to the clayey soils which route more of the rainfall off into the river systems.

It can be concluded that the soil hydraulic parameters and the rainfall–runoff model parameters are equally important in defining the ratio of runoff that originates at the surface to that which originates at the base of the soil column. This indicates that similar investment, both in terms of model complexity and in terms of obtaining the parameters, should be made in representing the vertical flow of soil water and the hydrologically driven runoff.

Acknowledgement The work carried out for this study was funded by the Met Office (Bracknell, UK).

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