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The impact of using area-averaged land surface properties —topography, vegetation condition, soil wetness—in calculations of intermediate scale (approximately 10 km²) surface–atmosphere heat and moisture fluxes

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Abstract

It is commonly assumed that biophysically based soil–vegetation–atmosphere transfer (SVAT) models are scale-invariant with respect to the initial boundary conditions of topography, vegetation condition and soil moisture. In practice, SVAT models that have been developed and tested at the local scale (a few meters or a few tens of meters) are applied almost unmodified within general circulation models (GCMs) of the atmosphere, which have grid areas of 50–500 km². This study, which draws much of its substantive material from the papers of Sellers et al. (1992c, *J. Geophys. Res.*, 97(D17): 19033–19060) and Sellers et al. (1995, *J. Geophys. Res.*, 100(D12): 25607–25629), explores the validity of doing this. The work makes use of the FIFE-89 data set which was collected over a 2 km × 15 km grassland area in Kansas. The site was characterized by high variability in soil moisture and vegetation condition during the late growing season of 1989. The area also has moderate topography.

The 2 km × 15 km ‘testbed’ area was divided into 68 × 501 pixels of 30 m × 30 m spatial resolution, each of which could be assigned topographic, vegetation condition and soil moisture parameters from satellite and in situ observations gathered in FIFE-89. One or more of these surface fields was area-averaged in a series of simulation runs to determine the impact of using large-area means of these initial or boundary conditions on the area-integrated (aggregated) surface fluxes. The results of the study can be summarized as follows:

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1. analyses and some of the simulations indicated that the relationships describing the effects of moderate topography on the surface radiation budget are near-linear and thus largely scale-invariant. The relationships linking the simple ratio vegetation index (*SR*), the canopy conductance parameter (∇_F) and the canopy transpiration flux are also near-linear and similarly scale-invariant to first order. Because of this, it appears that simple area-averaging operations can be applied to these fields with relatively little impact on the calculated surface heat flux.
2. The relationships linking surface and root-zone soil wetness to the soil surface and canopy transpiration rates are non-linear. However, simulation results and observations indicate that soil moisture variability decreases significantly as an area dries out, which partially cancels out the effects of these non-linear functions.

In conclusion, it appears that simple averages of topographic slope and vegetation parameters can be used to calculate surface energy and heat fluxes over a wide range of spatial scales, from a few meters up to many kilometers at least for grassland sites and areas with moderate topography. Although the relationships between soil moisture and evapotranspiration are non-linear for intermediate soil wetnesses, the dynamics of soil drying act to progressively reduce soil moisture variability and thus the impacts of these non-linearities on the area-averaged surface fluxes. These findings indicate that we may be able to use mean values of topography, vegetation condition and soil moisture to calculate the surface–atmosphere fluxes of energy, heat and moisture at larger length scales, to within an acceptable accuracy for climate modeling work. However, further tests over areas with different vegetation types, soils and more extreme topography are required to improve our confidence in this approach.

1. Introduction

The land surface–atmosphere models currently used within general circulation models (GCMs) of the atmosphere are largely based on formulations of processes as observed or conceptualized at relatively small spatial scales, i.e. a few meters. In applying these models within GCMs, there is an implicit assumption of scale-invariance in almost all of the model components. Whereas there are some notable exceptions to this generalization (for example, the treatment of non-spatially uniform precipitation in the Simple Biosphere model as reported by Sato et al. (1989) and of soil wetness spatial distributions by Wetzel and Chang (1987)), by and large most modelers simply drive unaltered versions of their soil–vegetation–atmosphere-transfer (SVAT) models at GCM grid scales using assumed area-averaged lower boundary conditions (vegetation properties, topography, soil moisture) and atmospheric and radiative forcings. The use of increasingly sophisticated SVATs in GCMs (see, e.g. Dickinson, 1984; Sellers et al., 1986, 1996a, b; Bonan, 1994), has almost always resulted in improved model performance and the simulation of more realistic continental climatologies (see, e.g. Dickinson and Henderson-Sellers, 1988; Sato et al., 1989; Nobre et al., 1991; Randall et al., 1996). However, in parallel with this trend of increasing model sophistication has been a growing concern that the scaling assumptions inherent in them may not be valid and that specifying parameters for these models will become increasingly difficult and, at some level, meaningless. These two issues were aired in many discussions in the early and mid-1980s and were the motivation for the International Satellite Land Surface Climatology Project (ISLSCP)

program of field experiments which have the following goals:

1. upscale integration of models: experiments were designed to test the soil–plant–atmosphere models developed by biometeorologists for small-scale applications (millimeters to meters) and to develop methods to apply them at the larger scales (kilometers) appropriate to atmospheric models and satellite remote sensing.
2. Application of satellite remote sensing: the experiments were tasked with exploring methods for using satellite data to quantify important biophysical states and rates for model input, with the ultimate aim of providing a means for initializing and validating SVATs on regional and global scales.

This paper reviews the results of some recent work carried out with a biophysical SVAT, the Simple Biosphere model (SiB) of Sellers et al. (1986). The work used data from the First ISLSCP Field Experiment (FIFE) to explore the impact of area-averaging initial or boundary conditions of topography, vegetation condition and soil moisture on the calculated fields of radiation, heat and water vapor (see Sellers et al., 1992b). In the process, methods of defining surface properties from satellite data were developed and tested.

Fig. 1. Schematic diagram of the simple biosphere model (SiB) as modified for use in the study by Sellers et al. (1992b) and in this study. This version differs from the original one published by Sellers et al. (1986) in that the ground vegetation cover has been removed and the FIFE site grass canopy has been 'promoted' to be the upper-story vegetation layer. The fluxes of sensible and latent heat flux from the surface to the atmosphere are depicted on the right-hand and left-hand sides of the figure, respectively.

2. Theoretical considerations in ‘upscaling’ soil–vegetation–atmosphere transfer (SVAT) models

2.1. The simple biosphere model (SiB)

The current generation of biophysically realistic SVATs is well represented by the BATS model of Dickinson (1984), SiB of Sellers et al. (1986), and LSM of Bonan (1994). In many respects, these models are very similar in their underlying approach and assumptions. All three have explicit representations of the vegetation canopy overlying multi-layer soil models and all three use some version of the Penman–Monteith equation (Monteith, 1973) to calculate surface heat fluxes.

SiB was modified by Sellers et al. (1992a) to run independently of a GCM with the required atmospheric forcings provided by near-surface observations of meteorological conditions and downwelling radiation fluxes. In the study by Sellers et al. (1992a), data from the FIFE network of automatic meteorological stations (AMS) were used to provide these upper boundary conditions.

Fig. 1 shows the structure of SiB: it can be seen that the transfer of soil water into the atmosphere is regulated by a canopy resistance (r_c), a soil surface resistance (r_{surf}) and a network of aerodynamic resistances (r_a , r_b , r_d). Like almost all SVATs in use today, SiB has been demonstrated to provide realistic local-scale simulations of the surface sensible and latent heat fluxes for many different vegetation types (see, e.g. Sato et al., 1989; Sellers et al., 1992a).

2.2. Scale-invariance in SVATs

The issue of scale-invariance with regard to SVAT performance is the extent to which the calculated large-domain land–atmosphere fluxes of heat and water are affected by the use of area-averaged surface boundary conditions; in particular, the fields of topography, vegetation condition and soil moisture. We can address this issue in abstract terms by defining the dependence of a surface flux, Q , which in practice would be sensible heat flux (H), or latent heat flux (LE); on some surface quantities x_i ; which in practice could be the three surface boundary conditions mentioned above.

Thus, if we have a model for Q as a function of x_i which we believe works well at the local scale, we may write

$$Q = f(x_1, x_2, x_3, \dots) \quad (1)$$

An exact calculation of Q over a large domain of area A , i.e. a representative area-averaged value of Q , $\langle Q \rangle$, may be obtained from

$$\langle Q \rangle_1 = \frac{1}{A} \oint_A f(x_1, x_2, x_3) da \quad (2)$$

where $\langle Q \rangle_1$ is the estimate of area-averaged value of surface flux provided by numerical integration.

The angle brackets ‘ $\langle \rangle$ ’ denote ‘area-average’. Fig. 2 (left-hand side) shows how a numerical calculation may be done to yield $\langle Q \rangle_1$ by dividing up the domain A into many

Fig. 2. Left-hand side: a model relating flux Q at a point to a surface variable x_i (e.g. soil moisture) can be numerically integrated over a domain to provide an accurate estimate of the domain-scale flux, $\langle Q \rangle_I$. In practice, this is done by dividing the domain into many small-scale pixels where the length scale is known to be compatible with the model, $Q = f(x_i)$, and summing the results of the individual pixel scale calculations. Right-hand side: the same model can be applied to area-averaged surface parameter values, $\langle x_i \rangle$ to yield an estimate of the area-averaged flux $\langle Q \rangle_M$. However, if $f(x_i)$ is non-linear and the spatial variability of x_i is significant, then $\langle Q \rangle_M$ will be different from the 'true' value as defined by $\langle Q \rangle_I$.

small-scale segments for each of which Eq. (1) can be assumed to be valid. In this approach, the fields of x_i are discretized over the domain so that we can calculate Q for each contributing pixel. For practical reasons, this is not done in GCMs. Instead, it is normally assumed that

$$\langle Q \rangle_M \approx f(\langle x_1 \rangle, \langle x_2 \rangle, \langle x_3 \rangle) \quad (3)$$

where $\langle Q \rangle_M$ is the estimate of area-averaged value of Q provided by taking mean values of x_i over the domain.

Fig. 2 (right-hand side) shows how this use of Eq. (3) reduces the tedious numerical integration of Eq. (2) to a single calculation but, as $f(x_i)$ becomes more non-linear, the difference between the 'true' value of $\langle Q \rangle_I$ as given by Eq. (2), and the area-averaged estimate $\langle Q \rangle_M$ provided by Eq. (3) will increase.

Under what conditions will the two estimates of $\langle Q \rangle$ be approximately equal? Fig. 3(a) shows how for linear functions of $f(x_i)$, the two methods will yield the same result. This we can write

$$\langle Q \rangle_M \rightarrow \langle Q \rangle_I; \text{ as } \frac{\partial^2 f(x_i)}{\partial x_i^2} \rightarrow 0 \quad (4a)$$

Alternatively, if the spatial variability of the x_i field over the domain is very low, the two methods will agree even if the functions are non-linear (see Fig. 3(b)). Thus

$$\langle Q \rangle_M \rightarrow \langle Q \rangle_I, \text{ as } \frac{\sigma_{x_i}}{\langle x_i \rangle} \rightarrow 0 \quad (4b)$$

where σ_{x_i} is the standard deviation of x_i inside the domain.

It is more likely in nature for $f(x_i)$ to be somewhat non-linear and the fields of x_i to show

Fig. 3. Effect of using area-averaged values of surface properties x_i to calculate surface fluxes, $Q = f(x_i)$. (a) For a linear function $f(x_i)$, there is no difference between a fully resolved integration of $Q = f(x_i)$ and a calculation based on $\langle Q \rangle = f(x_i)$; the geometry in this figure explains why this is so. ●, Area-integrated calculation, $\langle Q \rangle_I$, corresponding to the left-hand side of Fig. 2; ○, area-averaged calculation, $\langle Q \rangle_M$, corresponding to the right-hand side of Fig. 2. (b) For a non-linear function $f(x_i)$, there can be a significant difference between $\langle Q \rangle_I$ and $\langle Q \rangle_M$ where the spatial variability in x_i is large (upper case). When the spatial variability in x_i is low, $\langle Q \rangle_M \rightarrow \langle Q \rangle_I$.

some degree of spatial variability, so it is the combination of these factors that is important. Thus we may write a general criterion which combines Eq. (4a) and Eq. (4b):

$$\langle Q \rangle_M \rightarrow \langle Q \rangle_I, \text{ as } \frac{\partial^2 f(x_i)}{\partial x_i^2} \frac{\sigma x_i}{\langle x_i \rangle} \rightarrow 0 \quad (4c)$$

How does this analysis transfer to the actual functioning of a typical biophysical SVAT?

2.3. Theoretical effects of variability in surface quantities on the surface energy balance

The surface quantities considered in this study are topography, vegetation condition and soil moisture. The effects of spatial variation in each of these on the surface energy balance are briefly reviewed below.

(a) Direct Beam

(b) Diffuse Radiation

Fig. 4. Interception of (a) direct beam radiation and (b) diffuse radiation as treated in this modified version of SiB. The factor β_b in eqn (5) represents the ratio of radiation intercepted by a sloping pixel compared with a horizontal one. In (a), the solar beam has the same azimuth as the slope and β_b is equivalent to the area shadowed by the pixel $(1 + x)$. In (b), each portion of the pixel views a fractional area of sky above the horizon of $(\pi - V)/\pi$ and an area of landscape below the horizon of fraction area V/π .

2.3.1. Topography

Topographic variations have a direct effect on the absorption and reflectances or emission of radiation by the surface. In addition, topography determines how soil moisture is redistributed when the soil is wet and thereby to a large extent determines an area's streamflow response. The work in this paper addresses only the effects of topography on the radiation balance; for the examples considered, the role of interflow and runoff is small (see discussion at the end of this paper).

The amount of shortwave (S_N) and longwave (L_N) radiation absorbed by a pixel within the domain depends on slope and aspect (see Fig. 4). In this study, we adapted the SiB radiation submodel to take account of illumination geometry. The modification is simple and does not take into account the effects of topographic shadowing by other pixel elements in the scene.

$$\begin{aligned}
S_N = & [S \downarrow_{b,v} (1 - a_{b,v}) + S \downarrow_{b,n} (1 - a_{b,n})] \beta_b + [S \downarrow_{d,v} (1 - a_{d,v}) \\
& + S \downarrow_{d,n} (1 - a_{d,n})] \beta_d + [S \downarrow_{b,v} a_{b,v} + S \downarrow_{d,v} a_{d,v}] (1 - a_{d,v}) \beta_r \\
& + [S \downarrow_{b,n} a_{b,n} + S \downarrow_{d,n} a_{d,n}] (1 - a_{d,n}) \beta_r
\end{aligned} \quad (5a)$$

where S_N is net shortwave radiation absorbed by a pixel (W m^{-2}), $S \downarrow_{\mu,\Lambda}$ is normal downward shortwave flux for angular component μ and waveband Λ (μ is b for solar beam or d for diffuse); Λ is v for visible or n for near-IR), $a_{\mu,\Lambda}$ is surface spectral reflectance,

$$\beta_b = 1 + \tan V \tan Z \cos (\alpha_v - \alpha_z) \quad (5b)$$

$$\beta_d = \left(\frac{\pi - V}{\pi} \right) \frac{1}{\cos V} \quad (5c)$$

$$\beta_r = \frac{V}{\pi \cos V} \quad (5d)$$

where β_b is direct beam radiation correction factor for slope and aspect, β_d is the fraction of sky dome visible to the pixel, β_r is the fraction of sky dome below the horizon for a sloping pixel, Z is solar zenith angle, V is slope angle, and α_v and α_z are slope azimuth (aspect) and solar azimuth, respectively

The first term in Eq. (5a) deals with the absorption of the direct beam radiation, the second term deals with the interception of diffuse sky radiation, and the third and fourth terms cover the absorption of radiation reflected from the surrounding landscape which would be exposed to a sloping pixel from the horizon downwards. It should be noted that we assume that the landscape has the same reflectance properties as the subject pixel and that there is no shadowing of the subject pixel by distant topography. (A complete radiative transfer treatment of these issues would be complex and computationally expensive).

The direct beam surface reflectances, $a_{b,v}$ and $a_{b,n}$, are a function of the solar beam illumination angle (see Sellers, 1985). A correction is made in this treatment to make these reflectances a function of the solar zenith angle as referenced to the slope normal. This 'effective' solar zenith angle, Z' , which is used to calculate $a_{b,\Lambda}$, is given by

$$\cos Z' = \cos Z \cos V + \sin Z \sin V \cos (\alpha_z - \alpha_v) \quad (5e)$$

The treatment of longwave radiation exchanges is similar to that for shortwave diffuse radiation:

$$L_N = L \downarrow \beta_d + \varepsilon \sigma [f_c T_c^4 + (1 - f_c) T_g^4] \left(\beta_r - \frac{1}{\cos V} \right) \quad (5f)$$

where L_N is net longwave radiation absorbed by the pixel (W m^{-2}), $L \downarrow$ is normal longwave flux (diffuse; thermal) (W m^{-2}), σ is the Stefan-Boltzman constant ($\text{W m}^{-2} \text{K}^{-4}$), f_c is fraction of ground covered by vegetation, T_c and T_g are canopy and ground surface temperature (K), and ε is emissivity (assumed $\varepsilon = 1$).

The effects of topography on the radiation budget are represented by the factors β_b , β_d and β_r (see Eq. (5b,c,d)). Generally speaking, deviations from a flat surface usually lead to

an increase in the absorption and emission of longwave radiation with a small net energy gain for the surface as the fraction of cold sky viewed by the surface decreases. The impact of topography on the interception of diffuse radiation is fairly small, as the terms multiplied by β_d and β_r tend to decrease slowly from the solution for a level surface as V increases. The absorption of direction beam radiation is almost linearly related to V through $\tan V$ in β_b .

An inspection of Eqs. (5a), (5b) and (5c) shows that for moderate values of V , say 25° or less, the impact of slope angle on absorbed radiation is almost linearly related to V . (The $\cos V$ terms are relatively weak for values of V of 0 – 25° .) Because of this, we might expect some small changes in net radiation over a domain owing to increased topography, but, overall, the effects are partially self-canceling over a day (south slope gains roughly cancel out north slope losses for direct beam radiation, for example) and should be well behaved owing to the near-linearity of the dependence of S_N on V . The formulation of Eq. (5) thus roughly conforms to the condition specified in Eq. (4), and so we anticipate that simple area-averaging of topography will have little impact on the net radiation and the total heat fluxes, $\langle Q \rangle$, for a domain with no significant net slope.

2.3.2. Vegetation condition

In SiB, the canopy transpiration rate, λE_{ct} , is calculated from a simple resistance model (see Fig. 1):

$$\lambda E_{ct} = \frac{[e^*(T_c) - e_a] \rho c_p}{r_c + r_b} \gamma \quad (6)$$

where $e^*(T_c)$ is the saturated vapor pressure inside canopy foliage (Pa), e_a is the vapor pressure in the canopy air space (Pa), r_c is canopy resistance ($s\ m^{-1}$), r_b is canopy bulk leaf boundary layer resistance ($s\ m^{-1}$), ρ is density of air ($kg\ m^{-3}$), c_p is specific heat of air ($J\ kg^{-1}\ K^{-1}$) and γ is the psychrometric constant ($Pa\ K^{-1}$).

Under normal daytime conditions, $r_c \gg r_b$ so λE_{ct} is roughly proportional to the canopy conductance $g_c = 1/r_c$. Sellers et al. (1992a) showed that for the grass cover found in the FIFE area, g_c was well described by

$$g_c = \frac{1}{r_c} = \nabla_F F_0 [f(\delta e) f(T) f(\psi_l)] \quad (7)$$

where g_c is canopy conductance ($m\ s^{-1}$), r_c is canopy resistance ($s\ m^{-1}$), $\nabla_F = (\partial g_c^* / \partial F_0)$ ($m^3\ J^{-1}$), g_c^* is unstressed (vegetation- and light-dependent components only) value of g_c ($m\ s^{-1}$) and F_0 is incident flux of PAR ($W\ m^{-2}$). $f(\delta e)$, $f(\psi_l)$ and $f(T)$ are environmental stress terms which take into account the effects of vapor pressure deficit (δe), leaf water potential (ψ_l), which is a function of soil moisture potential and stress, and temperature T .

Eq. (7) is similar to the empirical equation set of Jarvis (1976), which was put forward to describe the stomatal conductance of individual leaves as a function of environmental variables. The critical term in Eq. (7) is ∇_F , which should be a function of the fraction of incident photosynthetically active radiation (PAR) absorbed by the green portion of the vegetation canopy (FPAR), and vegetation physiology only (see Sellers et al., 1992a). Fig. 5(a) shows how g_c^* , the unstressed or maximum canopy conductance, is supposed to vary

Fig. 5. Schematic figures illustrating the hypothesized relationships between the unstressed canopy conductance, g_c^* , and the simple ratio (SR) vegetation index. (a) Changes in unstressed canopy conductance, g_c^* , for a range of canopy states and incident photosynthetically active radiation (PAR) fluxes, F_0 . (Note that all the canopy conductances saturate at the same level of F_0 , F_0^{sat} ; that is, V_{max0} is invariant in this example.) Derivatives of g_c^* with respect to F_0 , ($F_0 < F_0^{\text{sat}}$) are shown. (b) Derivatives of g_c^* with respect to F_0 , $\nabla_F = \partial g_c^* / \partial F_0$, are plotted against SR vegetation index. The letters on the line correspond to the cases shown in (a). These figures illustrate the principles discussed in the text; they contain no real data. (Figure reproduced from Sellers et al., 1992c).

as a function of canopy greenness and incident PAR: as the density of green vegetation increases, the area-averaged response of g_c^* to changes in PAR is expected to be more lively, leading to an increase in ∇_F .

The theoretical work of Sellers (1985, 1987) and Sellers et al. (1992a) indicated that ∇_F should be near-linearly related to spectral vegetation indices obtained from satellite data (see Fig. 5(b)). Sellers et al. (1992c) used the FIFE-87 surface flux station data and associated remote sensing data to test the validity of this hypothesis. SiB was adapted to operate in the inverse mode; that is, the model was forced to reproduce latent and sensible heat fluxes that converged on above-canopy flux observations by iterative adjustment of one or two parameters controlling evapotranspiration. This procedure resulted in retrieval of ∇_F values for several surface flux stations for different times of the year which were then compared with spectral reflectances as observed by satellites for

Fig. 6. Retrieved values of ∇_F for one FIFE-89 and two FIFE-87 flux stations plotted against remotely sensed variables derived from Landsat TM (continuous lines) and SPOT (dashed lines) data. Data for the FIFE-87 Stations 16 and 18, and for the FIFE-89 916 station are shown (Sites 16 and 916 are exactly equivalent, Sites 16 and 18 were colocated). The ∇_F values are plotted against Landsat TM and SPOT reflectance-based SR values; error bars represent the uncertainties owing to the soil moisture formulation, and the overall model goodness-of-fit.

the $90 \text{ m} \times 90 \text{ m}$ area (nine Landsat thematic mapper pixels) centered on each flux station. The results are shown in Fig. 6 for a pair of flux stations that operated in 1987 and 1989; ∇_F is compared with the simple ratio vegetation index, SR , as given by the ratio of the near-IR and visible surface reflectances. From Fig. 6, we obtain an expression which should be a reasonable descriptor of ∇_F for the C_4 vegetation which dominates the FIFE area.

$$\nabla_F = a SR + b; \quad r^2 = 0.717 \quad (8)$$

where $a = 0.495 \times 10^{-5} \text{ m}^3 \text{ J}^{-1}$, $b = -1.033 \times 10^{-5} \text{ m}^3 \text{ J}^{-1}$ and SR is ρ_N/ρ_V (where ρ_N and ρ_V are near-IR and visible surface reflectances derived from Landsat and SPOT data, respectively).

From Eq. (6), Eq. (7) and Eq. (8), we see that the canopy transpiration rate, λE_{ct} is roughly proportional to g_c , which in turn is near-linearly related to SR . This chain of near-linear relationships suggests that if we were to use area-averaged values of SR to specify $\langle g_c^* \rangle$, there would be little impact on the area-averaged latent heat flux, i.e. $\langle LE \rangle_M \approx \langle LE \rangle_I$ in the absence of significant environmental stress. This is because the transpiration formulation approximately satisfies the condition represented by Eq. (4).

2.3.3. Soil moisture

Soil moisture is a critical regulator of evapotranspiration at the local scale; as soil

moisture decreases, vegetation stomates close in response to hormones flowing from the drying root system (see Gollan et al. (1986, 1992) and Mansfield and de Silva (1994)). At the same time, the direct evaporation flux from the soil surface rapidly falls off.

The effect of soil moisture stress on transpiration is represented in SiB by $f(\psi_1)$, the leaf water potential stress factor in Eq. (7). An expression for $f(\psi_1)$ and the other stress factors, $f(\delta e)$ and $f(T)$, were obtained from curve fits to the data of Verma et al. (1993) such that

$$f(\delta e) = \frac{1}{1 + b\delta e} \quad (9a)$$

where $b = 0.0002 \text{ Pa}^{-1}$ and δe is vapor pressure deficit in the canopy air space (Pa);

$$f(\psi_1) = \frac{\psi_1 - \psi_{c_2}}{\psi_{c_1} - \psi_{c_2}} \quad (9b)$$

where ψ_1 is leaf water potential (Pa) and ψ_{c_1} and ψ_{c_2} are leaf water potentials at which stomates start to close and are completely closed, respectively (Pa);

$$f(T) = 1 \quad (9c)$$

The factors $f(\delta e)$, $f(\psi_1)$ and $f(T)$ are allowed to vary between zero, where stress is simulated to completely shut down transpiration, and unity, where no stress limitations are applied to g_c^* .

In SiB, the wetnesses of the three soil layers, W_1 , W_2 and W_3 , are defined by dividing the volumetric water content of each layer by the porosity, thus W_i varies between zero and unity. The upper soil layer (W_1) is 0.05 m thick, the root zone (W_2) is 0.20 m thick and the recharge layer (W_3) varies from 0.25 to 1.75 m thick in this study. The leaf water potential, ψ_1 , is determined from a catenary equation with the water source defined by the root zone soil moisture potential, a direct function of W_2 (see Sellers et al., 1986). In practice, the leaf water potential is very close to the root zone soil moisture potential:

$$\psi_1 \rightarrow \psi_2 = \psi_s W_2^{-b} \quad (10)$$

where ψ_2 is root zone (second soil layer) soil moisture potential (Pa), ψ_s is the saturated value of ψ_2 (Pa), W_2 is root zone soil wetness fraction and equals unity when saturated, and b is a sorptivity parameter.

The soil hydraulic parameters ψ_s , b , soil porosity and soil hydraulic conductivity are all functions of soil type. With regard to $f(T)$, the work of Polley et al. (1992) and Verma et al. (1993) indicated that leaf photosynthesis was only weakly dependent on leaf temperature in the C_4 species found in the FIFE area and so $f(T)$ was set to unity.

Soil evaporation can make a significant contribution to the total evapotranspiration flux (see Villalobos and Fereres (1990) and Sellers et al. (1992c)). In the study by Sellers et al. (1992c), a series of inverse-mode runs with SiB were used to estimate soil evaporation rates and soil surface resistances, r_{surf} , from the time-series of flux station data; this procedure resulted in the formulation of an empirical expression relating r_{surf} to the wetness of the uppermost soil layer, W_1 :

$$r_{\text{surf}} = \exp(8.206 - 4.255W_1) \quad (11)$$

where r_{surf} is soil surface resistance (s m^{-1}), which is 150 s m^{-1} or more, and W_1 is soil wetness of the (0–5 cm) upper soil layer.

Fig. 7. Dependence of the soil surface resistance, r_{surf} , on W_1 (eqn (11)) and the vegetation moisture stress function, $f(\psi_1)$, on W_2 (eqn (9)). The following should be noted: (1) r_{surf} is kept constant for values of $W > 0.75$; in practice, such high levels of surface soil moisture cannot be maintained for very long; (2) $f(\psi_1)$ is plotted as a direct function of W_2 for the purpose of illustration; i.e. the plant and soil resistance terms are taken as negligible here, so $\psi_1 = \psi_2$.

Fig. 7 shows the variation of $f(\psi_1)$ and r_{surf} with soil wetness (W_2 and W_1 , respectively). Clearly, these functions are very non-linear for drier soil conditions. It is also widely reported that soil moisture variability can be significant, particularly for wet soil moisture conditions (see, e.g. Charpentier and Groffman, 1992, Tables 3 and 4). We might therefore expect that as neither Eq. (4a) nor Eq. (4b) are likely to be satisfied by the soil moisture–surface flux relationships, then area-averaging soil moisture over an intermediate scale domain could have severe impacts on the calculated fluxes, i.e. $\langle Q \rangle_1 \neq \langle Q \rangle_M$.

3. The FIFE data and model validation at the local scale

3.1. The FIFE-89 data set

The FIFE-89 ‘testbed’ data set was selected for the study of Sellers et al. (1995), which focused on the impacts of spatial variability on area-averaged fluxes. Much of the work reported by Sellers et al. (1995) is summarized here.

The data collection effort in FIFE-89 was concentrated primarily within a 2 km × 15 km strip oriented north–south in the middle of the FIFE area. This so-called ‘testbed’ area contained most of the surface flux stations and was also made the first priority for airborne data collection.

Fig. 8. Schematic diagram showing the FIFE-89 testbed area data sets used in this study. The figure also shows the location of the testbed area in the FIFE area. (A) Initial or boundary conditions: (i) vegetation condition: SPOT and TM data were used to calculate surface reflectances (ρ_N , ρ_V) and simple ratio vegetation index (SR) fields at $30\text{ m} \times 30\text{ m}$ resolution; (ii) surface soil moisture: the airborne pushbroom microwave radiometer (PBM) mounted on the NASA C-130 mapped surface (0–5 cm) soil moisture at $50\text{ m} \times 50\text{ m}$ resolution; (iii) root zone, deep soil moisture: neutron probe and gravimetric data were used to develop regression equations to specify these soil moisture fields; (iv) topography: digital elevation model (DEM) from USGS data, aggregated to $30\text{ m} \times 30\text{ m}$ resolution. (B) Forcing variables: (v) near-surface meteorology and downwelling radiation fluxes: automatic meteorological stations (AMS) measured near-surface meteorological variables and downwelling shortwave and longwave radiation fluxes. (C) Validation data: (vi) surface fluxes: surface flux stations measure fluxes at half-hourly time resolution; (vii) airborne flux measurements: the Canadian Twin Otter aircraft flies at 15 m altitude to map surface fluxes at coarse ($2\text{ km} \times 4\text{ km}$) resolution.

There was considerable spatial variability in the fields of surface soil moisture and green vegetation cover in FIFE-89, owing to the after-effects of convective storms that covered the southeastern corner of the area on Days 204 and 212. In addition, the area was subjected to a long dry period from Day 216 to Day 224 during which a series of airborne remote sensing missions collected microwave data to monitor the drying-out of the surface soil layer. Airborne and surface flux measurements tracked the decrease in the surface latent heat fluxes over the same period. All in all, the FIFE-89 data set gives us a good example of a landscape with well-developed spatial heterogeneity in topography, vegetation condition and soil moisture with some significant changes in soil moisture over time as the site dried out. It should be remembered, however, that the vegetation type is an almost uniform grassland within the domain.

The critical data collected in FIFE-89 are depicted schematically in Fig. 8. They include the following.

3.1.1. Surface boundary conditions and model initialization data

(i) Atmospherically corrected SPOT and Landsat TM surface spectral reflectances at $30\text{ m} \times 30\text{ m}$, combined into *SR* fields.

(ii) Spatially continuous airborne microwave data collected by the NASA C-130 aircraft equipped with the pushbroom microwave radiometer (PBMR). The PBMR data, described in detail by Wang et al. (1992), were used to infer surface (0–5 cm) soil moisture fields using a regression equation based on a large number of gravimetric samples acquired at the same time as the PBMR data. The PBMR data were mapped by J.R. Wang et al. (personal communication, 1995) to a $15\text{ m} \times 15\text{ m}$ grid.

(iii) Soil moisture profiles. In addition to the daily near-surface soil gravimetric samples taken around the flux station sites, the soil moisture profile from 10 cm to the bedrock (usually at 50–150 cm depth) was monitored using neutron probes.

(iv) Digital elevation model (DEM). A 25 m resolution DEM was prepared from US Geological Survey data by the FIFE Information System (FIS). The vertical resolution of these data appears to be good to better than 5 m (see Davis et al., 1992).

3.1.2. Model forcing data

(v) The network of automatic meteorological stations (AMS) and instruments mounted on the surface flux stations provided continuous measurements of the near-surface meteorological variables: temperature (T), mixing ratio (q), wind speed (u), downwelling shortwave, longwave and net radiation fluxes ($S\downarrow$, $L\downarrow$ and R_n), and precipitation (P) over the area. Some comparisons showed that except for precipitation there was very little spatial variability in these forcings during the dry period, Days 216–224, and so spatial means of the AMS variables were used to provide uniform upper boundary conditions for the model runs.

3.1.3. Model validation data

(vi) Surface fluxes: the surface flux stations shown in Fig. 8 provided near-continuous time-series estimates of net radiation, latent heat, sensible heat, momentum and in some cases total CO_2 flux. The instruments and techniques have been described by Kanemasu et al. (1992).

(vii) Airborne fluxes: the National Research Council of Canada contributed the Twin Otter flux aircraft to both FIFE-87 and FIFE-89. The aircraft is equipped to measure the turbulent fluxes of sensible and latent heat, CO_2 and momentum see Desjardins et al. (1992b) and MacPherson et al. (1992).

3.2. Model initialization

The Landsat 30 m $\times$ 30 m data grid defined for the FIFE area by FIS was used as the basis for the full-resolution studies within the FIFE-89 testbed area. The digital elevation model (DEM), soil moisture and other data sets were all gridded to this 68 $\times$ 501 pixel domain. The full-resolution calculations of energy and water balances discussed in the next section were performed for each of these pixels.

The ∇_F , FPAR and leaf area index (LAI) values for each pixel were specified from *SR* fields as derived from TM or SPOT data using Eq. (8), and methods discussed by Sellers et al. (1992a,b). The LAI, canopy greenness fraction and soil background reflectance values are used by the SiB radiation submodel to calculate surface albedo and the absorption of radiation by the canopy and ground. Likewise, these LAI data enter into the turbulent transfer submodel to calculate roughness lengths (z_0) and other aerodynamic parameters. Soil physical properties (porosity, θ_s ; saturated hydraulic conductivity, K_s ; saturation moisture potential, ψ_s ; sorption parameter, B) were specified from data collected by the FIFE Staff Science soil survey work and the figures of Clapp and Hornberger (1978) (see Sellers et al., 1992c). Soil physical properties are assumed to be uniform over the site.

The prognostic variables of SiB are the canopy, ground surface and deep soil temperatures (T_c , T_g and T_d , respectively), the canopy and soil interception stores (M_c and M_g) and the soil wetnesses (W_1 , W_2 and W_3). T_c and T_g were initialized from the air temperature and T_d was set equal to the 50 cm soil temperature as measured by the AMS network; mean values were used for the entire testbed area. M_c and M_g were both set to zero for the first timestep. The surface layer (0–5 cm) soil wetness, W_1 , was initialized using the PBMR data of Wang et al. (1992). The soil moisture contents for the root zone, W_2 , and the deep soil layer, W_3 , are more difficult to specify directly from observations; all we have are the neutron probe readings for the AMS and flux station sites. Using these and the other available FIFE-89 data, a study was made into which of the following variables are the best predictors of W_2 and W_3 : *SR*, W_1 (from the PBMR data), slope and elevation. The best fits were provided by regression equations operating on local slope angle, simple ratio and W_1 (see Sellers et al., 1996a).

The deepest neutron probe reading taken at AMS and flux station sites generally corresponded to the bedrock depth. These readings were compared with the local slope and elevation data to specify the spatial variation of the total soil profile depth (see Sellers et al., 1995): this depth varied from 0.5 to 2.0 m in the testbed area.

3.3. Model operation

Generally, model runs were started at some time close to the acquisition of the PBMR data used to initialize W_1 . This is because all of the other initialization data are more or less continuously available (temperatures), slowly changing in time (∇_F , FPAR and LAI) or

Fig. 9. Schematic diagram comparing full-resolution (68×501 pixels) and area-averaged 'block' calculations for the FIFE-89 testbed area. In this example, all the fields represented in detail in the full-resolution calculation are area-averaged to provide a set of mean initial or boundary conditions for the area-averaged run, which can then be treated as one large pixel. Any intermediate calculation, where at least one field is represented at full resolution, must be carried out at full resolution. The full-resolution flux fields may be area-integrated or aggregated for direct comparison with the area-averaged results.

constant (soil properties, topography). For the full resolution calculations, the energy and water balances were calculated at hourly time resolution for each of the ($68 \times 501 = 34\,068$) $30\text{ m} \times 30\text{ m}$ pixels within the testbed domain.

With the incorporation of the modified radiation code (see Eq. (5)), SiB can be used to

compare full resolution (area-integrated) with block (area-averaged) calculations of the surface heat fluxes. Area-integrated values of prognostic or diagnostic variables calculated by these full-resolution runs are provided by simply taking the mean of the appropriate quantities over all the pixels in the testbed domain (see the left-hand side of Fig. 9). Area-averaged calculations are performed by using a single model run with area-averaged initial conditions and prognostic variables (see the right-hand side of Fig. 9). The initial conditions for the area-averaged runs are provided by averaging the initial fields used by the full-resolution runs. A mean of the absolute values of the slopes of all the pixels is used to calculate the baseflow from the soil recharge layer (W_3). For the radiation calculations, the area-averaged slope and aspect over the domain work out to be almost negligible.

3.4. Model validation using surface and airborne flux data

The $90\text{ m} \times 90\text{ m}$, 3×3 pixel cluster centered on each flux site is assumed to be representative of the surface flux 'footprint' characterized by that flux station (see Schuepp et al., 1990). Fig. 10(a) shows model time-series of fluxes produced by the SiB model for Site 916, Day 216, using the initial conditions and parameter sets specified in the previous section; the model does a reasonable job of reproducing the observed fluxes. Fig. 10(b) shows the results of model calculations for a number of flux stations for a day with relatively wet soil moisture conditions, Day 216, and days with drier soil moisture conditions, Days 222, 223 and 224. The evaporative fractions shown in Fig. 10(b) are

Fig. 10. Comparisons between surfaces fluxes calculated by SiB in this study and as measured in the field in FIFE-89. (a) Comparison of observed and calculated diurnal courses of surface heat and energy fluxes for Site 916 (ECV-4439) for Day 216. Calculated heat fluxes for a $90\text{ m} \times 90\text{ m}$ area centered on the flux station site (3×3 Landsat pixels) are shown. (b) Comparison of observed and calculated evaporative fractions for a number of surface flux stations in FIFE-89; the numbers in the figure represent individual sites. The points above the dashed line cutting the 1:1 line are for the period 13:45–15:45 h local time, Day 216 (relatively wet soil conditions) and the points below the dashed line are for the periods 10:45–12:15 h, Day 222; 11:45–13:45 h, Day 223; 11:45–13:45 h local time, Day 224 (relatively dry soil moisture conditions). (c) Comparison of observed and calculated evaporative fractions for a series of blocks aligned along the center of the testbed area (see Fig. 8). The observed values were produced for $2\text{ km} \times 4\text{ km}$ blocks from airborne eddy correlation data. The calculated values represent $2\text{ km} \times 2\text{ km}$ blocks centered on the airborne data blocks; they were 'averaged-up' from the $30\text{ m} \times 30\text{ m}$ calculations for the same periods and days as specified in (b) above. The dashed line delineates between Day 216 and Day 224 results.

the means of the calculated and observed fluxes for the periods when the Twin Otter flux aircraft was flying over the area. The results from the same full-resolution runs were area-integrated to provide estimates for $2 \text{ km} \times 2 \text{ km}$ blocks for the same periods, for comparison with overlapping $2 \text{ km} \times 4 \text{ km}$ blocks of footprint-adjusted surface flux estimates produced by the Twin Otter (see Fig. 10(c) and Desjardins et al., 1992b). The flux data of Desjardins et al. (1992b) are not directly comparable with the calculations as the domains are different ($2 \text{ km} \times 4 \text{ km}$ vs. $2 \text{ km} \times 2 \text{ km}$) and it is known that estimates of the surface fluxes provided by the flux aircraft are consistently low (see Desjardins et al. (1992a) and Sellers et al. (1992c)). However, it appears that ratios of these estimates (e.g. the evaporative fraction, $EF = \lambda E / (\lambda E + H)$) are more consistent with equivalent surface observations (see Fig. 10(c) and Sellers et al. (1992c)).

The results shown in Fig. 10 and other work reported by Sellers et al. (1992b; 1995) indicate that all aspects of the model perform credibly at the local ($90 \text{ m} \times 90 \text{ m}$) scale when using initial conditions provided by meteorological observations for the prognostic variables, and remotely sensed measurements for the vegetation condition (∇_F , albedo,

Table 1
Results of exploratory tests described in Section 4.1

(a)														
<i>i</i>	Treatment				$R_a - G$		λE		H		G		EF	
					Mean	σ	Mean	σ	Mean	σ	Mean	σ	Mean	σ
1	top	SR	W		429.6	29.1	270.5	56.2	159.5	60.4	46.5	3.9	0.629	0.134
2	(top)	SR	W		429.0	12.8	270.7	55.7	158.6	57.2	46.5	3.4	0.631	0.132
3	top	(SR)	W		428.6	28.9	273.2	39.0	155.8	40.5	46.3	2.8	0.637	0.092
4	top	SR	(W)		429.4	29.1	274.1	33.4	155.7	40.8	46.9	3.4	0.638	0.084
5	(top)	(SR)	W		427.6	12.8	273.4	37.9	154.9	36.3	46.3	2.3	0.638	0.089
6	(top)	SR	(W)		428.5	12.7	274.3	32.3	154.8	36.8	46.9	2.8	0.639	0.080
7	top	(SR)	(W)		428.4	28.9	277.3	11.1	151.7	18.1	46.7	2.0	0.646	0.027
8	(top)	(SR)	(W)		429.2	0.0	277.6	0.0	151.6	0.0	46.7	0.0	0.647	0.000

(b)														
<i>i</i>	Treatment				$R_a - G$		λE		H		G		EF	
					Mean	σ	Mean	σ	Mean	σ	Mean	σ	Mean	σ
1	top	SR	W		346.9	23.5	176.0	51.2	171.4	55.2	33.6	3.5	0.507	0.151
2	(top)	SR	W		345.9	10.6	176.2	51.0	170.3	52.4	33.6	2.9	0.509	0.150
3	top	(SR)	W		345.7	23.5	177.8	30.6	168.4	34.2	33.4	2.3	0.514	0.090
4	top	SR	(W)		346.9	23.7	175.4	28.6	171.9	35.5	33.7	3.5	0.505	0.089
5	(top)	(SR)	W		344.7	11.0	178.1	30.1	167.2	29.8	33.4	1.4	0.516	0.088
6	(top)	SR	(W)		345.9	10.8	175.7	28.0	170.8	31.9	33.6	2.8	0.507	0.086
7	top	(SR)	(W)		346.2	23.6	179.5	7.0	167.1	17.1	33.3	2.2	0.518	0.027
8	(top)	(SR)	(W)		346.6	0.0	179.9	0.0	166.6	0.0	33.3	0.0	0.519	0.000

Initial or boundary conditions and forcings for (a) Day 216 and (b) Day 224 were used to run SiB for the tested domain. The results of model runs with fully resolved fields of topography (top), soil moisture (W) and simple ratio (SR) are shown in Lines 1 of (a) and (b); Line 8 shows equivalent results when all three fields are represented by area-averages which degenerate to a single energy balance calculation for the entire domain. Lines 2–7 show the results of using area-averages of one or two of the three fields.

etc.) and surface soil wetness (W_1). The model also appears to reproduce the responses of the energy balance to changes in soil moisture at this spatial scale, as can be seen by the range of EF values in Fig. 10.

4. Investigating the effects of inhomogeneity on intermediate ($2 \text{ km} \times 15 \text{ km}$) scale fluxes

4.1. Exploratory tests

Three kinds of surface heterogeneity are addressed in this study: topography (slope and aspect), vegetation condition (V_F and LAI), and soil wetness (principally variations in W_1). Table 1 and Fig. 11 show the results of exploratory tests performed using the FIFE-89 testbed data set. For Days 216 and 224, a series of 24 h simulations were performed starting with a full resolution calculation in which all these fields were fully resolved,

Fig. 11. Comparison of energy balance calculations conducted for the FIFE-89 testbed area using the full-resolution (continuous line with error bars) and area-averaged or 'block' (heavy dashed line) methods. The full-resolution results represent the mean and standard deviations of the 68×501 pixel calculations within the domain; the area-averaged results are based on a single 'big-pixel' calculation for the entire testbed area using mean initial or boundary conditions and prognostic variables. Day 216: (a) latent heat flux; (b) sensible heat flux. Day 224: (c) latent heat flux; (d) sensible heat flux.

followed by a succession of runs in which one or more of the initial fields was assigned a mean value over the whole domain. The final rows of Table 1(a) and Table 1(b) show the results of the area-average ‘block’ runs where a single simulation was done for the entire testbed area using area-averaged initial conditions and prognostic variables. This area-average calculation is identical to the area-integral of pixel-scale calculations carried out over the testbed domain where each pixel is given the same mean initial and boundary conditions.

Fig. 11 shows the diurnal course of the heat fluxes for the testbed domain as calculated using fully resolved and area-averaged fields of soil moisture, topography and SR. It is surprising how closely the results from the two methods agree (see also Fig. 14 of Sellers et al., 1992c).

The results in Table 1 show what we would expect from the analysis in Section 2 of this paper. The functions governing the response of the energy balance to changes in vegetation condition and topography are more or less linear in these variables, so the impact of area-averaging those variables is weak. This result was anticipated for the vegetation component by Sellers et al. (1992c) and for the effects of topography on the radiation balance by Dubayah et al. (1990).

However, the dependences of the soil surface resistance, r_{surf} , on W_1 , and the vegetation moisture stress function, $f(\psi_1)$, on W_2 are non-linear (see Fig. 7). As a result, the area-averaging of soil wetness fields leads to a relatively large divergence between the area-integrated and area-averaged surface flux estimates in Table 1. A comparison of Lines 1 (all three fields fully resolved) and Lines 5 (only the soil moisture field fully resolved) in Table 1(a) and Table 1(b) reveals that soil wetness variability accounts for most of the calculated variability in the latent and sensible heat flux fields (σ_{LE} and σ_{H}) and, more significantly, for the greatest mismatch between the area-integrated (Line 1) and area-averaged (Lines 5 and 8) estimates of heat fluxes.

Sellers et al. (1995) presented indices which quantified the differences between $\langle Q \rangle_1$ and $\langle Q \rangle_M$ relative to standard deviations in the surface property fields. These indices show that soil moisture is the most problematical of the three variables with respect to area-averaging. Given these findings, it may be concluded that simple area-averaging procedures can be used on the fields of SR (provided we are dealing with one type of vegetation in the domain) and topography (provided the topography is moderate) without greatly affecting the calculated larger-scale fields of sensible and latent heat flux. In the next section, we therefore concentrate on the influence of spatial variability in the soil wetness field on the mean surface heat fluxes.

4.2. Sensitivity tests: the impact of soil wetness variability on the surface heat fluxes

4.2.1. Case I: test with constant spatial variability in the soil wetness field

When we run a simulation forward from a set of initial conditions using observed meteorological forcings, it is often difficult to separate a clear signal of the impact of changes in soil moisture variability on the surface fluxes from other effects (e.g. changing weather conditions, changes in the mean soil moisture condition, etc.). To isolate the effects of soil moisture variability from other effects, the Day 216 initial conditions and meteorological forcings were used as a baseline. To start with, reference runs were done

Fig. 12. Results of the Case I simulation. Plot of the mean (spatial average over the FIFE-89 testbed area; time-average of fluxes over 2 h centered on midday) evaporative fraction (EF) as a function of mean initial soil wetness. Continuous lines, with error bars showing standard deviations of EF , represent the results of the full-resolution runs (top, W , SR); dashed lines represent the results of runs where area-averaged initial conditions ($\langle \text{top} \rangle$, $\langle W \rangle$, $\langle SR \rangle$) were used. These results were obtained by taking the initial Day 216 soil wetness field as a reference (marked with an arrow) and incrementing or decrementing the wetness of each pixel by successive multiples of ± 0.05 . (a) EF vs. W_1 ; (b) EF vs. W_2 . (The small rise in EF as the soil wetness falls through $W_2 \approx 0.27$ is due to an artifact produced by the soil model in reaction to the imposition of an unrealistic soil wetness profile).

with fully resolved and area-averaged fields of topography (top), soil wetness (W) and SR for Day 216; the results of these runs are shown in Lines 1 and 8 of Table 1(a) and in Fig. 12, where the area-averaged evaporative fraction (EF) is plotted against the average soil wetnesses $\langle W_1 \rangle$ and $\langle W_2 \rangle$ for the testbed domain. Subsequent runs, which used both fully resolved (top, W , SR) and area-averaged ($\langle \text{top} \rangle$, $\langle W \rangle$, $\langle SR \rangle$) initial conditions, were performed using the same forcings and initial conditions with the sole exception that the entire soil wetness field, i.e. all pixel-level values of W_1 , W_2 and W_3 , were raised or decreased by a constant fraction in each run. For example, if we represent the wetness of a pixel on Day 216 by W^{216} , subsequent runs in the series would start with values for that pixel of $W^{216} - 0.05$, $W^{216} - 0.10$, $W^{216} - 0.15$, ..., and also with $W^{216} + 0.05$, $W^{216} + 0.10$, etc. This procedure has the effect of preserving the soil moisture variability and forcing

Fig. 13. Results of the Case II simulations (see facing page). Time series of FIFE-89 testbed simulations for Days 216–224. The continuous lines, with light dashed lines showing standard deviations, represent the results of full-resolution runs (top, W , SR); the long dashed lines show the results of the area-averaged ($\langle \text{top} \rangle$, $\langle W \rangle$, $\langle SR \rangle$) runs. In all cases, mean values for the entire testbed area are shown, time-averaged for 2 h centered on local midday. (a) $R_n - G$; (b) evaporative fraction (EF); (c) W_1 ; (d) W_2 ; (e) standard deviation of W_1 , σ_{W_1} , plotted against the area-average of W_1 , $\langle W_1 \rangle$. Observations are also shown: $\blacklozenge$ and filled error bars, surface flux stations in the testbed area; $\circ$ and filled error bars, airborne fluxes; diamonds with numbers, estimates of σ_{W_1} calculated from surface soil moisture observations derived from the PBMR data for each day. The light dotted lines in (c) and (d) show the half-hourly values of W_1 and W_2 , respectively.

conditions for the testbed area while adjusting the mean soil wetness. (Individual pixel values of W were not permitted to drop below 0.05 or exceed 1.0 during this test; the values of $\langle W \rangle$ plotted in Fig. 12 are the means taken over the W field following these bounding adjustments so that at low and high values of $\langle W \rangle$, successive increments and decrements of $\langle W \rangle$ may be slightly less than 0.05). Fig. 12 shows that for the Day 216 pattern of soil wetness, the fully resolved and area-averaged calculations of EF are close to each other at high $\langle W \rangle$ but diverge as $\langle W \rangle$ decreases. Fig. 7 showed why this is: the soil moisture stress function, $f(\psi_s)$, and the soil surface resistance function, r_{surf} , exert little control on evapotranspiration when soil moisture is freely available but have strong non-linear influences as W decreases.

The results in Fig. 12 (Case I) indicate that these non-linear effects become a problem when W drops below about 0.5, i.e. below the transpiration control breakpoint shown in Fig. 7. However, it should be remembered that the pattern of spatial variability for W in Case I was artificially preserved as W decreased in this test. Is this pattern conserved in nature?

4.2.2. Case II: changes in soil moisture variability during a real drydown

Fig. 13 (Case II) shows the result of running fully resolved and area-averaged versions of SiB forward from Day 216 to Day 224 using the observed time-series of meteorological forcings and starting with the Day 216 initial conditions. Fig. 13(a) and Fig. 13(b) show time-series of calculated available energy ($R_n - G$) and evaporative fractions (EF) over the period; Fig. 13(c) and Fig. 13(d) show the changes in the surface and root zone soil moisture. If anything, the variability in EF is conserved over time. Also shown in the figures are observations of some of these quantities obtained by averaging available surface and airborne heat flux measurements. Although the figure indicates that the model is performing credibly, Fig. 13(c) is particularly interesting: the spatial variability in the surface layer wetness, σ_{W_1} , is seen to decrease as W_1 decreases (see also Fig. 13(e), which shows the same data replotted for clarity). The likely explanation for this trend is that wetter areas tend to dry out and/or drain much quicker than drier ones which has the effect of continually narrowing the range of soil moisture contents within the domain as the drydown progresses. The results in Fig. 13 can be summarized as follows: (1) the model appears to simulate the changes in soil wetness and heat fluxes reasonably well over the FIFE-89 drydown period (Days 216–224); (2) as the domain dries out, the spatial variability of the surface soil wetness decreases.

4.2.3. Case III: changes in the surface energy balance and soil wetness fields during an idealized drydown

The results from Case II are contaminated by day-to-day changes in the meteorological

Fig. 14. Results of the Case III simulations. Time-series of FIFE-89 testbed simulations starting with Day 216 initial conditions and forced thereafter with repeated Day 216 meteorological and radiation data. The continuous lines, with error bars showing standard deviations, represent the results of the full-resolution runs (top, W , SR); the dashed lines show the results of the area-averaged ((top), $\langle W \rangle$, $\langle SR \rangle$) runs. In all cases, mean values for the entire testbed area are shown, time-averaged for 2 h centered on local midday. (a) $R_n - G$; (b) evaporative fraction (EF); (c) W_1 ; (d) W_2 ; (e) standard deviation of W_1 , σ_{W_1} , plotted against $\langle W_1 \rangle$ from the full-resolution run; (f) hourly depiction of W_1 (top, W , SR) for the first 7 days, again from the full-resolution run (note diurnal variation in W_1 owing to daytime evaporation and night-time recharge from below).

forcings; for example, the observed and calculated evaporative fractions in Fig. 13(b) rise and fall in response to the weather, partially obscuring the overall downward trend in EF that we would expect to see as the soil moisture decreases.

Fig. 14 (Case III) compares the results of fully resolved and area-averaged SiB runs starting from Day 216 initial conditions but forced by repeated cycles of the Day 216 meteorological conditions for every day thereafter. Fig. 14(a), (b) and (c) correspond to Fig. 13(a), (b) and (c). The time-series of energy balance components, the evaporative fraction EF and the soil wetness values calculated by the two methods track each other closely over the first part of this drydown, say from Day 216 to about Day 230. This phase corresponds to the soil moisture conditions covered in Table 1 and Case II. Over the next 15 days (Days 230–245), the evaporation rates diverge slightly as the root zone soil moisture drops through $W_2 \approx 0.5$. This is because of the non-linear effects of the $f(\psi_1)$ relationship shown in Fig. 7; the negative second derivative of this relationship as W_2 decreases from saturation insures that the full-resolution model will calculate higher transpiration rates. Thereafter, the two methods yield progressively convergent results as the soils dry further (Days 245–270). These results qualitatively agree with what we expect from the Case I results (Fig. 12), which indicated that the two methods would produce increasingly divergent surface flux estimates as W decreases. However, the reduction in σ_w as W_1 and W_2 decrease partially counteracts this divergence, with the net result that the two methods produce comparable time-series of area-averaged surface fluxes and soil wetnesses. For example, the maximum difference between the evapotranspiration rates delivered by the two methods in Case III amounts to just over 5% of the available energy (see Fig. 14(a)). This is less than half the maximum difference calculated in Case I (around 13%). Thus, for intermediate and low values of soil wetness, the Case III full-resolution and block calculations agree with each other far more closely than the equivalent Case I results. The reasons for this are shown clearly in Fig. 14(e), Fig. 15(a) and Fig. 15(b); as the soil dries,

Fig. 15. Changes in (a) canopy transpiration and (b) soil evaporation rates as the soil dries out in the extended Case III runs. Continuous lines and vertical bars denote results from the full-resolution run (means and standard deviations, respectively). Dashed lines denote mean fluxes generated by the block (area-average) runs.

Fig. 16. Variation in σ_{w_1} as a function of $\langle W_1 \rangle$ for Case I (long dashed line), Case II (dotted line) and Case III (continuous line). Estimates of σ_{w_1} obtained from analyzing the PBMR data are also shown labeled by day number; these should be compared with the Case II trajectory.

the variability in canopy transpiration and soil evaporation decrease along with the decrease in variability in soil wetness. This is in contrast to the effects produced by the constant spatial variability in soil wetness that was artificially maintained in Case I.

Fig. 16 shows the variation in σ_{w_1} as a function of W_1 for all three cases. The uppermost dashed line represents Case I where σ_{w_1} was effectively conserved (see Fig. 12). The dotted line and data points represent the modeled (full resolution) and observed (from the PBMR data) values of σ_{w_1} for Case II, the real drydown case for Days 216–224 shown in Fig. 13. The continuous line shows the results of Case III, the idealized (cyclically repeated Day 216 meteorological forcings) drydown. Clearly, Case I is unrealistic. On the other hand, the Case II and Case III results show qualitative agreement with the observed σ_{w_1} values.

From these results, we conclude that:

1. as the soil moisture field dries out, its spatial variability is reduced. This has the effect of reducing variabilities in the surface heat fluxes below levels they would maintain if σ_w were conserved.
2. Because of the above effects, the area-averaged and area-integrated estimates of the surface heat fluxes and soil wetness track each other fairly closely during a drydown.

5. Summary

Soil–vegetation–atmosphere transfer (SVAT) models are usually developed from formulations which are conceptualized and tested at local scales, a few meters to a few tens of meters. It is the accepted practice to apply these models directly as land surface

parameterizations (LSPs) in atmospheric general circulation models (GCMs), a procedure which implicitly assumes that the dependence of the land–atmosphere fluxes on critical land surface parameters, principally topography, vegetation condition and soil moisture, are scale-invariant. Analyses in the second section of this paper demonstrate that for this assumption of scale-invariance to hold true, either the models linking fluxes to surface conditions must be linear or nearly so, or the spatial variability in the critical surface initial or boundary conditions must be small. The analyses suggest that the negative impacts of area-averaging surface fields should be relatively small for topography and vegetation condition but could present some problems when dealing with soil moisture. This analysis provided the motivation for a numerical study of the issue using a modified version of the simple biosphere model (SiB) of Sellers et al. (1986) and the FIFE-89 data set.

The third section evaluated the performance of SiB using initial or boundary conditions (vegetation condition, represented by the conductance parameter ∇_F (from *SR*), soil moisture from airborne microwave and in situ data), meteorological forcings (from the automatic meteorological station network), and validation data (surface and airborne flux measurements) gathered for the 2 km $\times$ 15 km testbed area located in the center of the FIFE area. The model heat flux simulations agreed convincingly with available observations over a range of soil moisture conditions.

The fourth section addresses the central question of the paper: how does spatial variability in topography, vegetation conditions and soil wetness affect the area-averaged fluxes? The approach for investigating this issue was to perform ‘fully resolved’ calculations for every 30 m $\times$ 30 m pixel within the testbed area, i.e. for a total of 68 $\times$ 501 pixels. This 30 m grid is fine enough to resolve landscape-scale variability in the area; also, values for the fields can be assigned at this resolution from the available observations. These full-resolution runs were followed by similar runs in which one or more of the spatially varying fields was replaced by a domain-average value. It was found that the use of area-averaged fields of topography and the vegetation index, *SR*, had relatively small impacts on the mean fluxes. This is mainly because the relations linking topography and *SR* to the surface energy and heat fluxes are near-linear and so area-averaging and area-integration operations are mathematically almost equivalent. Some of these runs (Case I) indicated that if a large spatial variability in the soil wetness field could be maintained as the area dried out, the area-averaged and full-resolution (area-integrated) flux estimates would diverge significantly. This is because the relationships between soil moisture content

Fig. 17. Conceptual diagrams showing the relationships between the value of a surface initial or boundary condition and the evaporative fraction (*EF*). (a) Topography, represented by the absolute value of the slope; (b) vegetation condition, represented by variations in the canopy conductance parameter, ∇_F ; (c) soil wetness, *W*. In all three figures: ●, estimate of *EF* obtained from an area-averaged value of the initial or boundary condition; ○, area-integrated estimate of *EF*, which is partly a function of the standard deviation of the initial or boundary condition. The two estimates converge when the functions are linear or the standard deviations are small. It should be noted that, as the relationships with slope and ∇_F are more or less linear, the use of area-averaged values of these variables over an intermediate-scale (2–10 km²) domain has little impact on the area-averaged fluxes and *EF*. The relationship between *W* and *EF* is near-linear at high and very low values, so area-averaging *W* has little impact under these conditions. During a drydown, intermediate values of *W* are produced in the non-linear portion of the *W*–*EF* relationship; however, the dynamics of the drying process act to reduce the spatial variability of *W*, σ_w , so once again the impact of averaging *W* is small.

and soil and canopy evapotranspiration become increasingly non-linear as the soil dries. Further simulations tracked the dynamics of the soil wetness field during the FIFE-89 drydown (Days 216–224) for which comprehensive data sets were available. Both the simulations and observations indicated that as the soil wetness decreased, its spatial variability also decreased, with the result that the area-averaged and area-integrated calculations of the surface fluxes tracked each other closely as the drydown progressed. In short, when the soil is wet, the soil wetness field has high spatial variability but the evapotranspiration rate is insensitive to these variations for these conditions. By contrast, when the area-averaged soil wetness is low, its spatial variability is also low and so even though the soil wetness–evapotranspiration functions are highly non-linear under these conditions, their impact on the area-averaged fields is reduced. These findings are summarized in Fig. 17. In conclusion:

1. the relationships describing the effects of moderate topography on the surface radiation budget are near-linear and so are largely scale-invariant. The relationships linking the simple ratio vegetation index (*SR*), the canopy conductance parameter (∇_F) and the canopy transpiration flux are also near-linear and similarly scale-invariant to first order. Because of this, simple area-averaging operations can be applied to these fields with relatively little impact on the calculated fields of the surface heat fluxes.
2. The relationships linking surface and root zone soil wetness to the soil surface and canopy transpiration rates become increasingly non-linear as the soil dries out. However, simulation results and observations both showed that soil wetness variability decreases significantly as an area dries out, which partially cancels out the effects of these non-linear functions. As a result, the time-series of heat fluxes produced by fully resolved and area-averaged soil wetness fields track each other closely.

The combination of the effects described in (1) and (2) above indicate that we can use area-averaged values of topography, vegetation condition and soil wetness to describe fluxes on the scale of a few kilometers over a period of a few days. They also indicate that, for practical purposes, the biophysical evapotranspiration models in use for calculating these fluxes are relatively robust and scale-invariant with respect to these variables. It should be noted that these results apply only to the calculation of surface flux fields; we suspect that such simple area-averaging techniques will not work well in describing runoff rates, particularly under very wet soil moisture conditions which were not addressed here. How much confidence do we have that these findings translate to the larger scales associated with atmospheric models, i.e. 50–500 km? The results of this paper and those of Sellers et al. (1992c) indicate that simple area-averages of (moderate) topography and vegetation parameters are probably sufficient over a wide range of spatial scales provided that the vegetation type does not change radically within the domain; the work of Noilhan et al. (1991) and Townshend and Justice (1988, 1990) supports this view. However, the issue is not so straightforward with respect to soil moisture. To address it properly, it would be necessary to characterize the dynamics of the spatial variability of soil moisture and atmospheric conditions at these larger scales. If we see the same trends of decreasing spatial variability in soil wetness with decreasing area-averaged wetness, then the entire biophysical modeling approach can be treated as effectively scale-invariant over the spatial scale range of a few meters to a few tens or hundreds of kilometers. If this

turns out not to be the case, then scale-dependent area-averaged forms of the equations relating soil moisture to evaporation will have to be developed which take these effects into account. However, it must be said that the results of this study suggest that the impact of using area-averaged values of soil moisture, vegetation cover and topography on the calculated fluxes of sensible and latent heat are much less than were originally feared by the authors. This finding in turn suggests that inaccuracies associated with the assumption of biophysical model scale-invariance are likely to be small. However, further work is needed to explore the generality of these findings for areas with different kinds of vegetation cover, soil properties and topography.

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