The ISBA land surface parameterisation scheme

Joël Noilhan, Jean-Francois Mahfouf

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J. Noilhan, J.-F. Mahfouf

Météo-France/CNRM, 42 Avenue Coriolis, 31057 Toulouse Cedex, France

Abstract

An updated description of the land surface scheme ISBA (Interaction Soil Biosphere Atmosphere) is given. Recent improvements mostly concern the evaporation from bare soil, the inclusion of gravitational drainage, a continuous formulation of soil transfer coefficients for heat and moisture, and a modification of the surface drag coefficients. Validations of this scheme undertaken at local scale using data from various field campaign experiments are summarized. The ISBA scheme has been implemented in mesoscale and global scale models. Mesoscale modeling was used to define effective parameters taking into account sub-grid scale heteorogeneities of the surface. Finally, developments on soil moisture initialization for potential applications to numerical weather prediction are described.

1. General presentation

This paper presents an updated description of the land-surface parameterisation scheme ISBA (Interactions Soil Biosphere Atmosphere) developed by Noilhan and Planton (1989). This scheme describes the exchanges of heat and water between the low-level atmosphere, the vegetation and the soil. As ISBA was designed for meteorological models, it is a relatively simple scheme, but it embraces the most important components of the land surface processes. The simplicity of the scheme is achieved by the calibration of several important coefficients with more sophisticated models and experimental data.

The scheme includes the treatment of soil heat content, soil water content, water interception by vegetation and aerodynamic transfer processes in the atmospheric surface layer. The scheme uses the force restore model for soil heat and water content (Blackadar, 1976; Deardorff, 1977) and the α method for evaporation (Mahfouf and Noilhan, 1991). Other important features of the scheme include a representation of gravitational drainage (Mahfouf and Noilhan, 1996), the introduction of soil/vegetation heat capacity C_T and the calibration of the force restore coefficients (for water content) with a multi-layer soil water model

After describing the main features of the scheme, the various studies conducted with ISBA applied at the local, meso-scale and global scales are summarized.

*This original version of the published paper in *Global and Planetary Change* contains a number of errors corrected in a companion document.

2. Prognostic equations and surface fluxes

The scheme incudes a representation of snow which is not described here (Douville et al., 1995). There are five prognostic equations (list of symbols in Table 1) for deep soil temperature, T_2, deep soil water content, w_2, surface soil/vegetation temperature, T_s, top soil water content, w_g and interception water storage, W_r.

$$\frac{\partial T_s}{\partial t} = C_T(R_n - H - LE) - \frac{2\pi}{\tau}(T_s - T_2) \quad (1)$$

$$\frac{\partial T_2}{\partial t} = \frac{1}{\tau}(T_s - T_2) \quad (2)$$

$$\frac{\partial w_g}{\partial t} = \frac{C_1}{\rho_w d_1}(P_g - E_g) - \frac{C_2}{\tau}(w_g - w_{geq}); \quad 0 \leq w_g \leq w_{sat} \quad (3)$$

$$\frac{\partial w_2}{\partial t} = \frac{1}{\rho_w d_2}(P_g - E_g - E_{tr}) - \frac{C_3}{d_2 \tau}\max[0., (w_2 - w_{fc})]; \quad 0 \leq w_2 \leq w_{sat} \quad (4)$$

$$\frac{\partial W_r}{\partial t} = vegP - (E_v - E_{tr}) - R_r; \quad 0 \leq W_r \leq W_{rmax} \quad (5)$$

where veg is the fractional vegetation cover. A list of symbols is given in Table 1 for an explicit definition of other parameters.

Only one energy balance is considered for the whole system ground-vegetation. As a result, heat and mass transfers between the surface and the atmosphere are related to the mean values T_s and w_g.

The net radiation at the surface is the sum of the absorbed fractions of the incoming solar radiation R_G and of the atmospheric infrared radiation R_A, reduced by the emitted infrared radiation:

$$R_n = R_G(1 - \alpha_t) + \epsilon_t(R_A - \sigma T_s^4) = H + LE + G \quad (6)$$

where σ is the Stefan-Boltzmann constant, H the sensible heat flux, LE the latent heat flux, and G the ground heat flux.

The turbulent fluxes are calculated by means of the classical aerodynamic formulae (Louis, 1979). For the sensible heat flux:

$$H = \rho_a C_p C_H V_a (T_s - T_a) \quad (7)$$

where C_p is the specific heat of air; ρ_a, V_a, and T_a are respectively the air density, the wind speed, and the temperature at the lowest atmospheric level; and C_H is the drag coefficient depending upon thermal stability of the atmosphere. The water vapor flux E is the sum of the evaporation from the soil surface (i.e., E_g) and from the vegetation (i.e., E_v):

$$LE = LE_g + LE_v \quad (8)$$

$$E_g = (1 - veg)\, \rho_a C_H V_a (h_u q_{sat}(T_s) - q_a) \quad (9)$$

$$E_v = veg\, \rho_a C_H V_a h_v [q_{sat}(T_s) - q_a] \quad (10)$$

where L is the latent heat of vaporization, $q_{sat}(T_s)$ is the saturated specific humidity at the temperature T_s,

Table 1: List of symbols

Surface temperature	T_s
Deep soil temperature	T_2
Top-soil volumetric water content	w_g
Deep soil volumetric water content	w_2
Intercepted water by the canopy	W_R
Precipitation at screen and ground level	P, P_g
Net radiation	R_n
Incoming solar radiation	R_G
Soil heat flux	G
Sensible heat flux	H
Total evapotranspiration	E
Evapotranspiration from the vegetation	E_v
Transpiration	E_{tr}
Surface evaporation from bare soil	E_g
Evaporation from intercepted reservoir	E_r
Force restore coefficients for soil moisture	C_1, C_2
Coefficient for gravitational drainage	C_3
Surface soil/vegetation heat capacity	C_T
Density of liquid water	ρ_w
Depths of the top and deep soil	d_1, d_2
Time constant of one day	τ
Saturated volumetric water content	w_{sat}
Field capacity volumetric water content	w_{fc}
Wilting point volumetric water content	w_{wilt}
Slope of the retention curve	b
Soil thermal coefficient at saturation	$C_{G_{sat}}$
Value of C_1 at saturation	C_{1sat}
Value of C_2 at $w_2 = 0.5w_{sat}$	C_{2ref}
Coefficients of w_{geq} formulation	a, p
Depth of soil column	d_2
Minimum leaf surface resistance	R_{smin}
Leaf area index	LAI
Vegetation cover	veg
Roughness length for momentum	z_0
Surface albedo	α_t
Emissivity	ϵ_t

and q_a is the atmospheric specific humidity at the lowest atmospheric level.

The relative humidity h_u at the ground surface is related to the superficial soil moisture w_g following:

$$h_u = \frac{1}{2}\left[1 - \cos\left(\frac{w_g}{w_{fc}}\pi\right)\right], \text{ if } w_g < w_{fc} \quad (11)$$

$$h_u = 1, \text{ if } w_g \geq w_{fc} \quad (12)$$

where w_{fc} is the volumetric soil water content at the field capacity.

In case of dew flux when $q_{sat}(T_s) < q_a$, h_u is also set to 1 (see Mahfouf and Noilhan (1991) for details). When the flux E_v is upward from the surface to the atmosphere, the Halstead coefficient h_v takes into account the direct evaporation E_r from the fraction δ of the foliage covered by intercepted water, as well as the transpiration E_{tr} of the remaining part of the leaves:

$$h_v = (1-\delta)R_s/(R_a + R_s) + \delta \quad (13)$$

$$E_r = veg\frac{\delta}{R_a}[q_{sat}(T_s) - q_a] \quad (14)$$

$$E_{tr} = veg\frac{1-\delta}{R_a + R_s}[q_{sat}(T_s) - q_a] \quad (15)$$

When E_v is downward from the atmosphere to the surface, the dew flux is supposed to occur at the potential rate, and h_v is taken equal to 1.

Following Deardorff (1978), δ is a power function of the moisture content of the interception reservoir:

$$\delta = (W_r/W_{rmax})^{2/3} \quad (16)$$

The aerodynamic resistance is $R_a = (C_H V_a)^{-1}$. The canopy surface resistance, R_s, depends upon both atmospheric factors and available water in the soil.

The expressions for the various coefficients of ISBA are given in the Appendices.

With respect to the initial description of ISBA given by Noilhan and Planton (1989), differences concern:

- the inclusion of gravitational drainage in Eq. (4)
- the use of w_{fc} from the definition of Wetzel and Chang (1987) as threshold for bare soil evaporation in Eq. (11) instead of $0.75 w_{sat}$.
- the continuous formulation of soil moisture transfer coefficients (Appendix A).
- the modification of the drag surface coefficients in order to consider different roughness length values for heat and momentum.
- the improvement of C_1 formulation for low soil water content.

Table 2: A summary of ISBA validation at the "local" scale

Reference	Data	Temporal scale	Vegetation/soil type
Noilhan and Planton (1989)	HAPEX-MOBILHY 86	clear days	Crops/loam-sand
Jacquemin and Noilhan (1990)	HAPEX-MOBILHY 86	clear days	Pine forest/sand
Mahfouf and Jacquemin (1989)	HAPEX-MOBILHY 86	rainy days	Crops/loam-sand
Mahfouf (1990)	HAPEX-MOBILHY 86	month	Crops/loam-sand
Mahfouf and Noilhan (1996)	HAPEX-MOBILHY 86	one year	Soya/loam
Mahfouf and Noilhan (1991)	AVIGNON 84	one week	Bare ground/loam
Braud et al. (1993)	EFEDA 91	two weeks	Bare ground/silt loam
Giordani et al. (1996)	EFEDA 91	one month	Vine/silt-loam
Germain (1990)	FIFE 87	clear days	Grassland/loam
Noilhan et al. (1993)	ARME 84	two years	Amaz. forest/s. clay-loam
Mahfouf and Noilhan (1994)	HAPEX SAHEL 92	clear days	Savanna/sand
Goutorbe et al. (1997)	HAPEX SAHEL 92	two months	Savanna/sand

3. Validations at the local scale

Most of the validation tests were performed in a stand-alone model, by constraining the atmospheric conditions (temperature, humidity, wind speed and radiative forcing) to follow the observations. Because of the major influence of the initial soil moisture content for daily prediction of surface fluxes, the soil water content at the beginning of the simulations was prescribed from available observations.

Series of tests were run to investigate the ability of the surface scheme to reproduce the partitioning of available energy for constrasted meteorological conditions (clear sky and rainy days) and for various land surface covers (bare ground, grasslands, crops and forest). Table 2 gives a summary of these experiments and indicates some specific points which were examined.

3.1 Exchanges over bare ground

Special attention has been paid to exchanges over bare ground using AVIGNON 84 and EFEDA 91 data sets. Of particular interest is the magnitude of day-time soil heat flux G which may represent a significant fraction of the net radiation ($\approx$ 30%), particularly in dry conditions. Because the local measurements of thermal properties were directly introduced in the surface scheme, the predictions of G were in fair agreement with the observations (for simulations lasting one week). Previous run using the parameterisation of thermal conductivity proposed by Al Nakshabandi and Kohnke (1965) provided too large values of G by about 20%.

Another important feature concerning heat exchanges over bare ground is the significant departure between the roughness length for momentum z_{om} and for heat z_{0h} transfers. Although confirmed from observations since many years (Garratt and Hicks, 1973), few land-surface schemes distinguish between these

two roughness lengths. Observations show that over natural surfaces the ratio z_{0m}/z_{0h} is close to 10. Such an assumption was tested with the EFEDA data set at Barrax location (Braud et al., 1993; Giordani et al., 1996) leading to satisfactory prediction of the sensible heat flux H. Using the classical assumption $z_{0m} = z_{0h}$, the sensible heat flux was clearly underestimated.

The formulation of surface evaporation in simple parameterisation schemes is also an important issue. Indeed, evaporation over bare ground is a very complex mechanism because liquid water is not generally available at the surface but evaporation occurs at a certain depth depending on the near-surface soil moisture. Accurate prediction of surface evaporation requires the use of detailed multi-layer soil models with very high spatial and temporal resolutions. In the parameterisation context, the challenge is to reproduce the surface hydrology with a limited number of soil variables (e.g. the surface temperature and the water content of a thin superficial layer) and surface parameters. Most of evaporation formulations in surface schemes are based on bulk aerodynamic expressions by using a parameterisation of the surface specific humidity (Deardorff, 1978; Kondo et al., 1990). Some parameterisations make use of threshold methods based on the concept of water supply and demand close to the surface (Mahrt and Pan, 1984; Dickinson, 1984; Wetzel and Chang, 1987). In this last approach, considerable information about hydraulic properties in a thin layer is needed, making them very sensitive to the averaging depth to which the near-surface water content is known (Abramopoulos et al., 1988). Using AVIGNON 84 data set, Mahfouf and Noilhan (1991) examined various formulations of surface evaporation. A series of tests in stand-alone model were run, starting from observed soil water contents and changing only the formulation of surface evaporation. In the context of these observations, the diffusion-limiting moisture flux was strongly underestimated by threshold methods. On the other hand, the most widely used bulk aerodynamic formulations (so-called α- and β-methods) provided comparable results during day-time, whereas the β-methods overestimated the nocturnal evaporation.

In the case of very dry soils, effect of vapor phase diffusion have to be taken into account in the time variation of superficial soil moisture. A specific investigation of heat and water exchanges for very dry soils was performed by Braud et al. (1993) and Giordani et al. (1996) who improved ISBA by taking into account soil moisture transfers in vapor phase (see expression for the coefficient C_1 in the Appendix A). This approach was tested favourably against observations of surface evaporation and near-surface water content collected during EFEDA.

3.2 Rainfall interception

Rainfall interception by vegetative canopies modifies the surface water balance because water evaporates back into the atmosphere at potential rate. Interception loss can be significantly higher than transpiration rate, especially over forest canopies because of their high value of roughness length. Forests have also a significant water storage capacity (several millimeters of water can be retained on the foliage leaf surfaces) as compared to short vegetation canopies.

A sensitivity study of ISBA by Mahfouf and Jacquemin (1989) confirmed the importance of high values of

roughness length in reinforcing the evaporation of intercepted rainfall. The drying time of the interception reservoir is closely linked to the reservoir capacity and therefore to the value of LAI. Validations of interception schemes have mainly been carried out over forest canopies for which long enough time series of precipitation and throughfall were recorded (Gash et al., 1980; Lloyd et al., 1988). The interception data obtained continuously during 25 months in the Reserva Ducke during ARME 84 have been used by Noilhan et al. (1993) to compare four models (Rutter et al. (1971); Gash (1979); SiB and ISBA models) applied to the same forcing during two years. The ISBA integration provided comparable results with the Rutter model which can be considered as a reference for rainfall interception. On the other hand, ISBA predictions were larger than SiB results with comparable conditions. The difference might be due to different formulation of the aerodynamic resistance in the two models. For short canopies, direct measurement of interception is nearly impossible. However, evidence of re-evaporation of water retained by leaves can be proved from measurements of Bowen ratio. As an example,the HAPEX-MOBILHY data set was used by Mahfouf and Jacquemin (1989) to analyse the ISBA simulations of interception processes by crops fields.

3.3 Plant transpiration and soil water budget

Plant transpiration is probably the main issue of Soil Vegetation Atmosphere Transfer Schemes because of its major influence on deep soil moisture. Water extraction from the rooting system depends on vegetation physiology which is largely driven by environmental conditions and soil moisture availability: extraction of water is enhanced in conditions of full radiation, moderate atmospheric specific humidity deficit, optimum air temperature and freely available soil water in the rooting zone. As in many surface schemes, the general approach of Jarvis (1976) is adopted by ISBA for the expression of R_s (Appendix B).

Because ISBA does not takes into account processes inside the canopy, the parametersation of R_s attempts to describe the whole canopy. The minimum surface resistance R_{smin} is depending on the plant type. Scaling up R_s to the whole canopy is simply attempted by attenuation of R_s with increasing leaf area index. Typically, ISBA calibration studies gave R_{smin} of the order of 40 s/m for a closed live crop canopy (Noilhan and Planton, 1989; Jacquemin and Noilhan, 1990; Mahfouf, 1990), reaching higher values $\approx$ 500 s/m when plants/crops experience maturation (Noilhan and Planton, 1989) or senescence (Germain, 1990). A value of $R_{smin} = 100$ s/m was calibrated during HAPEX-MOBILHY over Les Landes pine forest for $LAI =$ 2.3 m^2/m^2 (Gash et al., 1989), while $R_{smin} = 250$ s/m was found for the Amazon forest ($LAI = 5$ m^2/m^2) by Manzi and Planton (1994). In these two cases of forested canopies, it is interesting to observe that the ratio R_{smin}/LAI was nearly constant.

The function $F_1(R_G)$ takes into account the effect of photosynthetically active radiation (PAR) assumed to be 55 % of the global radiation R_G (Eq. 38 in Appendix B). Noilhan et al. (1993) showed that the methods used in the ISBA and SiB models provide comparable results, despite the extremely simplified calculation in ISBA (particularly for radiative transfers in the canopy).

The function F_3 accounts for stomata closure in conditions of dry environment, a process particularly

important for forest canopies (Eq. 40 in Appendix B). The coefficient γ in F_3 has been calibrated. For example, the R_s dependence of F_3 is the most important environmental constraint for the transpiration of the Amazon forest and SiB calibration for the Amazon tropical forest provided $\gamma = 0.0273$ hPa^{-1}. This strong dependence was also observed for the Landes forest during HAPEX-MOBILHY, involving a lower latent heat flux over the pine forest than the nearby crop area. Jacquemin and Noilhan (1990) proposed $\gamma = 0.037$ hPa^{-1} for the Landes forest, in agreement with observations (Gash et al., 1989; Stewart, 1988). Such calibration of F_3 over Les Landes forest (Jacquemin and Noilhan, 1990) leads to a fair prediction of the canopy surface conductance estimated for the entire data set. The vapor deficit term F_3 seems less important for short canopies. However, it was found particularly significant for the prairie grass canopy in FIFE 87 (Sellers et al., 1988). Thanks to this effect, Germain (1990) calibrated ISBA for 6 selected FIFE 87 "golden" days and for 7 flux stations.

The function $F_4(T_s)$ used in ISBA follows the proposal of Dickinson (1984) (Eq. 41 in Appendix B). The parabolic shape of $F_4(T_s)$ is centered around an optimum air temperature of 25°C when the stress is negligible. Transpiration stops when $T_a = 0$°C or $T_a = 50$°C, an assumption which sounds valid for temperate latitudes. So far no clear validation of $F_4(T_s)$ has been obtained with ISBA.

The function $F_2(soil\ moisture)$ expresses the plant stress versus soil moisture in the root zone (Eq. 39 in Appendix B). In many land surface parameterizations, this stress factor is computed from the leaf water potential which depends on plant physiology and soil moisture. The threshold values of w_{wilt} and w_{fc} are computed from soil texture. The water content at wilting point w_{wilt} is calculated for a soil moisture potential of -15 bar below which plants cannot extract any more water from the soil matrix (see table I in Jacquemin and Noilhan (1990)). The water content at field capacity w_{fc} corresponds to a low value of hydraulic conductivity (0.1 mm/day according to Wetzel and Chang (1987)). The main difficulty to calibrate the soil water stress function is to find suitable data set, i.e. long enough time series of micro-meteorological quantities together with soil moisture and hydraulic properties. These observations were collected during the HAPEX-MOBILHY Special Observing Period of 3 months (Goutorbe et al., 1989). Although the soil water stress was moderate during this period, this data was used by Mahfouf (1990) to evaluate the prediction of soil moisture by ISBA for periods of one to two months in crop fields. Mahfouf (1990) showed that the predictions of the components of the water budget agree reasonably well with observations. In the course of ISBA development, the result of Mahfouf (1990) was particularly important since it gave substantial confidence to the ISBA functioning when used over long time periods.

The one year time series of soil moisture from HAPEX-MOBILHY devised for the PILPS-RICE Workshop has highlighted the need for a gravitational drainage at the bottom of the soil layer to be included in ISBA (Mahfouf and Noilhan, 1996). When this term is omitted, soil moisture is strongly overestimated in the winter time period (Fig. 1).

A second example of calibration of ISBA against observations is provided in Tables 3 and 4 and in Figs. 2-4. The data were collected during the EFEDA field program (Bolle et al., 1993) in a semi-arid area. Three

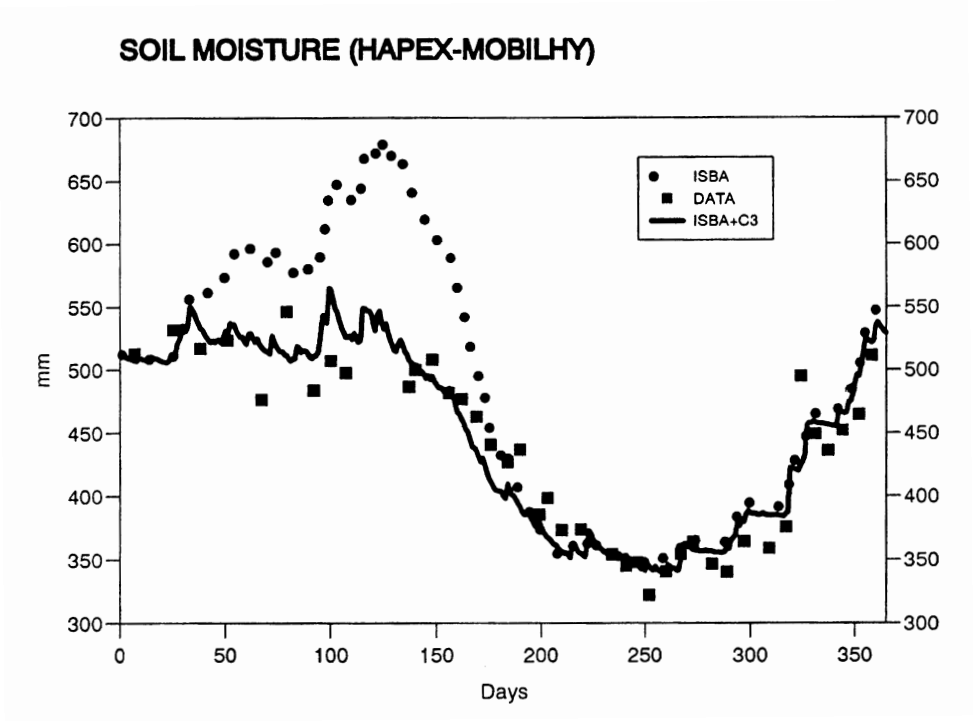

Figure 1: Daily averages of total soil water (mm) in the top 1.6 m simulated by ISBA with (solid line) and without drainage (dotted line) and in comparison with HAPEX data (solid square)

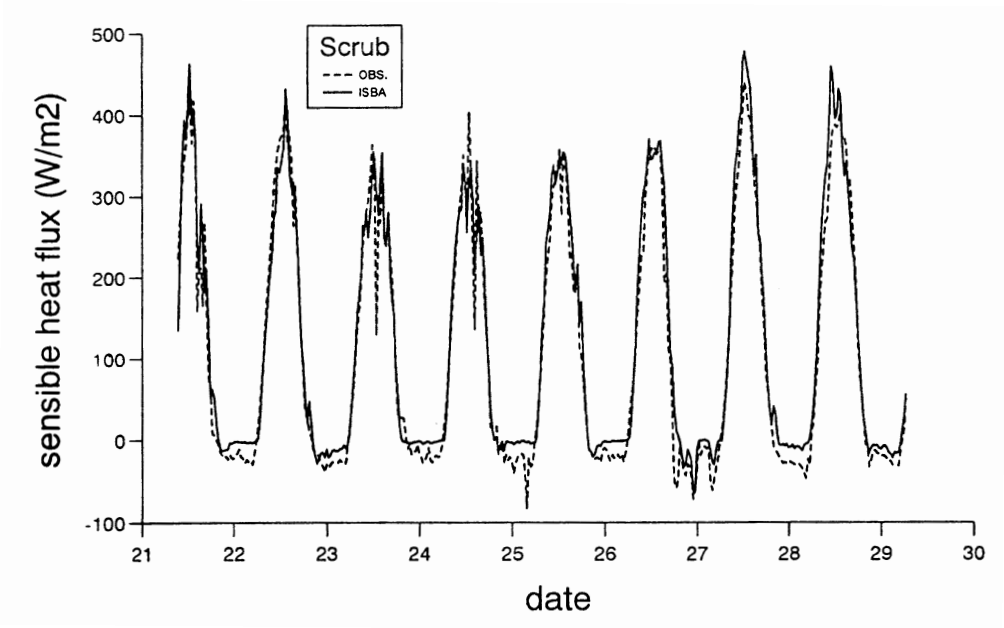

Figure 2: Daily variations of observed (dotted line) and simulated (solid line) sensible heat fluxes at the SCRUB super-site of the EFEDA 1991 field experiment (Bolle et al., 1993)

Table 3: Soil and vegetation parameters for the EFEDA sites

Parameter	Vineyard (Tomelloso)	Small scrub (Belmonte)	Bare ground (Barrax)
Period	June 4-30	June 21-29	June 19-30
Z_0 (m)	**0.01-0.05**	**0.02**	**0.01**
Z_{0h} (m)	$Z_0/20$	$Z_0/10$	$Z_0/10$
veg	**0.05** (June 1) - **0.15** (June 30)	0.4	0.0
LAI (m^2/m^2)	**0.1** (June 1) - **0.4** (June 30)	1.5	-
R_{smin} (s/m)	140	120	-
Soil type	**3**	**5**	**4**
λ (W/mK)	**0.3**	**1.0**	**0.42**
d_2 (m)	**0.70**	**1.4**	**0.7**

Table 4: Mean surface fluxes (W/m^2) and surface temperature (°C) computed (ISBA) and observed at the three EFEDA sites. The rms of the errors of flux simulations are also given

	exp.	H	rms_H	LE	rms_{LE}	G	rms_G	Rn	rms_{Rn}	T_s
ISBA	vine	95	28	24	31	2	35	122	19	29.8
OBS	vine	87		28		14		129		29.2
ISBA	scrub	119	31	38	28	0	39	157	13	30.4
OBS	scrub	107		43		12		162		29.7
ISBA	bare	96	37	57	57	0	24	153	11	28.1
OBS	bare	106		32		13		151		30.3

data sets representative of the experimental zone were prepared in order to calibrate land-surface schemes for sparse vegetation canopy under dry soil conditions. Low-level atmospheric quantities as well as surface fluxes were measured during nearly one month (June, 1991) over plots of small-natural scrub, vineyard and bare ground. Table 3 summarizes the values of the parameters prescribed to ISBA for the simulation of the three data sets. The bold parameters were observed (e.g. roughness lengths for momentum, fractional coverage and leaf area index for the vine, total albedo). The soil type corresponds to the vertically-averaged textural profile measured at each sites. The soil parameters were computed from the sand and clay contents associated with the soil type (Appendix A), except for the thermal conductivity because in-situ observations were available. The soil depth d_2 corresponds to the depth of soil moisture measurements. The initial values of soil moistures and temperatures in ISBA were taken from observations. Some aspects of this calibration work can be found in Braud et al. (1993) and Giordani et al. (1996). As depicted in Figs. 2-4, the sensible heat flux was very high reaching 400 W/m^2 around midday the last days of June as a result of very dry soils.

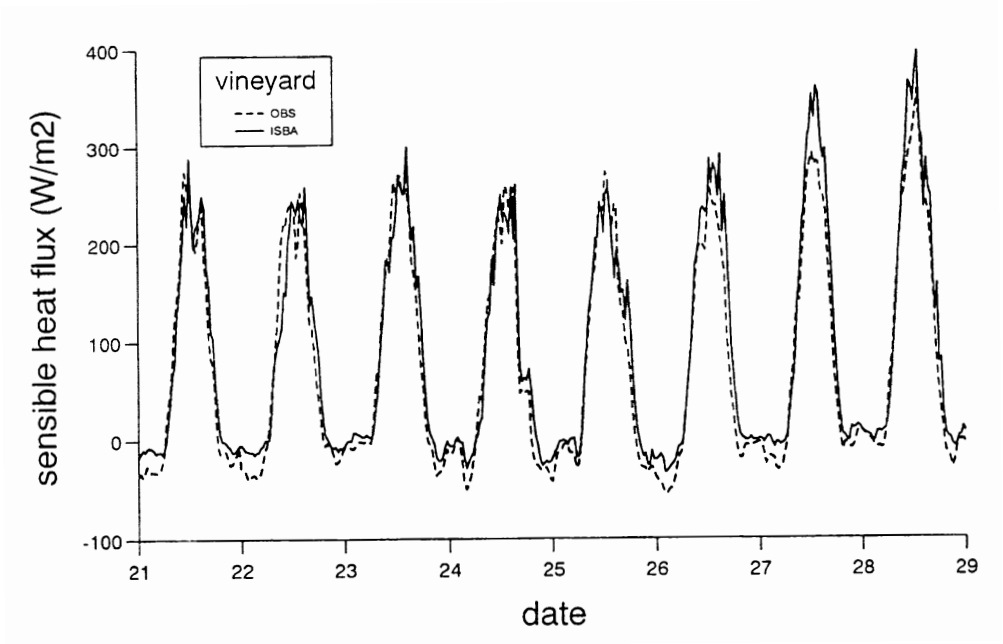

Figure 3: Daily variations of observed (dotted line) and simulated (solid line) sensible heat fluxes at the VINEYARD super-site of the EFEDA 1991 field experiment (Bolle et al., 1993)

The ISBA scheme was found able to simulate correctly the general patterns of the surface fluxes for the three types of surfaces. In the cases of the sparse-vegetation canopies (vine and scrub) the parameter R_{smin} was tuned to simulate the Bowen ratio of the two first days of the calibrations and then, kept unchanged during the rest of the runs. In the bare ground case, the inclusion of a simple representation of vapor transfers in the soil improved significantly the simulation. Finally, it was found that the revised scheme was able to simulate properly the surface energy balance of such complex canopies despite the use of a single surface temperature for the soil and vegetation medium. The performance of ISBA in terms of mean values and rms of the errors of surface flux simulations can be assessed in Table 4 ($rms_x^2 = \overline{(x_{ISBA} - x_{obs})^2}$). On a monthly basis, the rms for latent and sensible heat flux simulations exceeds 20 W/m^2, reaching high values in the bare ground case. These errors are rather large but they are comparable to the simulation errors produced by more complex schemes tested with the same data sets (see Linder et al. (1996) for details).

As a result of the experiments briefly presented in this section, some parameters of the scheme were adjusted. Since significant changes were included relative to the initial version (drainage, vapor phase transfers within very dry soil, continuous formulation of soil parameters), we think now that there is no basis to complicate ISBA any more in the context of the chosen philosophy for the treatment of heat and moisture transfers (force-restore method).

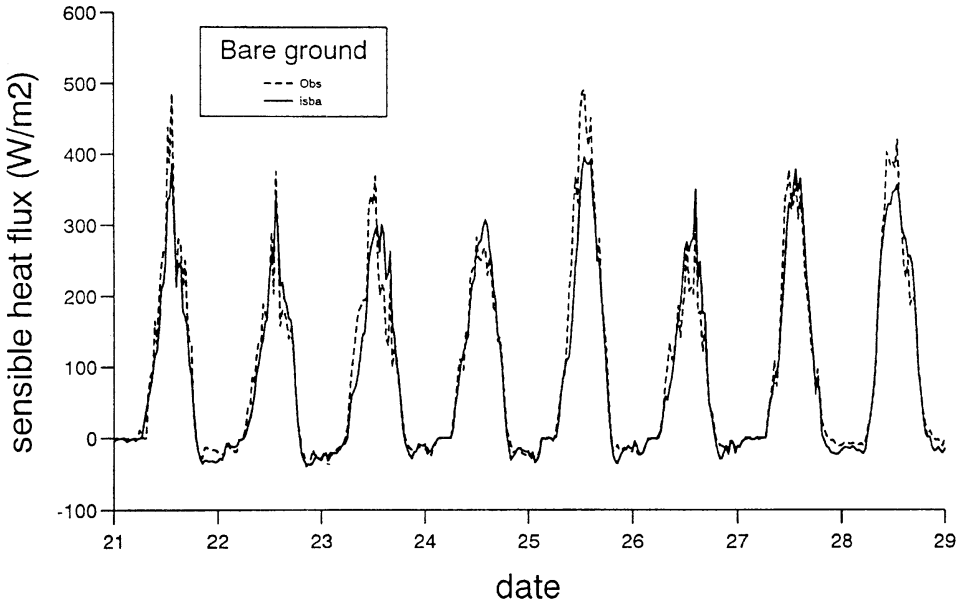

Figure 4: Daily variations of observed (dotted line) and simulated (solid line) sensible heat fluxes at the BAREGROUND super-site of the EFEDA 1991 field experiment (Bolle et al., 1993)

4. Integrations at meso-scale and global scale

4.1 Meso-scale integrations and parameter aggregation over large area

A mesoscale modeling strategy has been developed (André and Bougeault, 1988; André et al., 1989; Bougeault et al., 1991b,a; Noilhan et al., 1991c,a) to provide an integration tool as well as a useful framework for the interpretation of the experimental results collected during HAPEX-MOBILHY 1986 and EFEDA 91. Briefly summarized the modeling work was as follows:

- Use the results of the 1D calibration tests to setup a general table of possible values of the surface parameters and to produce maps of these parameters within the 3D simulation domains from detailed classification of both soil and vegetation types (Phulpin et al., 1989; Noilhan et al., 1991b).
- Integrate the model and compare the results with all available data (surface network, radiosonde and aircraft fluxes measurements).

Several one-day integrations for clear days have been carried out showing clearly the impact of surface processes on the prediction of the lower atmospheric fields at the regional scale. Thanks to an improved representation of the surface processes, the mesoscale model was able to reproduce the evidence of land-breeze between a Pine forest and surrounding crop fields (HAPEX-MOBILHY 86) as well as sea-breeze circulation observed during EFEDA 91 (Habets et al., 1994).

The mesoscale results have been used to address the issue of spatial variability in land surface properties from horizontal scales ranging from 10 km, a possible resolution for mesoscale models, to 100 km, the size of a GCM grid box. Noilhan and Lacarrère (1995) discussed methods for estimating effective vegetation

properties and soil characteristics of a large area from the observations and the 3D numerical work carried out with the HAPEX-MOBILHY data base. The averaging operators derived to compute effective vegetation properties were chosen in order to be consistent with an arithmetic averaging of the fluxes themselves. Aggregation of soil types was realized from continuous relationships relating the hydraulic properties to either the percent of clay and sand estimated over the large area (see Appendix A). Despite the non-linear dependence of surface fluxes on both vegetation and soil water content, it is found that the effective surface fluxes computed from effective parameters with a 1D column model match the spatial-averaged fluxes estimated from 3D mesoscale model results with a relative error less that 10%. On the other hand, fluxes computed with prescribed surface properties associated to the dominant land-use of the large domain depart significantly from the averaged fluxes. For the cases examined, the effects of non-linearity are found to be smaller for the vegetation behavior than for the soil water transfers.

The parameter aggregation method has been tested successfully for a long time period within the context of a 1D GCM grid cell representing the HAPEX-MOBILHY 1986 instrumented area. Given precipitation and solar fluxes, predictions of soil water content and total evaporation for 25 days compare well with observations available within the large area.

A similar approach was applied to the subgrid distribution of precipitation and interception from mesoscale results obtained during a rainy day from HAPEX-MOBILHY 86 (Blyth et al., 1994). An aggregation scheme was proposed accounting for sub-grid interception loss based on the knowledge of the fraction S of the grid concerned by precipitation. The fraction S, computed in this study by the mesoscale model, should be estimated in GCMs with respect to this problem.

4.2 Implementing ISBA with a GCM

Finally, the ISBA scheme has been implemented in the AGCM (Atmospheric General Circulation Model) EMERAUDE (Manzi and Planton, 1994) and ARPEGE (Mahfouf et al., 1995) developed at Météo-France. This coupled ISBA/AGCM is a powerful tool to investigate natural or anthropogenic climate changes in continental surfaces, like desertification or deforestation processes in the tropical regions. In this context the 1D tests performed with ARME data set have provided a very valuable check of the representation by the model of the physical processes in the particular case of the tropical rain forest.

The next step was the mapping of the scheme parameters at the AGCM grid scale. Following the general principles stated in sections 3 and 4.1, these maps have been obtained from basic classifications of soil and vegetation types with their spatial distribution at the earth surface with a 1° horizontal resolution (Wilson and Henderson-Sellers, 1985). Each secondary parameter of the scheme has been inferred from these types by means of correspondence table which accounts either of 1D calibration tests or of the bibliography. A parameter dependent averaging process has been applied to represent sub-grid variability. Mahfouf et al. (1995) performed a 3-year integration for the present-day climate with T42L30 version of the ARPEGE climate model. Results obtained were compared favourably to observed climatologies related to the various

components of the continental surface energy and water balances. Differences noticed come most often from a poor simulation of the precipitation field. However, some differences suggest specific improvements on ISBA to be brought concerning the representation of bare soil albedo, surface runoff and soil moisture initialization.

5. Assimilation of soil moisture for NWP models

Because of the slow time scale of water content, any erroneous initialization of soil moisture in Numerical Weather Prediction (NWP) models will cause a poor evaluation of surface exchanges. This justifies the need for an initialization scheme of soil moisture for short-to-medium range atmospheric forecast. A sequential assimilation based on optimum interpolation was tested by (Mahfouf, 1991; Bouttier et al., 1993a,b) with ISBA. The method relates linearly near-surface atmospheric forecast errors (air temperature and relative humidity) to soil moisture increments for both the superficial and deep reservoirs. One-dimensional sensitivity and validation studies Mahfouf (1991) and Bouttier et al. (1993a,b) showed that screen-level atmospheric quantities contain a lot of information about the soil-vegetation-atmosphere interactions as described by ISBA. The most critical aspect of the algorithm is the formulation of the optimum coefficients relating atmospheric forecast errors to soil moisture corrections, depending mainly upon local time and surface characteristics (vegetation coverage and soil texture). A continuous formulation described in Bouttier et al. (1993a) allows for an easy computation of these coefficients at each grid point of a three-dimensional atmospheric model when enough information about surface properties is available.

As a first attempt to test the method with a 3-D model, a 48-hour clear-sky period from the HAPEX-MOBILHY experiment was selected. The results of the assimilation demonstrated that the convergence of the method is rapid either with a dry or wet soil moisture guess. However, some limitations of the method were observed for places where the atmosphere above the point of interest do not contain information about soil moisture. Such situations occured when mesoscale circulations generated by large horizontal advections, at lateral boundary conditions with prescribed atmospheric parameters from analysis, and at grid points with isolated surface properties influenced by their surrounding environment. These limitations are due to the one-dimensional nature of the assimilation process and are to be overcome in order to avoid erroneous corrections of soil moisture when atmospheric forecast errors do not come from errors on surface processes representation. This study opens the way for an operational use of this technique, even if more efficient variational techniques should also be developed in the future.

6. Concluding summary

This paper described the various steps involved in the validation of the ISBA land-surface scheme for numerical weather prediction and climate simulation studies. Special attention is devoted to the validation of the various components of the Interaction Soil-Biosphere-Atmosphere (ISBA) scheme at the local scale. Because

relevant validation data set are mainly available at this scale, very much can be learned on the model functioning as regard with similated basic processes: exchanges over bare ground, rainfall interception by plants, transpiration and soil moisture budget. It appears that considerable amount of observations is now available from the various field experiments realized during the past decade over a broad variety of characteristic biomes. However, a recommendation is made for collecting micro-meteorological and soil moisture measurements on the long term (at least one year) in order to provide the necessary information for validating the simulation of the annual budgets of water and heat.

The implementation of ISBA within a meso-β-scale model provided the necessary framework for the interpretation of the various data collected during large field experiments. Because the spatial variability of land cover at lengths of 10 km or greater was reproduced by ISBA, the mesoscale model was able to simulate reasonably the induced flow perturbations in the boundary layer, as observed. Moreover, mesoscale modeling allows to investigate how the spatial variability of surface properties and associated fluxes can be combined into spatial averaged values of surface fluxes against which GCMs might be calibrated.

Since ISBA was designed to be mainly used in NWP, it was mandatory to develop an assimilation scheme for soil moisture. A simple method is proposed, based on optimum analysis in which correction increments of the soil water content are made dependent on forecast errors of near-surface air temperature and relative humidity.

Appendix A. Important coefficients

A.1. Soil/vegetation heat capacity

The surface thermal characteristics is described by the soil/vegetation heat capacity

$$C_T = \left(\frac{1 - veg}{C_G} + \frac{veg}{C_V} \right)^{-1} \tag{17}$$

where $C_V = 2 \times 10^{-5}$ Km^{2}J^{-1} and

$$C_G = C_{G_{sat}} \left(\frac{w_{sat}}{w_2} \right)^{b/2 \log 10} \tag{18}$$

In the above equation, w_{sat}, $C_{G_{sat}}$ and b are parameters which are related to soil texture.

A.2. Volumetric moisture at the balance of gravity and capillary forces

The surface volumetric moisture at the balance of gravity and capillarity forces w_{geq} is derived from Clapp and Hornberger's specifications for soil hydraulic properties:

$$\frac{w_{geq}}{w_{sat}} = \frac{w_2}{w_{sat}} - a \left\{ \left(\frac{w_2}{w_{sat}} \right)^p \left[1 - \left(\frac{w_2}{w_{sat}} \right)^{8p} \right] \right\} \tag{19}$$

where a and p are empirical parameters.

A.3. Force-restore coefficients

The force and restore coefficients C_1 and C_2 are calibrated against a multi-layer soil moisture model and are respectively

$$C_1 = C_{1sat}\left(\frac{w_{sat}}{w_g}\right)^{b/2+1} \tag{20}$$

and

$$C_2 = C_{2ref}\left(\frac{w_2}{w_{sat} - w_2 + w_l}\right) \tag{21}$$

where C_{1sat} and C_{2ref} are parameters and w_l is a small numerical value.

A specific formulation of the C_1 coefficient was derived in order to describe implicitly the vapor phase transfers within very dry soils (i.e. $w_g < w_{wilt}$) (Braud et al., 1993; Giordani et al., 1996). The C_1 coefficient is approximated in this range of soil moisture values by the Gaussian distribution:

$$C_1 = C_{1max}\exp\left[-\frac{w - w_{max}}{2\sigma^2}\right] \tag{22}$$

In this expression,

$$C_{1max} = (1.19w_{wilt} - 5.09) \times 10^{-2}T_s + (1.464w_{wilt} + 17.86) \tag{23}$$

$$w_{max} = \eta w_{wilt} \tag{24}$$

with

$$\eta = (-1.815 \times 10^{-2}T_s + 6.41)w_{wilt} + (6.5 \times 10^{-3}T_s - 1.4) \tag{25}$$

and

$$\sigma^2 = -\frac{w_{max}^2}{2\ln\left(\frac{0.01}{C_{1max}}\right)} \tag{26}$$

A.4. Continuous formulation of the soil secondary parameters

The following relationships were derived from the Clapp and Hornberger (1978) and ISBA coefficients (Giordani, 1993; Noilhan and Lacarrère, 1995). Each hydraulic parameter is estimated when the fraction of sand or the fraction of clay is known. The sand and clay composition (i.e., $SAND$ and $CLAY$) are expressed in percentage.

The saturated volumetric water content (m^3m^{-3}):

$$w_{sat} = (-1.08 \times SAND + 494.305) \times 10^{-3} \tag{27}$$

The wilting point volumetric water content (m^3m^{-3}):

$$w_{wilt} = 37.1342 \times 10^{-3}(CLAY)^{0.5} \tag{28}$$

The field capacity volumetric water content (m^3m^{-3}):

$$w_{fc} = 89.0467 \times 10^{-3}(CLAY)^{0.3496} \tag{29}$$

The slope of the retention curve:

$$b = 0.137 \times CLAY + 3.501 \tag{30}$$

The soil thermal coefficient at saturation (Km^2J^{-1}):

$$C_{G_{sat}} = -1.557 \times 10^{-2} SAND - 1.441 \times 10^{-2} CLAY + 4.7021 \tag{31}$$

The value of C_1 at saturation:

$$C_{1sat} = (5.58 \times CLAY + 84.88) \times 10^{-2} \tag{32}$$

The value of C_2 for $w = 0.5 w_{sat}$:

$$C_{2ref} = 13.815 \times CLAY^{-0.954} \tag{33}$$

The coefficient C_3:

$$C_3 = 5.327 \times CLAY^{-1.043} \tag{34}$$

The coefficients for the w_{geq} formulation:

$$a = 732.42 \times 10^{-3} CLAY^{-0.539} \tag{35}$$

$$p = 0.134 \times CLAY + 3.4 \tag{36}$$

Appendix B. Surface resistance

The surface resistance used for transpiration is:

$$R_s = \frac{R_{smin}}{LAI} F_1 F_2^{-1} F_3^{-1} F_4^{-1} \tag{37}$$

where R_{smin} is unconstrained stomatal resistance, F_1 parameterises the effect of photosynthetically active radiation

$$F_1 = \frac{1+f}{f + R_{smin}/5000} \tag{38}$$

where $f = 1.1 R_G/(R_{GL} LAI)$. R_{GL} is a limit value of global radiation species-dependent.

F_2 describes availability of water in the root zone for transpiration

$$F_2 = \begin{cases} 1 & : \quad w_2 > w_{fc} \\ \frac{w_2 - w_{wilt}}{w_{fc} - w_{wilt}} & : \quad w_{wilt} \leq w_2 \leq w_{fc} \\ 0 & : \quad w_2 < w_{wilt} \end{cases} \tag{39}$$

F_3 represents the effect of the vapor pressure deficit in air

$$F_3 = 1 - \gamma[e^*(T_a) - e_a] \tag{40}$$

where γ is an empirical parameter, $e^*(T_a)$ is the saturated vapor pressure at temperature T_a and e_a is vapor pressure. F_4 allows for a temperature dependence of the stomatal resistance

$$F_4 = 1 - 0.0016(298 - T_a)^2 \tag{41}$$

Appendix C. Drag surface coefficients

The formulation of Louis (1979) giving drag coefficients C_H and C_D in the surface boundary layer was modified in order to consider different roughness length values for heat z_{0h} and momentum z_0 (Mascart et al., 1995):

$$C_D = C_{DN} F_m \tag{42}$$

$$C_H = C_{DN} F_h \tag{43}$$

with

$$C_{DN} = \frac{k^2}{[\ln(z/z_0)]^2} \tag{44}$$

where k is the Von Karman constant. Also

$$F_m = 1 - \frac{10Ri}{1 + C_m\sqrt{|Ri|}} \quad \text{if} \quad Ri \leq 0 \tag{45}$$

$$F_h = \frac{1}{1 + \frac{10Ri}{\sqrt{1+5Ri}}} \quad \text{if} \quad Ri > 0 \tag{46}$$

and

$$F_h = \left[1 - \frac{15Ri}{1 + C_h\sqrt{|Ri|}}\right] \times \left[\frac{\ln(z/z_0)}{\ln(z/z_{0h})}\right] \quad \text{if} \quad Ri \leq 0 \tag{47}$$

$$F_h = \frac{1}{1 + 15Ri\sqrt{1+5Ri}} \times \left[\frac{\ln(z/z_0)}{\ln(z/z_{0h})}\right] \quad \text{if} \quad Ri > 0 \tag{48}$$

where Ri is the Richardson's number. The coefficients C_m and C_h of the unstable case are given by:

$$C_m = 10C_m^* C_{DN}(z/z_0)^{p_m} \tag{49}$$

$$C_h = 15C_h^* C_{DN}(z/z_{0h})^{p_h} \times \left[\frac{\ln(z/z_0)}{\ln(z/z_{0h})}\right] \tag{50}$$

where C_m^*, C_h^*, p_m, and p_h are functions of the ratio $\mu = \ln(z_0/z_{0h})$ only:

$$C_h^* = 3.2165 + 4.3431 \times \mu + 0.5360 \times \mu^2 - 0.0781 \times \mu^3 \tag{51}$$

$$C_m^* = 6.8741 + 2.6933 \times \mu - 0.3601 \times \mu^2 + 0.0154 \times \mu^3 \tag{52}$$

$$p_h = 0.5802 - 0.1571 \times \mu + 0.0327 \times \mu^2 - 0.0026 \times \mu^3 \tag{53}$$

$$p_m = 0.5233 - 0.0815 \times \mu + 0.0135 \times \mu^2 - 0.0010 \times \mu^3 \tag{54}$$

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