Validation of land-surface parameterizations: Developments and experiments at the French Weather Service

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1 Introduction

This paper summarizes the various steps involved in the validation of a Land-Surface Parameterization (LSP) for Numerical Weather Prediction (NWP) and climate simulation studies. A simple Interaction Soil-Biosphere-Atmosphere (ISBA) scheme, amenable to the structure of a two-soil layer model, including a representation of the vegetation, has been designed for use in meteorological models (Noilhan and Planton 1989: NP89). The scheme was developed in reducing as much as possible the number of input parameters, while attempting to preserve the representation of the physics which control the surface and subsurface processes. A modelling strategy for evaluating the impact of the land surface scheme at various space and time scales has been developed and is briefly described in the following. A brief review of available data set for LSP validation is first given. The HAPEX-MOBILHY surface data set (André et al. 1988) has proven to be a fruitful source of information for testing ISBA at the local scale. However, other data sets have also been considered (FIFE87, ARME84, AVIGNON84 and EFEDA91). Some examples of local validations and calibrations of the scheme over a wide range of surface types and atmospheric conditions are then presented. Finally, developments related to the inclusion of the scheme in 3D-atmospheric models are discussed.

Brief description of the Interaction Soil Biosphere Atmosphere transfers scheme (ISBA)

ISBA is concerned with the following properties:

- The exchanges of heat and water within the first meter of soil depending both on the soil texture and the soil moisture.
- The water transfers from a vegetation layer: Interception/evaporation of precipitation or dew and transpiration, with some dependence on plant species (stomatal control).

The scheme has five prognostic variables:

- The volumetric near-surface water content associated with a thin upper layer (1 cm)
- The bulk soil moisture corresponding to the total soil layer (about 1 m)
- The surface temperature of the ground/vegetation medium
- The mean soil temperature
- The amount of water retained by the foliage.

A minimum of land surface information is required prior to using ISBA. Table 1 lists all the parameters needed, mostly dependent upon the soil texture and the type of vegetation. The soil parameters are related to the texture (Table 2) and have been calibrated from the USDA classification. For the parameters characterizing the plants, a sensitivity study (Jacquemin and Noilhan 1990) has revealed that two of them had a major influence on the prediction of energy partitionning:

- The fraction of vegetation *veg* within a model grid cell, since this quantity controls the partition of the latent heat flux between ground surface evaporation and plant transpiration.
- The minimum surface resistance R_{smin}, controlling the Bowen ratio.

In summary, the most important surface parameters required for ISBA are:

- The soil texture
- The fractional area of vegetation *veg*
- The minimum stomatal resistance R_{smin}
- The leaf area index LAI.

2 An overview of land surface observational studies

Shuttleworth (1991) provided a detailed description of past and proposed large-scale observational studies of land surface interactions. These data set can be used to calibrate land/atmosphere transfer schemes in meteorological model.

2.1 Single site data

The Amazonian Region Micrometeorological Experiment (ARME) was designed to document the energy-water budget of the tropical rainforest (Shuttleworth et al. 1984). At a single point situated in the Reserva Ducke near Manaus (central Amazonia) two automatic weather stations and an HYDRA correlation device were mounted on a 45 m tower above the forest. Measurements of

Primary Parameters	Secondary Parameters	Notation	Units
Dominant type of soil texture	Saturated volumetric moisture	w_{sat}	m^3m^{-3}
	Wilting point volumetric moisture	w_{wilt}	m^3m^{-3}
	Slope of the retention curve	b	-
	Saturated soil thermal coefficient	C_{Gsat}	Km^2J^{-1}
	Value of C_1 at saturation	C_{1sat}	-
	Value of C_2 for $w_2 = 0.5w_{sat}$	C_{2ref}	-
	Coefficients of w_{geq} formulation	a, p	-
Depth of soil	Depth of soil	d_2	m
Dominant type of vegetation	Fraction of vegetation	veg	-
	Minimum surface resistance	R_{smin}	sm^{-1}
	Leaf area index	LAI	m^2m^{-2}
	Roughness length	z_0	m
Albedo	Albedo	α	-
	Emissivity	ϵ	-

Table 1: A list of parameters for the ISBA surface scheme.

Soil texture	w_{sat}	b	c_{gsat}	p	a	C_{2ref}	C_{1sat}
Sand	0.395	4.05	3.222	4	0.387	3.9	0.082
Loamy sand	0.410	4.38	3.057	4	0.404	3.7	0.098
Sandy loam	0.435	4.90	3.560	4	0.219	1.8	0.132
Silt loam	0.485	5.30	4.418	6	0.105	0.8	0.153
Loam	0.451	5.39	4.111	6	0.148	0.8	0.191
Sandy clay loam	0.420	7.12	3.670	6	0.135	0.8	0.213
Silty clay loam	0.477	7.75	3.593	8	0.127	0.4	0.385
Clay loam	0.476	8.52	3.995	10	0.084	0.6	0.227
Sandy clay	0.426	10.4	3.058	8	0.139	0.3	0.421
Silty clay	0.482	10.4	3.729	10	0.075	0.3	0.375
Clay	0.482	11.4	3.600	12	0.083	0.3	0.342

Table 2: Values of the soil parameters according to the textural classification of Clapp and Hornberger (1978) and Noilhan and PLanton 's (1989) calibration: b is the slope of the retention curve, the saturation moisture content w_{sat} is in m^3/m^3, C_{gsat} is in $10^{-6}Km^2J^{-1}$ and all other coefficients are dimensionless.

radiation fluxes (thermal, solar, net), temperature, humidity, wind speed, turbulent heat fluxes and friction velocity were made for a 2-year period (09/83-08/85). Interception loss was determined from bucket raingauges measurements above and below the canopy. From these measurements average interception loss by Amazonian rain forest was estimated to 11.1% of gross rainfall (Lloyd et al. 1988).

The AVIGNON84 experiment was performed during a one-week period (19-26 June 84) in South France to document the progressive drying of a bare soil after an irrigation period (Passerat et al. 1989). A micrometeorological station recorded temperature, wind speed, water vapour and radiation fluxes. Turbulent fluxes were deduced from three classical methods (aerodynamic, Bowen ratio, energy balance). An original feature of this data set is the large amount of information about the soil. Detailed measurements were made within the soil for water content, hydraulic head and temperature profiles up to 80 cm. The soil heat flux has been deduced from the harmonic method (Fourier analysis of the soil surface temperature). Soil samples were taken at different locations to infer hydrodynamics and textural properties (bulk density, textural components, saturated hydraulic conductivity, thermal conductivity).

2.2 Large-scale observational studies

The HAPEX-MOBILHY (Hydrologic Atmospheric Pilot Experiment - Modélisation du Bilan Hydrique) experiment took place in Southwestern France during a two-year period (85-86) to estimate area-averaged evaporation fluxes over a 100 $\times$ 100 km^2 domain (André et al. 1986). A high-density network of raingauges, automatic weather stations, surface energy budget measuring stations, soil moisture measurement stations and streamgauges was established. During a Special Observing Period (May-July 86) the components of the water balance were accurately measured at 12 locations over crop areas (evapotranspiration by energy balance method, soil moisture by neutron-sounding technique, rainfall), plus one location above a pine forest with turbulent fluxes estimated by the eddy-correlation technique. Area-averaged turbulent fluxes were taken by an instrumented aircraft.

The FIFE (First ISLCSP Field Experiment) experiment (Sellers et al. 1990) was organized to study the biophysiological processes controlling the exchanges (radiation, heat, water, CO_2) between a vegetated area and the atmosphere, and to use satellite data to allow integration of local understanding to regional scales. The experiment was centred on a 15 $\times$ 15 km^2 area of grassland in central USA (Kansas). From early 87 to late 89, satellite, meteorological, hydrological and biometric measurements were routinely taken over the site. Four times in 87 and once in 89, during 10-20 day SOP surface fluxes of momentum, heat, moisture and CO_2, radiances and biophysical and soil variables were measured. Six aircraft were used to obtain remote-sensing and turbulent flux data.

The EFEDA (ECHIVAL Field Experiment in Desertification-Threatned Area) experiment (Bolle et al. 1992) was design to investigate land surface changes that might occur as a result of the "greenhouse effect" in semi-arid regions. Surface-atmosphere exchanges were documented at various time and space scales over a 100 × 100 km^2 in the region of Castilla-La-Mancha (Spain) during June 91. At three locations energy fluxes and temperature as well as humidity profiles were measured in the atmosphere. Soil and vegetation properties were determined together with soil moisture measurements. For example, accurate estimates of the soil heat flux G were provided at the BARRAX site. Measurements of G at five centimetres beneath the surface using Thornthwaite plates were corrected to take account of the heat stored within the first centimetres of soil. In addition, soil thermal properties (volumetric heat capacity and thermal conductivity) were measured in the vicinity of this flux station. Aircraft data were also collected in order to provide large scale integrated fluxes.

3 Sensitivity and validation studies at the local scale

Most of the 1D tests were realized with a boundary layer model linked with the surface parameterization, allowing a full interaction between the atmospheric and surfaces processes. The advantage of this method is to facilitate the inclusion of ISBA within a meteorological model since the various feedbacks were handled during the calibration sequences. Some tests were performed in a stand-alone mode, 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 was prescribed from available observations.

Series of 1D tests were run to investigate the ability of the surface scheme to reproduce the partitioning of available energy for contrasting meteorological conditions (clear sky and rainy days) and for various land surface covers (bare ground, grasslands, crops and forest). Table 3 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 AVIGNON84 and EFEDA91 data sets. Of particular interest is the magnitude of day-time soil heat flux G which may represents a significant fraction of net radiation ($\approx 30\%$), particularly in dry conditions. Fig.1a shows a time evolution (10 consecutive days) of observed and predicted soil heat flux G at BARRAX. 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. Note the day-time values of G which peak to 200 W/m^2. Previous run using the parameterization of thermal conductivity proposed

REFERENCE	DATA	TIMESCALE	TYPE OF SOIL/VEGETATION
Noilhan and Planton, MWR89	Hapex-Mobilhy	Clear days	CROPS/LOAM-SAND
Jacquemin and Noilhan, BLM90	id	id	PINE FOREST/ SAND
Mahfouf and Jacquemin, JAM89	id	Rainy days	CROPS/LOAM-SAND
Mahfouf, BLM90	id	months	CROPS /LOAM-SAND
Mahfouf and Noilhan, JAM91	AVIGNON84	week	BARE/LOAM
Braud et al., BLM92 (sub.)	EFEDA91	2 weeks	BARE/Silty Clay Loam
Germain, CNRM91	FIFE87	Clear days	GRASSLAND / LOAM
Manzi, CNRM90	ARME84	year	AMAZONIAN FOREST / LOAM
Mahfouf, JAM91	HM86	week	ASSIMILATION OF SOIL MOISTURE
Bouttier, JAM92 (sub.)	id	id	id

Table 3: SUMMARY OF 1D TESTS WITH ISBA/CNRM

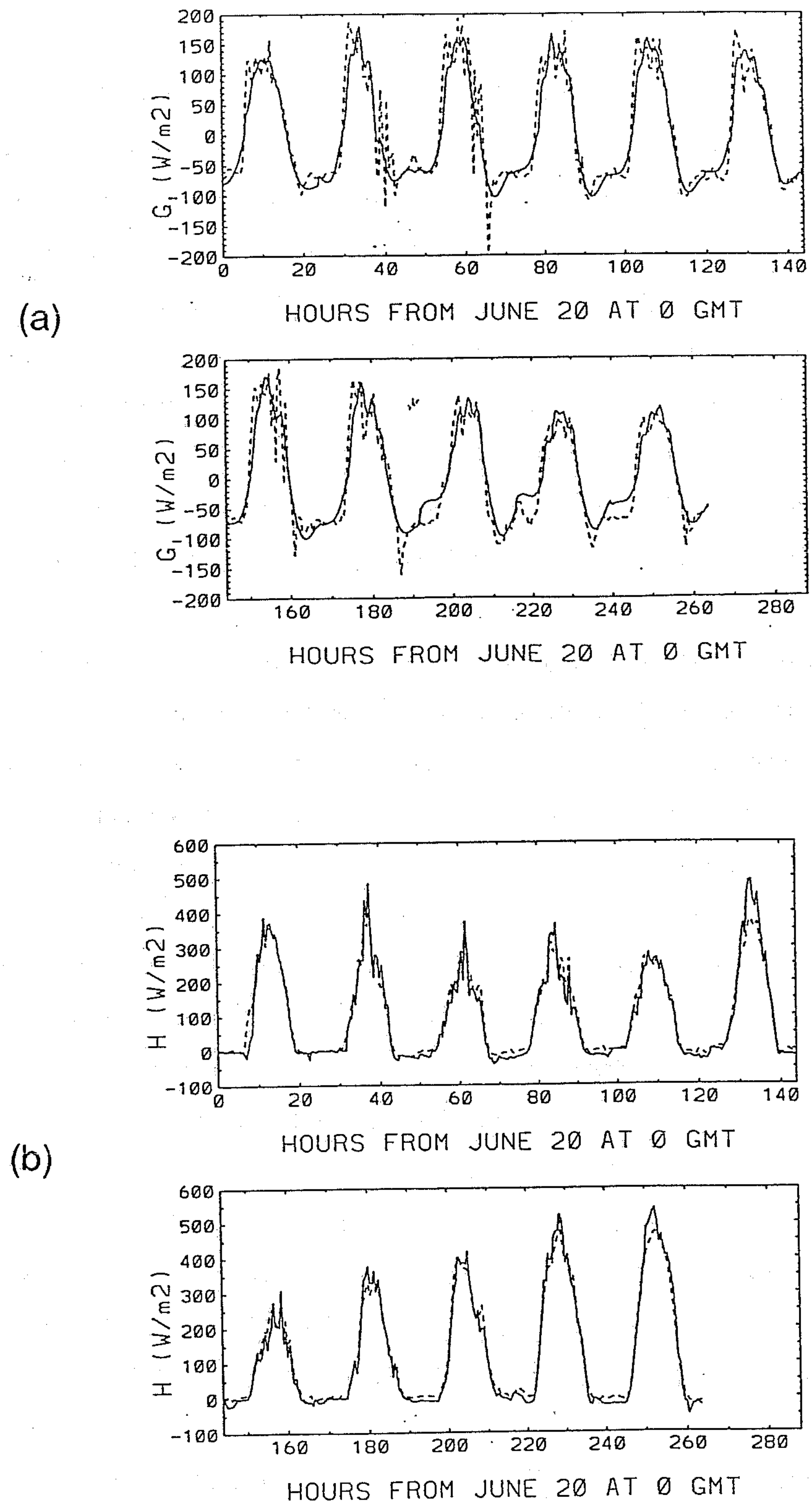


Figure 1: Comparison of the predicted soil heat flux G (a) and sensible heat H (b) with observations taken continuously during 11 days at the BARRAX site during EFEDA. (After Braud et al. 1992)

by Al Nakshabandi and Konhke (1965) provided overestimated values of G by about 20 %. A similar behaviour was observed by Mahfouf and Noilhan (1991) using the AVIGNON84 data set. From these two studies, relationships proposed by Van de Griend and O'Neil (1986) for thermal conductivity seem more appropriate than Al Nakshabandi and Konhke proposal, and should be used preferably when direct measurements are not available (particularly for wet soils).

Another important feature concerning heat exchanges over bare ground is the significant departure between the roughness length for momentum Z_{om} and heat Z_{oh} transfers. Although confirmed from observations since many years (Garratt and Hicks 1973, Brutsaert 1982; chap.4), few land-surface schemes distinguish between these two roughness lengths. Observations show that over natural surfaces the ratio Z_{om}/Z_{oh} is close to 10. Such an assumption was tested with the BARRAX data set (Braud et al. 1992) leading to a satisfactory prediction of the sensible heat flux H as shown in Fig.1b. Using the classical assumption that $Z_{oh} = Z_{om}$, the sensible heat flux was clearly overestimated.

Formulation of surface evaporation in simple parameterization 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 soil 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 (Sasamori 1970, Passerat et al. 1989, Mahfouf 1986, Noilhan 1987). In the parameterization context, the challenge is to reproduce the surface hydrology with a limited number both 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 scheme are based on bulk aerodynamical expressions by using a parametrization of the surface specific humidity (Deardorff 1978, Kondo et al. 1990). Some parametrizations 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 the AVIGNON84 data set, Mahfouf and Noilhan 1991 examined various formulations of surface evaporation (Fig. 2). A series of tests in a stand-alone mode were run, starting with observed soil water contents and changing only the formulation for surface evaporation. In the context of these observations, the diffusion-limiting moisture flux was strongly underestimated. On the other hand the most widely use bulk aerodynamic formulations (so-called α- and β-methods) provided comparable results during the day-time. On the other hand, the β-methods overestimated the nocturnal evaporation as depicted in Fig. 3 showing the better agreement with observed nocturnal dew flux using the α-method. Another difficulty with

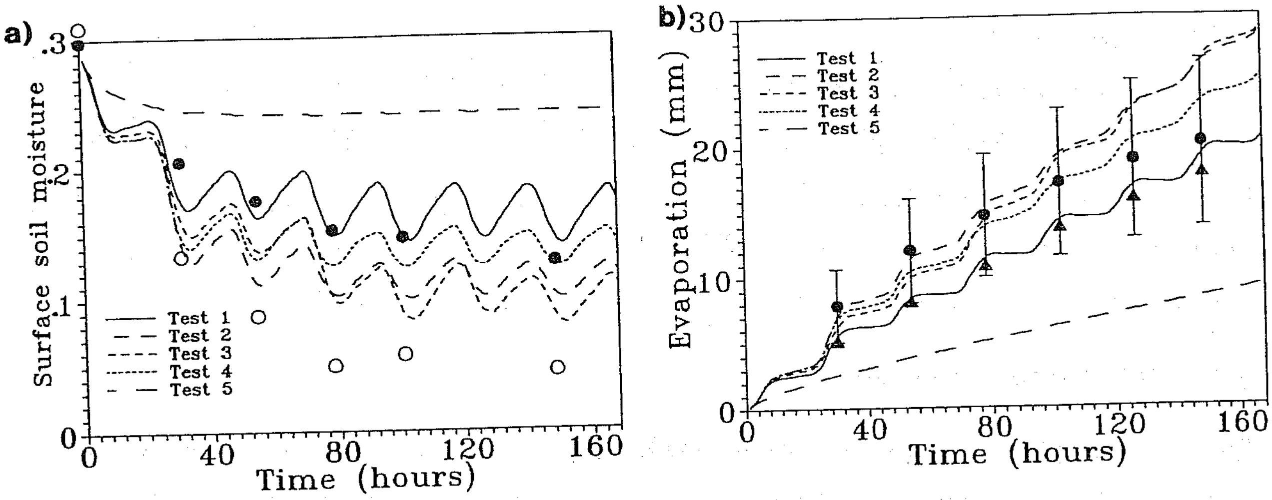


Figure 2: Evolution of the near-surface water content (a) and cumulative evaporation (b) for a plot of bare ground (after Mahfouf and Noilhan 1991). Daily observations of soil moisture are taken at 1.5 cm (full circles) and 0.5 cm (open circles). Estimations of evaporation are deduced from observed water budget (full circles with standard deviations) and aerodynamic measurements (full triangles) (AVIGNON84 data set). The different curves in (a) and (b) refer to ISBA predictions with various formulation of surface evaporation. Test 1 corresponds to the α method, while Test 4 refers to a β-scheme for surface evaporation.

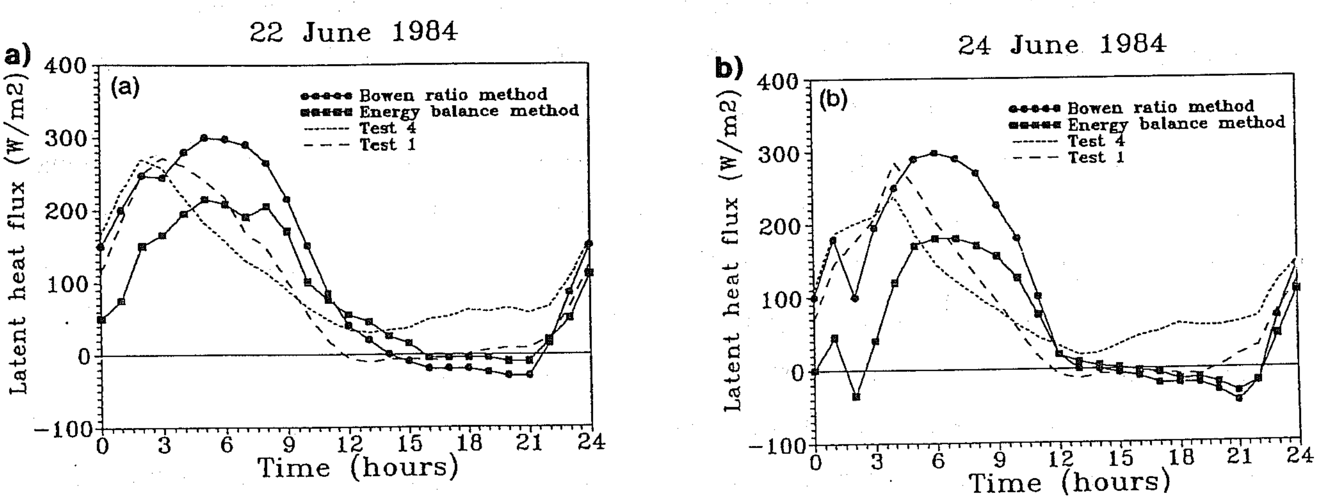


Figure 3: Comparison between observed and predicted latent heat fluxes for two days (AVIGNON84 data set). The nocturnal evaporation is better predicted using α-formulation (Test 1) than a β-method (Test 4). (after Mahfouf and Noilhan 1991)

β-formulations is that they necessitate prior calibration for a bare soil resistance with both soil moisture and soil texture.

In the case of very dry soils, effects of vapour 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. (1992) who improved ISBA by taking into account soil moisture transfers in vapour phase. This approach was tested favourably against observations of surface evaporation and near-surface water content at the BARRAX site.

3.2 Rainfall interception

Rainfall interception by vegetative canopies modifies the surface water balance because water evaporates back into the atmosphere at the potential rate. Interception loss can be significantly higher than transpiration rate, especially over forest canopies because their high value of roughness length. Forests have also a significant water storage capacity (several millimetres of water can be retained on the foliage leaf surfaces) as compare with short vegetative canopies. Interception also affects the vertical structure of the lower atmosphere. Stewart (1977) proved experimentally that forests, when their surface is wet, can draw energy from the overlying air to sustain their evaporation of intercepted rainfall involving negative values of the Bowen ratio. Thus, interception drives the development of a stably stratified boundary layer providing downward transport of heat and upward evaporation (Blyth et al. 1992). The link between interception processes and mesoscale meteorology was investigated by André et al. (1989). This study showed how the availability of additional water through interception can increase the regional rainfall through a positive feedback of atmospheric humidity. Finally, another interesting feature of rainfall interception is the rapid transition between wet and dry canopies (less than 10 hours). Thus, a correct estimation in a NWP model of the Bowen ratio and of the feedbacks between the boundary layer and the wetted surface will be only possible if the interception phenomenon is taken into account.

Rutter et al. (1971) derived the first predictive model of rainfall interception by computing the amount of water W_r retained on a one-layer vegetation canopy taking into account interception of rain, drainage rate and potential evaporation from the wetted part δ of the canopy. Most of recent interception schemes in LSP are conceptually similar to that of Rutter et al. (1971) approach. This is the case for ISBA, but the drainage rate from the canopy is simplified. Rainfall or dew are intercepted until the water storage reaches a critical saturation level W_{rmax}, at which stage excess water is assumed to drain immediately. The maximum reservoir capacity is set equal to $\alpha \times LAI \times veg$ where α is an empirical coefficient, often set to 0.2 mm (Dickinson 1984) which seems valid for coniferous canopies. In many calibration of interception schemes the α coefficient was tuned to reproduce observations.

A sensitivity study of ISBA by Mahfouf and Jacquemin (1989) confirmed the importance of high values of the 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 ARME84 have been widely used for calibrating interception schemes. As an example, Fig. 4 proposes different predictions of cumulative interception loss compared with observations. Measurements show that total interception loss was 428 mm while gross rainfall recorded in the 625 days of data was 4804 mm as reported by Lloyd et al. (1988). The measurements data set errors are mainly due to uncertainties in the cumulative measured throughfall (Lloyd et al. 1988) leading in some case to negative interception loss as shown in Fig. 4-2 between day 425-453. However, it is important to mention that the ARME data set is probably one of the most relevant for such calibration of LSP since long enough time series were recorded.

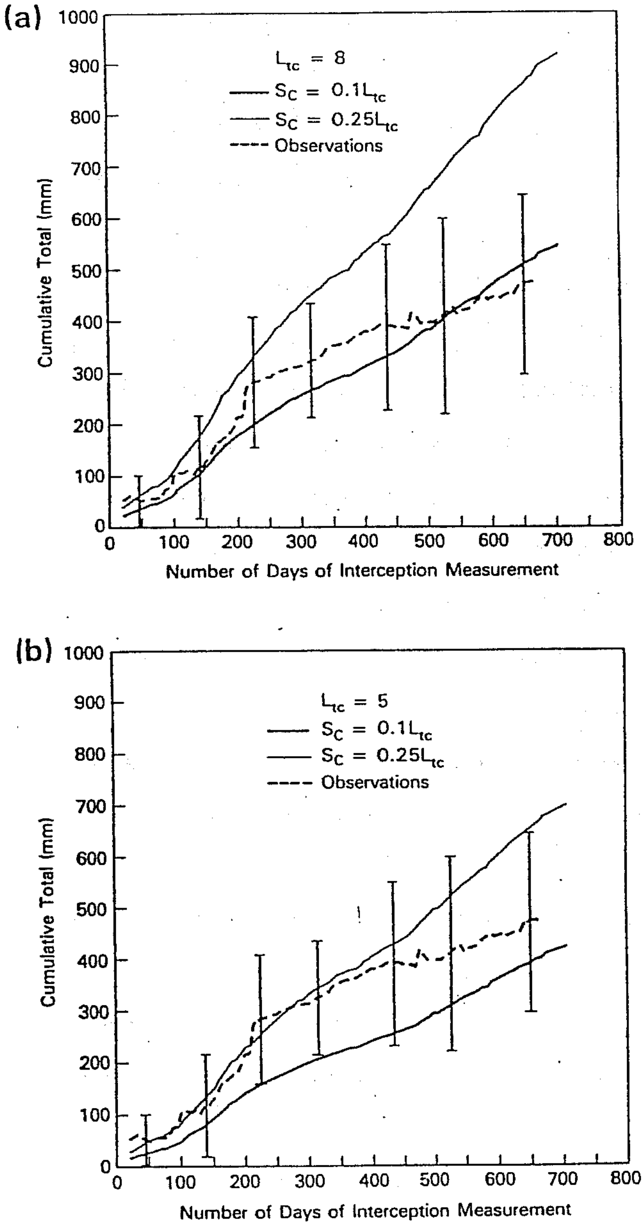
Four models (Rutter, Gash, SiB and ISBA models) were applied to the same automatic weather station forcing during two years. The Rutter model and Gash analytical model were used with the observed state parameters, in particular for the interception capacity estimated to 0.74 mm. The four simulations described in Sellers et al. (1989) are combinations of two leaf area indices and two values of W_{rmax}. Two simulations were run with ISBA. First of all, with $W_{rmax} = 0.1 \times LAI$ to be compare with one run of SiB. The second one with $W_{rmax} = 0.15 \times LAI$ =0.78 mm, for comparison with Rutter model results. From Fig. 4, all the schemes overestimate consistently the interception loss during the second year of measurements. It is difficult to conclude if these discrepancies are due to model imperfections or to measurement errors as advocated previously.

The SiB calculations show that $\alpha = 0.1$ yields a calculated interception loss in better agreement with observations than with $\alpha = 0.25$ (Fig. 4.1b). The integrations of Sellers et al. (1989) illustrate also the large sensitivity of the predictions to the value of LAI.

In the Rutter model, LAI is implicitly included since the observed interception capacity is used. Moreover, the aerodynamic resistance to the transport of water vapour is calculated from a simplified expression ($R_a = f/U_a$, where U_a is the wind speed at the reference level and f a constant).

The ISBA integration with $\alpha = 0.15$ provides comparable results with the Rutter model. On the other hand, ISBA predictions with $\alpha = 0.1$ are larger than the SiB results with comparable conditions. This difference may be due to different formulation of R_a in the two models. ISBA computes R_a from a single expression (Louis 1979) applied between the top of the canopy and the reference level (assuming that Z_0=3 m and a displacement height d=30 m), whereas in the SiB

Figure 4.1



Plot of cumulative interception loss over the 25-month period. The vertical error bars represent the range of interception loss estimates consistent within one standard error of the observations (see text). The dotted lines represent SiB calculated values for two different values of the total canopy leaf area index, L_{tc}, and two methods of calculating the canopy's maximum interception water storage capacity, S_c. (a) L_{tc} = 8.0, S_c = 0.25L_{tc} mm and S_c = 0.1L_{tc}; (b) L_{tc} = 5.01, S_c = 0.25L_{tc} mm and S_c = 0.1L_{tc} mm.

Figure 4.2

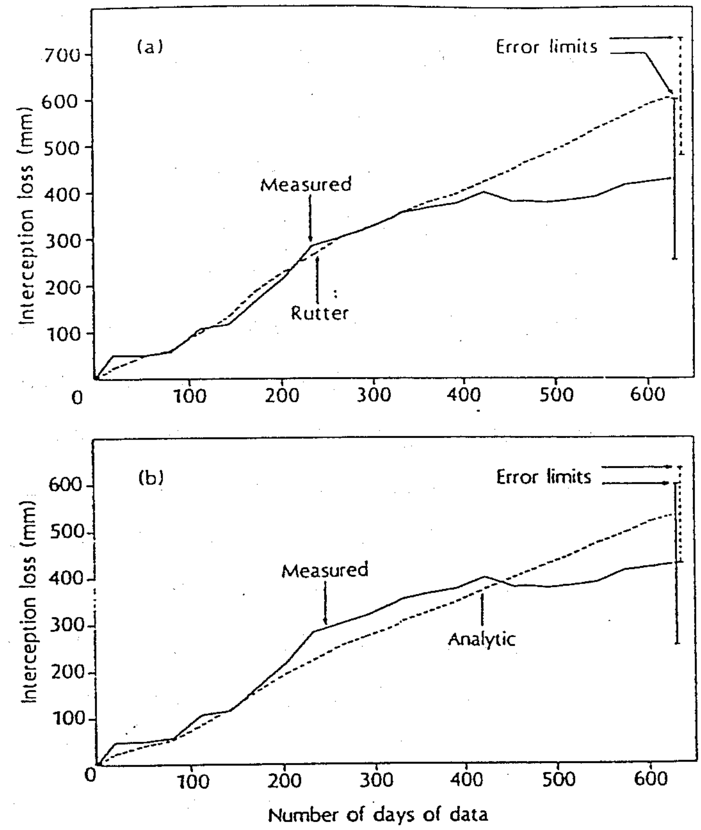


Figure 4.3

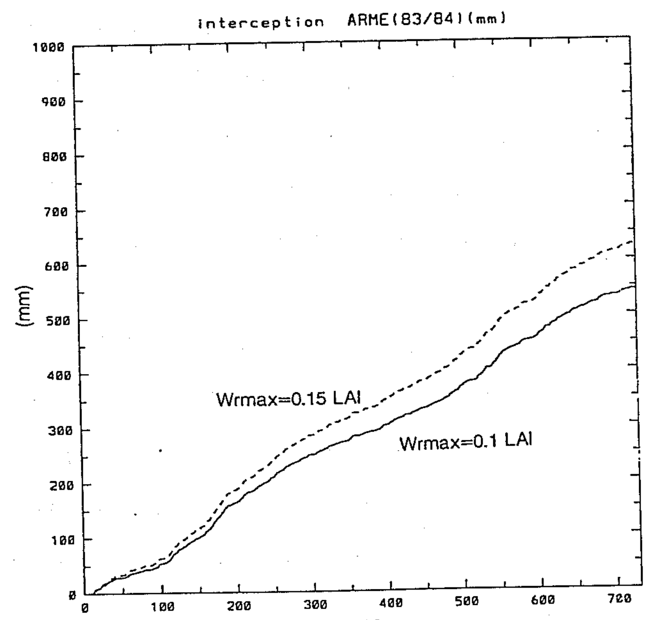


Figure 4: Prediction of interception loss of the Amazonian rain forest over a 25-month period (ARME84) using four interception models: SiB (Sellers et al. 1989) (4-1 ab); Rutter model and Gash's analytical model (4-2 ab) (Integrations described in Lloyd et al. 1988) and ISBA (4-3). The observations are reported in 4-1 and 4-2. The values of LAI and of the reservoir capacity W_{rmax} considered in the SiB integrations are given in 4-1a and 4-1b. ISBA was run with $LAI = 5.0$, while in Rutter and Gash models, W_{rmax} was set equal to 0.78 mm.

model the aerodynamic resistance between canopy air space and reference height, and the canopy bulk resistance are obtained from momentum transfers within and above the forest canopy.

For short canopies, direct measurement of interception is nearly impossible. However, evidence of reevaporation 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 prediction of interception processes by crop fields.

3.3 Plant transpiration and soil water budget

Plant transpiration is probably the main issue of LSP 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. Plants regulate the amount of water exchange for photosynthesis by means of the saturated tissues inside the leaf stomates. Most of LSP express plant transpiration as:

$$E_{tr} = veg\rho_a \frac{(q_{sat}(T_s) - q_a)}{R_a + R_s}$$

where $q_{sat}(T_s)$ is the saturation specific humidity at T_s, q_a the air specific humidity and R_s the bulk surface resistance.

In this expression, water saturation is assumed to be maintained within the plant stomates through biological processes. Plants adaptation to their environment is tentatively described in parameterizing the surface resistances R_s. The general approach of Jarvis (1976) is adopted by many LSP:

$$R_s = g(plant) f_1(R_g) f_2(soil\ moisture) f_3(\delta e) f_4(T_a)$$

The previous expression was initially developed for the resistance of water diffusion through a single leaf. The resistance for the whole canopy should be obtained by relating the profile of surface resistance through the canopy to the attenuation of solar radiation and atmospheric conditions within the canopy (Dolman et al. 1991). This seems particular important in the case of forest because of prevailing significant vertical gradients of temperature, humidity and wind speed. The study of Sellers et al. (1989) illustrates how this can be achieved with SiB for a given vegetation architecture. The stress factor f_2 in the expression of R_s are not readily available and must be determined either from complex physiological experiments and calibration of LSP versus observations.

Because ISBA does not takes into account processes inside the canopy, the parameterization of R_s attempts to describe the whole canopy. The function $g(plant)$ is set to R_{smin}/LAI where R_{smin} is a minimum surface resistance 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 (NP89, Jacquemin and Noilhan 1990, Mahfouf 1990a), reaching higher values $\approx$ 500 s/m when plants/crops experience maturation (NP89) 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 (1992). In these two cases of forested canopies, it is interesting to observe that the ratio R_{smin}/LAI is nearly constant.

The function $f_1(R_g)$ takes into account the effect of photosynthetically active radiation (PAR) assumed equal to 55% of the global radiation R_g. Fig. 5a shows the variation of $LAI \times f_1(R_g)/R_{smin}$ versus PAR in the cases of the Amazon forest for two values of LAI. The corresponding variation obtained with SiB (Sellers et al. 1989) is shown in Fig. 5b. It is rewarding to note that the methods used in the two models (SiB and ISBA) provides comparable results, despite the extremely simplified calculation in ISBA.

The function $f_3(\delta e)$ accounts for stomates closure in conditions of dry environment, a process particularly important for forest canopies:

$$f_3(\delta e) = [1. - h_e(e_{sat}(T_a) - e_a)]^{-1}$$

where $\delta e = e_{sat}(T_a) - e_a$ is the atmospheric vapour pressure deficit and h_e a parameter to be calibrated. The R_s dependence on $f_3(\delta e)$ is the most important environmental constraint for the transpiration of the Amazon forest and SiB calibration for the Amazon tropical forest provided h_e=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 h_e=0.037 hPa^{-1} for the Landes forest, in agreement with observations (Gash et al. 1989, Stewart 1988). Such calibration of $f_3(\delta e)$ over Les Landes forest (Jacquemin and Noilhan 1990) leads to a fair prediction of the canopy surface conductance when compared with the observed conductance estimated for the entire data set (Fig. 6). The maximum value of the canopy conductance is observed a few hours after sunrise, followed by a decrease all day long as a result of increasing atmospheric specific humidity deficit. The vapour deficit term $f_3(\delta e)$ seems less important for short canopies. However, it was found particularly significant for the prairie grass canopy in FIFE87 (Sellers et al. 1990). Thanks to this effect, Germain (1990) calibrated ISBA for 6 selected FIFE87 'golden' days and for 7 flux stations. Fig. 7 shows a comparison of hourly values of observed and predicted turbulent fluxes for the selected situations by Germain (1990).

The function $f_4(T_a)$ used in ISBA follows the Dickinson's 1984 proposal. The parabolic shape of $f_4(T_a)$ is centred around an optimum air temperature of 25 °C where the stress is negligible.

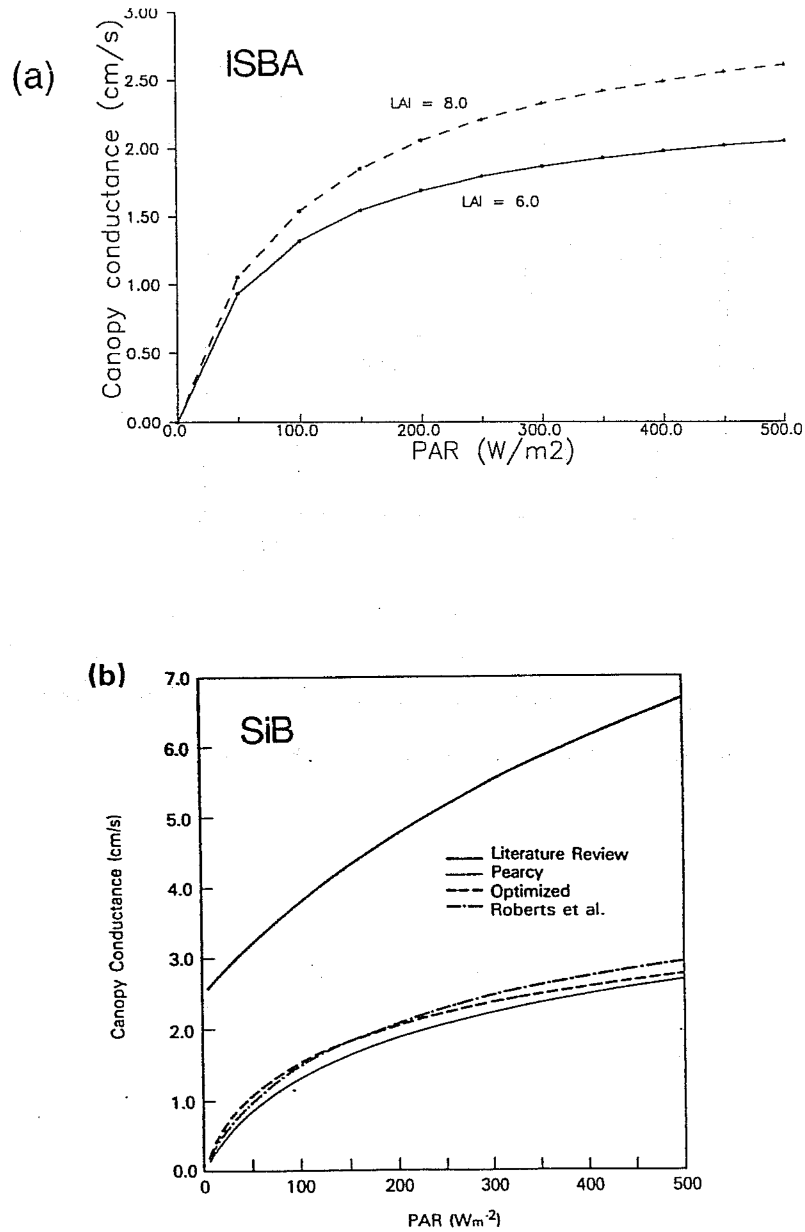


Figure 5: Variation of the light dependent (unstressed) canopy conductance versus PAR simulated by SiB **a)** (see Sellers et al. 1989 for calibration description) and ISBA **b)**

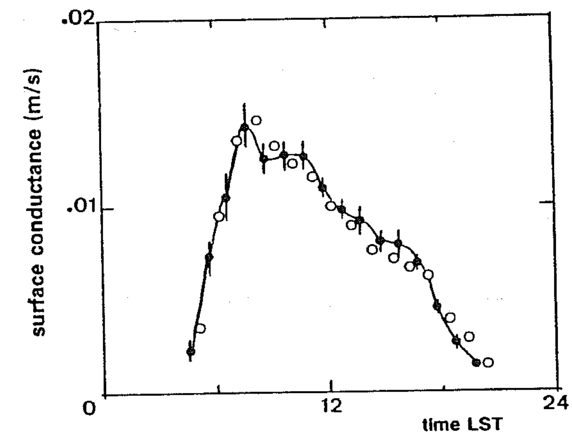


Figure 6: Daily variations of the mean observed (full circle) and predicted (open circle) surface conductance of the Landes forest (after Jacquemin and Noilhan 1990)

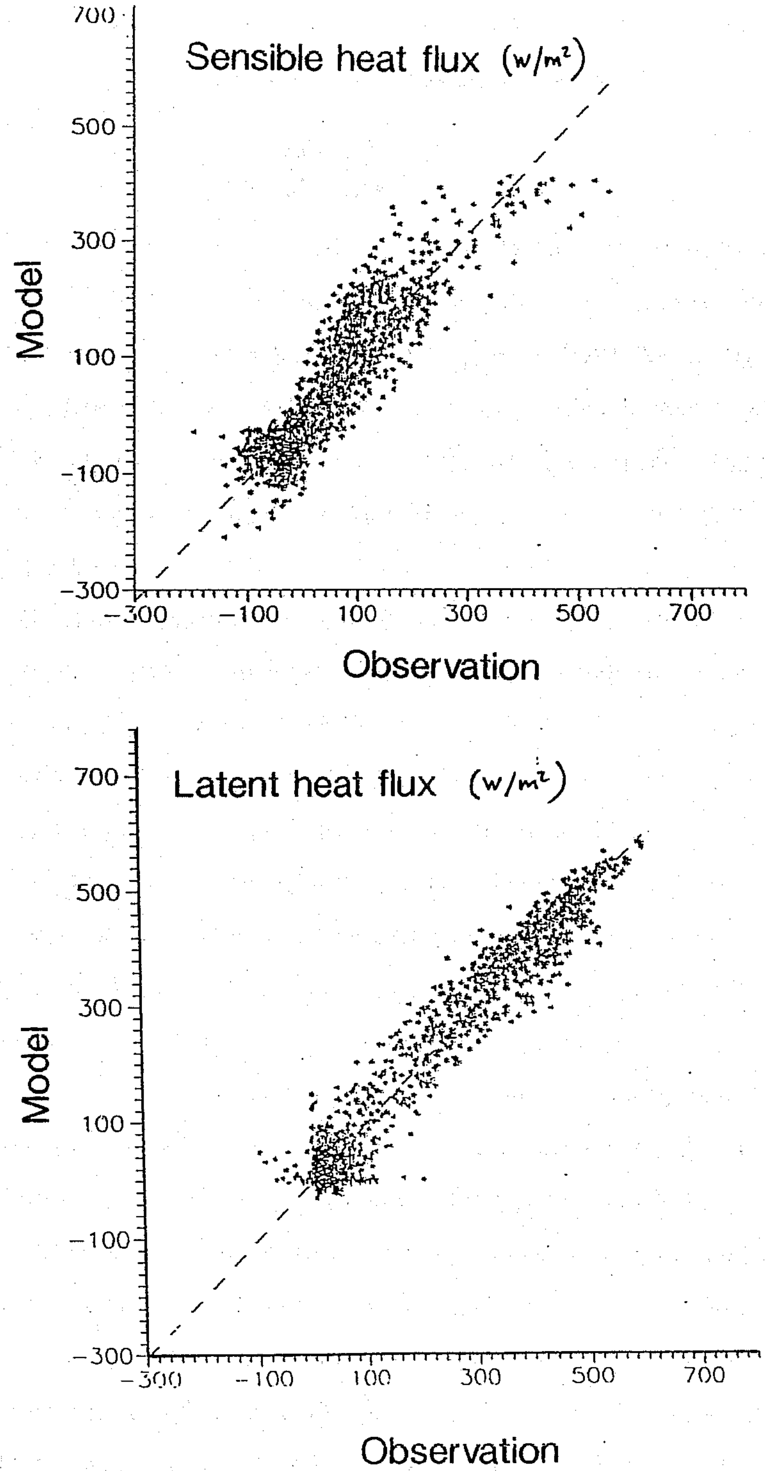


Figure 7: Scatter plot of predicted and observed sensible (a) and latent (b) heat fluxes over a prairie grass (FIFE87) (after Germain 1990)

Transpiration stops when T_a=0 °C or T_a=50°C, an assumption which sound valid for temperate latitudes. At the moment, no clear validation of $f_4(T_a)$ is obtained with ISBA.

The function f_2(*soil moisture*) expresses the plant stress versus soil moisture in the root zone. In many LSP, this stress factor is computed from the leaf water potential which depends on plant physiology and soil moisture. In ISBA, f_2 is computed as:

$$f_2 = \left(\frac{w_2 - w_{wilt}}{w_{fc} - w_{wilt}} \right)^{-1}$$

where the volumetric water contents w_2, w_{wilt} and w_{fc} refer respectively to soil moisture in the root zone, to the permanent wilting point and to field capacity. The values of w_{wilt} and w_{fc} are fixed for a given soil texture, w_{wilt} being computed 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). Choudhury (1983) was one of the first to simulate the effects of soil water potential on a corn canopy. The main difficulty to calibrate the soil water stress function is to find suitable data set, i.e. long enough time series of micrometeorological quantities together with soil moisture and soil hydraulics properties. These observations were collected during the HAPEX-MOBILHY Special Observing Period of 3 months (Goutorbe 1991, Goutorbe et al. 1989, Goutorbe and Tarrieu 1991, Cuenca and Noilhan 1991). Unfortunately, only a moderate soil water stress was observed. However, this data set was used by Mahfouf (1990a) to evaluate the prediction of soil moisture by ISBA for period of one or two months and for crop canopies. Fig. 8 shows examples of prediction of cumulative evaporation and soil moisture for three agricultural sites. In all cases, the predictions agree reasonably well with observations, although some scatter is observed between observations and predictions of soil moisture. It is important to mention that this data set seems fully consistent, e.g. the variation of the amount of soil moisture is reasonably well matched by exchanges at the soil-atmosphere interface (precipitation and evaporation) (Goutorbe et al. 1989). In the course of ISBA development, the result of Mahfouf (1990a) was particularly important since it gave substantial confidence to the ISBA functioning when used over long time periods.

Surface soil hydrology prediction was also examined with the ARME84 data set. Prediction of the near-surface and deep water contents for two years by ISBA is shown on Fig. 9. The soil and vegetation parameters used for this integration are given in Fig. 9 where details of modifications of the original interception scheme are given in Manzi and Planton (1992). Starting from the field capacity, the prediction of the soil moisture content (Fig. 9d) shows regular annual cycle with a wet season where soil moisture remains close to the saturation (w_{sat}=0.42 m^2/m^2) followed by a dry season with minimum values near about 0.25 m^2/m^2 (field capacity). Therefore, the soil water is always in ample supply and the forest does not experience significant water stress. Observations of near-surface and deep soil moisture using neutron probes reported by Cabral (personal

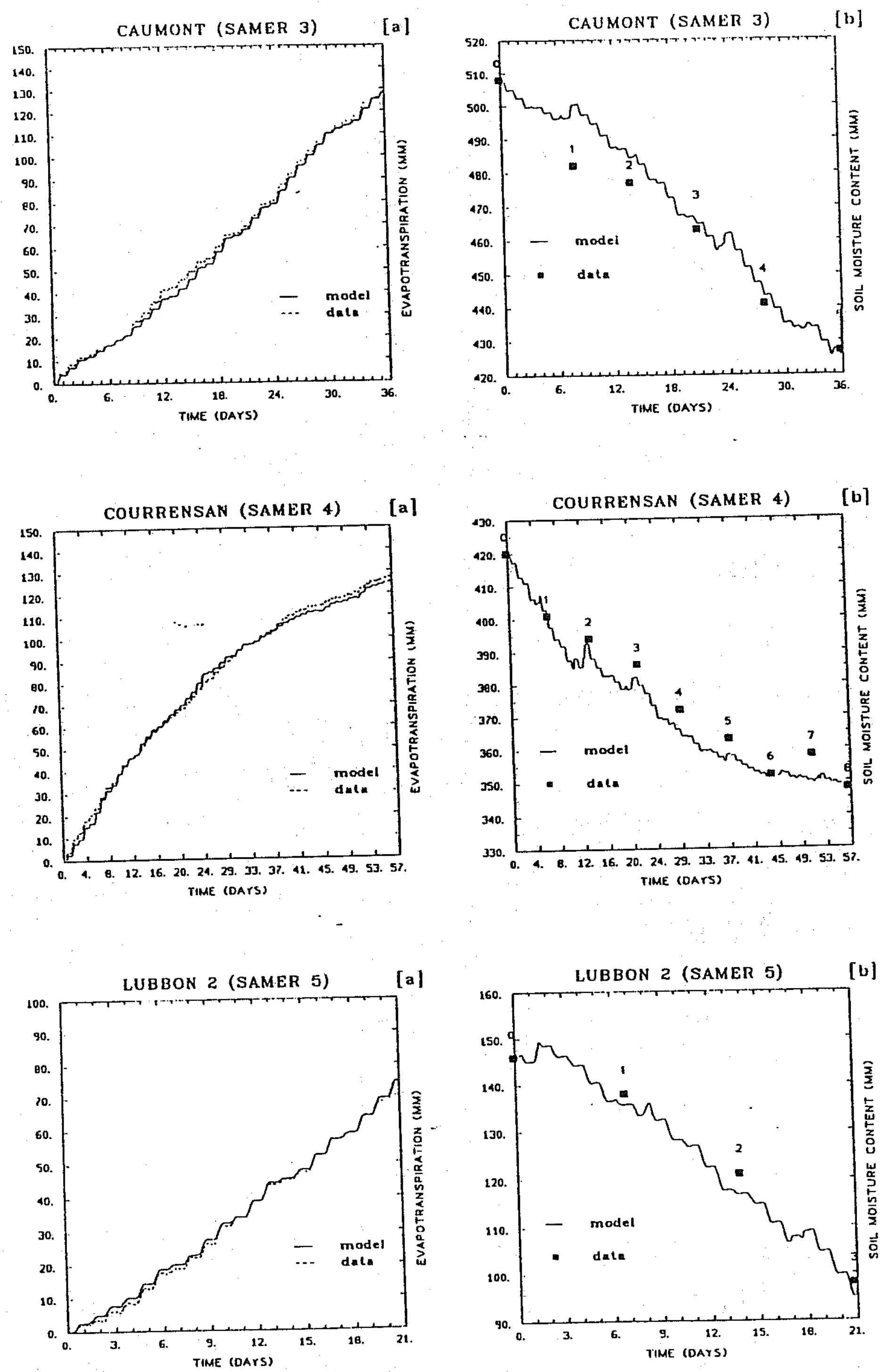


Figure 8:Time evolution of accumulated evaporation (left panels) and soil moisture content (right panels) observed (dashed line) and simulated (solid line) at three sites of HAPEX-MOBILHY 1986. The neutron sounding measurements of soil moisture are represented by full squares (Mahfouf 1990a).

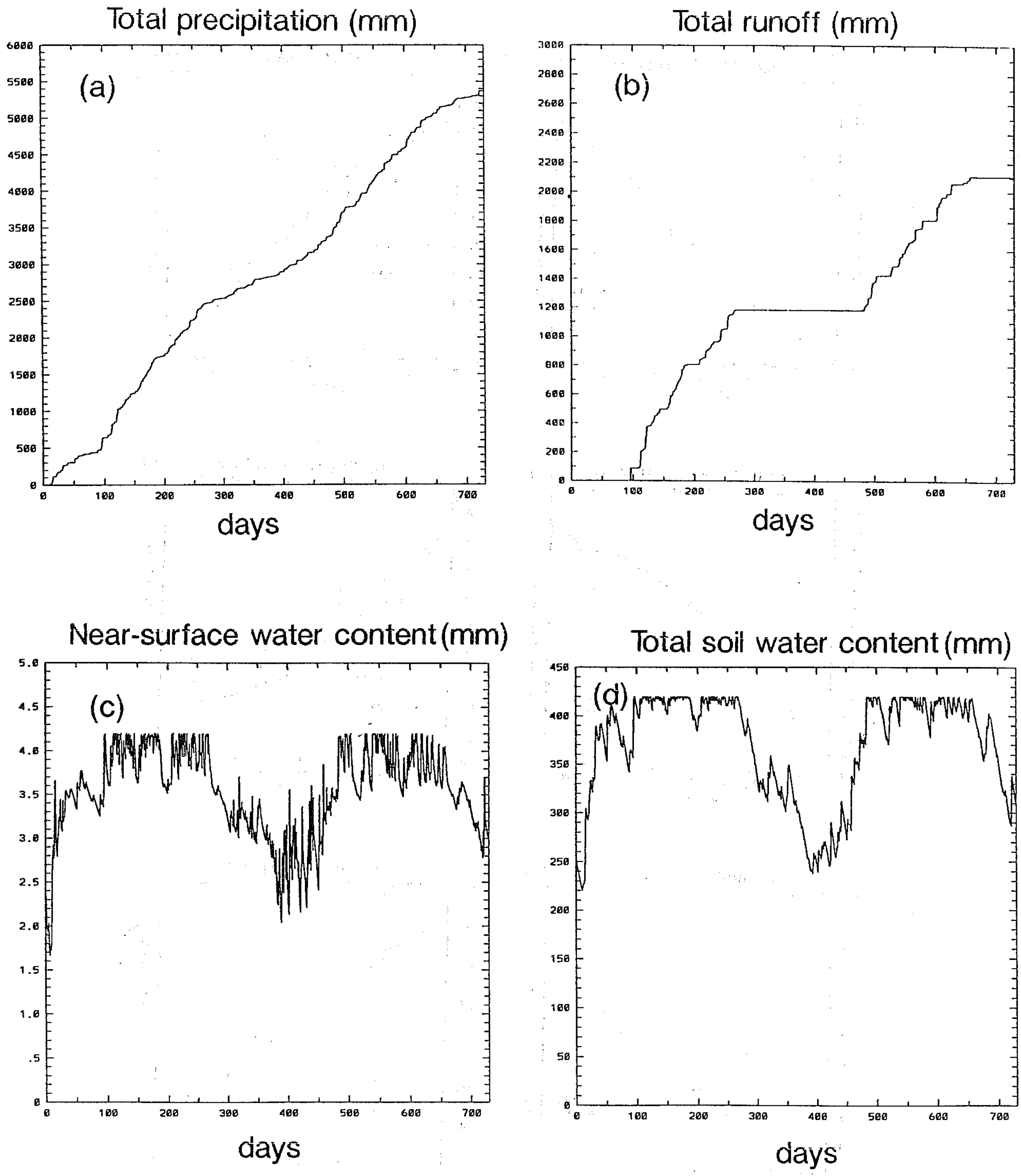


Figure 9: Prediction of soil water content and total runoff by ISBA for the Amazonian rain forest: total precipitation (a), near-surface water content (c), deep water content (d) and total runoff (b). Surface parameters are set to: R_{smin}=250 sm^{-1}, LAI=6, veg=0.99, d=30 m, soil type: silty clay loam, $w_{ginit} = w_{2init} = w_{fc}$, d_2=1 m.

communication) indicate that the prediction of soil water seems reasonable.

The ARME data set can also be used to investigate the runoff prediction. Indeed, $\approx$ 40% of total precipitation ($\approx$ 6000 mm in two years) is lost by the soil layer (Fig. 9b). Of course, runoff prediction is sensitive to the total depth of the soil layer d_2 assumed equal to 1 m in this run. On the other hand, given the large amount of rainfall, runoff prediction does not depend very much on initial values of soil moisture and on the soil texture.

As a result of these experiments, some parameters of the scheme were adjusted. Although slight changes were included relative to the initial version, we think now that there is no basis to complicate ISBA any more.

4 Integrations at meso-β and global scales (table 4)

4.1 Meso-β-scale integrations

A mesoscale modelling strategy has been developed (André and Bougeault 1988, André et al. 1989, Bougeault et al. 1991a, 1991b, Noilhan et al. 1991a, 1992b) to provide an integration tool as well as a useful framework for the interpretation of the experimental results collected during HAPEX-MOBILHY 1986. Briefly summarized the modelling work was as follows:

- Use the results of the 1D calibration tests to assemble a general table of possible values of the surface parameters and to produce maps of these parameters within the 3D simulation domain (400 km $\times$ 400 km) from detailed classification of both soil and vegetation types (Phulpin et al. 1989, Phulpin and Noilhan, 1989, Noilhan et al. 1991b).

- Integrate the model and compare the results with all the available data (surface network, radiosonde and aircraft fluxes measurements)

One-day integrations for four 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. Details of mesoscale integrations can also be found in Attié (1990) and Giordani (1991).

Concerning the impact of surface inhomogeneities on the atmosphere, the main striking pattern at the beginning of the observations was the development of a mesoscale circulation between the pine forest and the surrounding crop area. Because of high values of R_{smin} (Gash et al. 1989) and of the vegetation fraction veg over the forest, the sensible heat flux was higher than over the crop area both in the observations (André et al., 1988) and in the simulations on 16 and 19 June 1986 (Noilhan et al., 1991 ; Attié, 1990). On the other hand, at the end of the Special Observing Period (after July,1), the contrast between the crops and the forest was reduced as a result of crop development, resulting in a decrease in the soil heat storage (Giordani 1991).

REFERENCE	DATA	TIMESCALE	SPATIAL SCALE	GRID BOX SIZE
Bougeault et al., MWR91a	Hapex-Mobilhy	Clear day 16 june 86	MESOSCALE PERIDOT $400km \times 400km$	$10km$
Bougeault et al., MWR91b	id	id	id	id
Noilhan et al., MWR91	id	id	id	id
Noilh./Lacar., ECMWF91	id	id	id	id
Attié CNRM90	id	19 june 86	id	id
Noilhan et al., ESA91	id	1-8-10 july 86	id	id
Giordani CNRM91	id	id	id	id
Mahfouf, CNRM88	id	Rainy day	5 june 86	id
Blyth et al., JAM92 (sub.)	id	Rainy day	5 june 86	id
Noilhan-Lacarrère, CNRM92	EFEDA91	22-28 june 91	MESOSCALE $700km \times 700km$	$10km$
Manzi-Planton, JH92	-	1 year	GLOBAL EMERAUDE	2.8 degrees

Table 4: TESTING ISBA WITHIN 3D METEOROLOGICAL MODELS

4.2 Parameter aggregation over large areas

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 (1991) discussed methods for estimating *effective* vegetation properties and soil characteristics of a large area from the observations and the 3D numerical work realized 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. 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 than 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 behaviour 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 radiation fluxes, predictions of soil water content and total evaporation for 25 days compare well with observations available within the large area (Fig. 10).

A similar approach was applied to the subgrid distribution of precipitation and interception from mesoscale results obtained during a rainy day (Blyth et al. 1992). 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.

4.3 Implementing ISBA within a GCM

Finally, the ISBA scheme has been implemented in the AGCM (Atmospheric General Circulation Model) EMERAUDE. This coupled ISBA/AGCM is a powerful tool to investigate natural or anthropogenic changes in continental surfaces, like desertification or deforestation process 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 the two previous sections, these maps have been obtained from basic

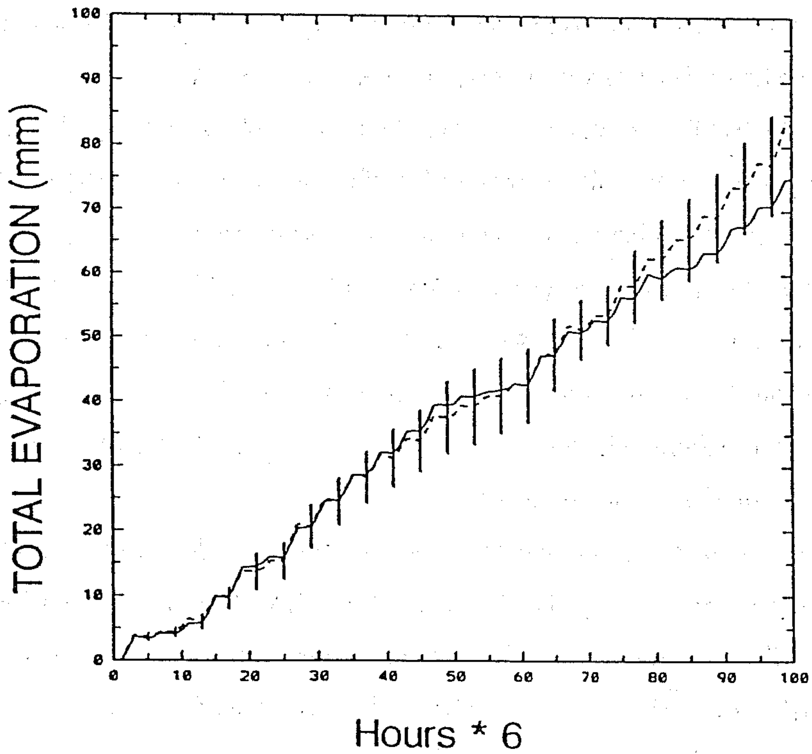


Figure 10: Area-average evaporation flux for the HAPEX-MOBILHY square (130 km × 70 km predicted by an atmospheric column model coupled with ISBA. The surface scheme uses effective values of surface properties (vegetation and soil) which takes into account the spatial variability of land-use categories within the experimental area. The aggregated flux is compared with an average of observations available during the 25 days of the integration (Noilhan and Lacarrère 1991).

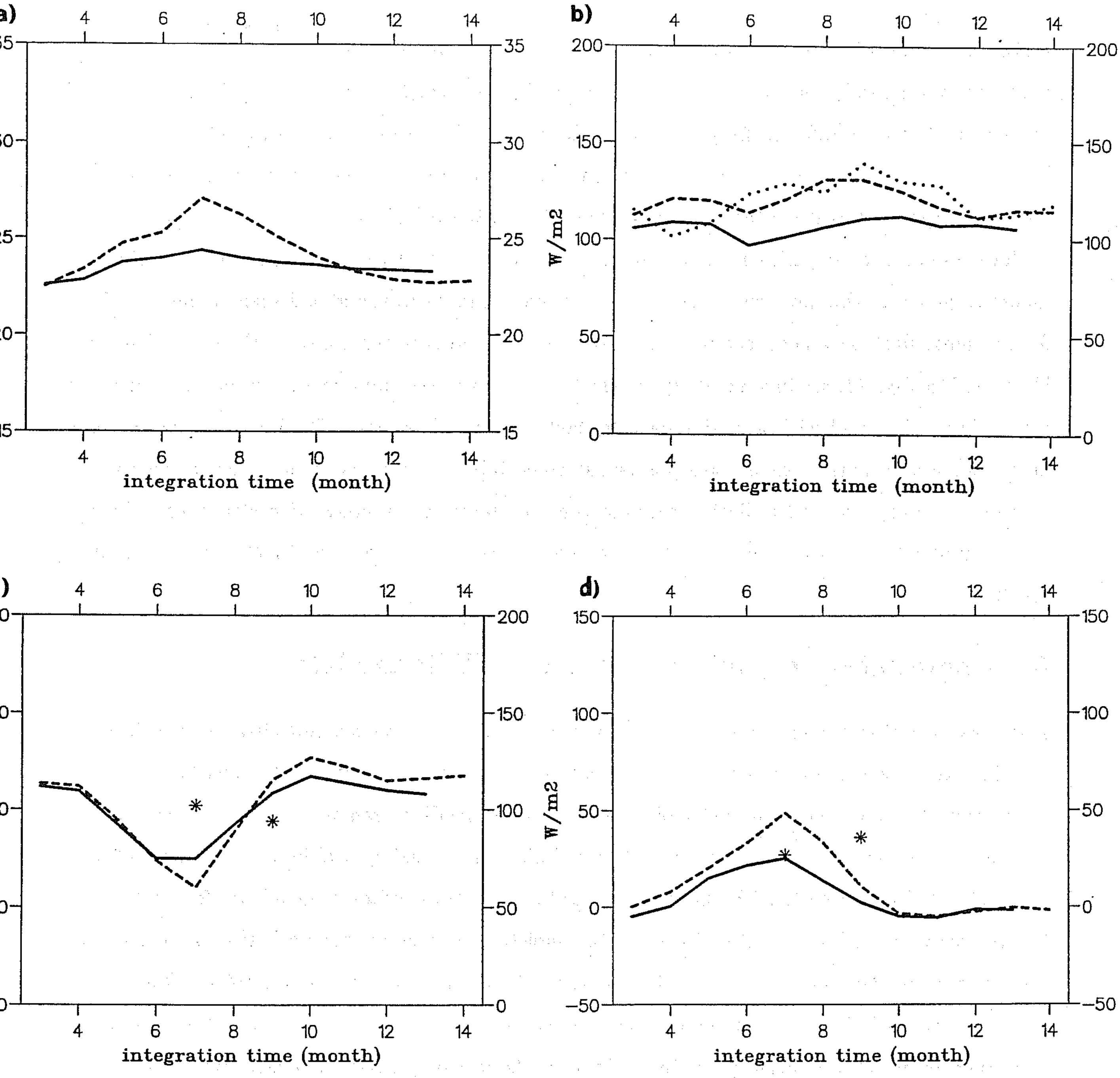


Figure 11: Monthly averaged Surface Temperature (a), Net Radiation (b), Latent Heat Flux (c), Sensible Heat Flux (d) over the Amazonian region. The solid lines refer to ISBA experiment and the dashed lines to the control experiment. Observed Net Radiation (dotted line), Latent and Sensible Heat Fluxes (stars) are also reported.

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 a correspondence table which account either of 1D calibration tests or of the bibliography. A parameter dependant averaging process has been then applied to reduce the horizontal resolution to the AGCM grid (near 2.8°).

Two annual cycle experiments have been carried out with and without ISBA. It is found that the model response to the new scheme is mainly regional, changes at the global scale being explained for the most part by albedo changes. The results over the Amazon forest with the new scheme illustrated in Fig. 11, are in general agreement both with similar numerical experiments performed with other GCM, and with available data set collected in this region. The impact of the scheme on the simulated surface fields also points out some interaction mechanisms associating surface processes, convection, and radiation through the representation of convective cloudiness. Details on the implementation and on the results of these experiments are given in Manzi and Planton (1992).

5 Assimilation of soil moisture for NWP models

Because of the slow time scale evolution of water content, any erroneous initialization of soil moisture in 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) and Bouttier et al. (1992a,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 1990b, Bouttier et al. 1992a) showed that screen-level atmospheric quantities contain a lot of information about the physics of 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. (1992a) 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 within a 3-D model, a 48-hour clear-sky period from the HAPEX-MOBILHY experiment (André et al. 1986) was selected. The results of the assimilation demonstrate that the convergence of the method is rapid either with a dry or wet soil moisture guess (Fig. 12). However, some limitations of the method where observed for places where the atmosphere above the point of interest do not contain information about soil moisture. Such

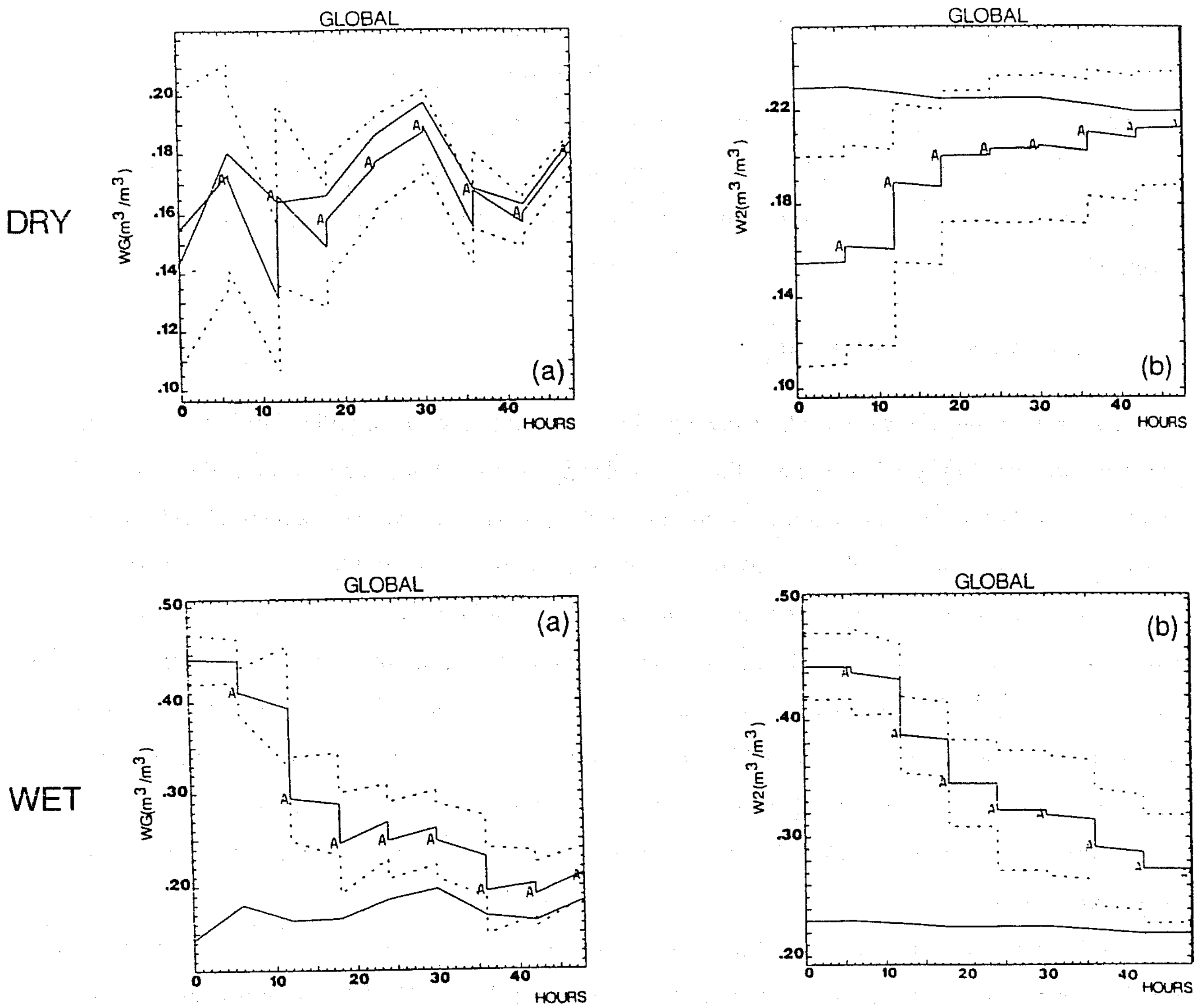


Figure 12: Evolution of the surface (a) and deep-soil (b) moisture contents during a 3D assimilation of 48 hours, starting from a dry guess and a wet guess. Dotted curves indicate the dispersion of the errors in the mesoscale domain. The evolution in the reference simulation is represented by the solid curves (Bouttier et al. 1992).

situations occured when mesoscale circulations generated 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 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 describes the various steps involved in the validation of a simplified 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 simulated 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 last decade over a broad variety of characteristic biomes. However, a recommendation is made for enhanced micrometeorological measurements in dry conditions to provide some guidance for improving the representation of surface exchanges in condition of low water content (soil water stress function and dew flux).

The implementation of ISBA within a meso-β-scale model provided the necessary framework for the interpretation of the various data collected during HAPEX-MOBILHY. 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 regularly during the experiment. Moreover, mesoscale modelling 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 dependant on forecast errors of near-surface air temperature and relative humidity.

All the developments reviewed in this paper should facilitate the ISBA implementation in the ARPEGE model promoted by Météo-France.

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