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Treatment of soil, vegetation and snow in land surface models: a test of the Biosphere–Atmosphere Transfer Scheme with the HAPEX-MOBILHY, ABRACOS and Russian data

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Abstract

Various components of the land surface, their individual hydrological processes and the process-oriented models are reviewed in this paper, with the focus on their application in global climate models (GCMs). The Biosphere–Atmosphere Transfer Scheme (BATS) is examined regarding its performance for three different surfaces (crop, forest and grass), with available data from HAPEX-MOBILHY, ABRACOS and Russian data sets. The simulations of the key land surface prognostic variables, such as soil moisture and snow cover, are examined in detail because such validation has been lacking. Using the HAPEX-MOBILHY data, the impact of errors in the forcing variables on the uncertainties in the partitioning of total run-off and evapotranspiration is investigated, and the influence of the periodic forcing on soil moisture simulations is examined. Furthermore, an alternative empirically based approach for the soil evaporation efficiency is tested. The current framework of BATS soil hydrology, vegetation and snow schemes adequately reproduces observed soil moisture profiles for the three surfaces considered, and captures the seasonal evolution of snow mass. The simulations can be enhanced when site-specific information on surface parameters is available. Because of the realism of the overall framework of BATS, its inclusion in a GCM [the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM)] leads to reasonably realistic simulations of surface hydroclimatological variables. Further improving surface hydrology in global climate models is dependent on thorough tests of the available models using the available data, on the collection of long-term, seasonal, high-quality data, both at point and on larger spatial scales, and on the effective representation of the surface types on GCM scales. © 1998 Elsevier Science B.V. All rights reserved.

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1. Introduction

The land surface consists of soil, vegetation, snow, glaciers, inland waters, mountains, animals, human beings, their shelters and much more. However, modern global climate models (GCMs) with a scale of 100 km × 100 km ''view'' the land surface simply as a mosaic of soil, snow, vegetation, inland water and orography. While the orography is treated independently in GCMs (e.g., McFarlane, 1987), the so-called

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land surface models (LSMs) deal specifically with the interactions among soil, vegetation, snow, inland water and their overlying atmospheres.

On a microscale of 10^{-2}–10^{3} m, or a local scale of 10^{2}–5×10^{4} m (Oke, 1987), detailed numerical models have been developed for soil (e.g., Sievers et al., 1983; Bach, 1992), vegetation (e.g., Shaw and Pereira, 1982; Meyers and Paw U, 1987; Raupach, 1989; Baldocchi, 1992; Kondo and Watanabe, 1992; Watanabe, 1993, and references therein) and snow (Anderson, 1976; Jordan, 1991). Such models are process oriented, designed to study the complicated transfer or diffusion of energy, water and trace gases within each study object and between each such object and the atmosphere. There are also complex models of lakes (e.g., Hostetler and Bartlein, 1990). These models may not be suitable for use in modern GCMs for three reasons: (1) in order to incorporate these models, GCMs should operate at resolutions commensurate to micro-scales or local scales, which would be computationally prohibitive with the current or near-future computing power; (2) if these models were directly implemented into the GCMs at the macro-scale (10^{4}–10^{6} m), their computations would overwhelm their atmospheric counterparts; and (3) these models require a large number of parameters that cannot be measured easily at global scales.

The development of LSMs compatible with modern GCMs has been an active research topic over the past decade. To date, there are approximately 30 LSMs available, and the number is growing quickly. Most of these are participating in the Project for Intercomparison of Land-surface Parameterizations (PILPS) (Henderson-Sellers et al., 1993; Henderson-Sellers et al., 1995). Generally, these LSMs are not very different from each other; all of them may be regarded as simplifications to different degrees of the detailed process-oriented models mentioned earlier. PILPS aims to compare how these ‘‘simple’’ models perform, both in the off-line mode and in the coupled mode (i.e. linked with GCMs), and examine how the performances of these LSMs can be improved.

Ideally, the performance of any LSM should be evaluated against observed field data collected for each vegetation type, soil type and the various climatic conditions before their implementation into any climate model. This so-called off-line mode allows use of observed, and hence realistic, data (in contrast to what may be computed by GCMs). This testing can identify any serious problems and improve the realism of the LSM code. As a result, a great deal of international effort has addressed this issue, including such projects as ABRACOS (Shuttleworth et al., 1991; Wright et al., 1996), ARME (Shuttleworth, 1988), BOREAS (Sellers et al., 1995), Cabauw (Beljaars and Holtslag, 1991), EFEDA (Bolle et al., 1993), FIFE (Sellers et al., 1992), GCIP (International GEWEX Project Office, 1993), HAPEX-MOBILHY (Andre et al., 1986), HAPEX-NIGER92 (Gash et al., 1991), LOTREX (Schadler et al., 1990), the Monsoon 90 (Kustas et al., 1991), the Russian data (Vinnikov and Yeserkepova, 1991) and SEBEX (Wallace et al., 1991). The number of such data sets is limited when compared to the range of possible environments. In addition, seasonal and multi-year measurements of surface fluxes are extremely rare. Therefore, the urgent need is not to develop a ‘‘new’’ LSM, but to evaluate and test the available models with the available data. In doing this, the available LSMs can be improved, and the results may provide guidance to the observational scientists.

The basic components of the land surface and their treatment in LSMs are reviewed in Section 2. Section 3 evaluates the Biosphere–Atmosphere Transfer Scheme (BATS) (Dickinson et al., 1986, 1993) as one example of an LSM. Furthermore, both soil moisture and snow simulations are compared against the HAPEX-MOBILHY, ABRACOS and Russian data. A summary of our work is presented in Section 4.

2. Review of treatments of soil, vegetation and snow in land surface models

2.1. Soil

In pure soil science, water flow in unsaturated or partly saturated soils has traditionally been described using the Richards equation. The soils were usually assumed to constitute a rigid porous medium in which air phase and temperature gradient were assumed to play insignificant roles in the liquid flow process. Philip and de Vries (1957) proposed a general theory of vapor, water and heat transfer within the soil. To obtain the vertical profiles of soil moisture and/or temperature, these equations need to be solved numerically under an appropriate set of boundary and initial conditions. Various versions of the schemes have been developed to solve these equations (e.g., Milly and Eagleson, 1982; Kool and van Genuchten, 1991; Bach, 1992). Despite the rapid development of sophisticated numerical and modeling systems, their successful application to actual field programs is somewhat limited because of the lack of information regarding the parameters of storage and transport coefficients entering the governing transfer equations (van Genuchten, 1980; Bach, 1992). For example, the unsaturated soil hydraulic properties and soil thermal properties are difficult to measure because most of the existing laboratory and field methods are relatively expensive, cumbersome and time consuming (van Genuchten et al., 1991). In particular, accurate in situ measurements of the unsaturated hydraulic conductivity and the soil thermal conductivity remain difficult to obtain.

A very popular alternative to direct measurements of the unsaturated soil hydraulic properties and soil thermal properties has been to use analytical functions, such as the moisture retention curve (e.g., Brooks and Corey, 1966; Clapp and Hornberger, 1978; van Genuchten, 1980), the moisture–hydraulic conductivity curve (e.g., Clapp and Hornberger, 1978; van Genuchten, 1980), and the relationship between the effective soil thermal properties and the volumetric fractions and thermal properties of the soil constituents (e.g., de Vries, 1963). The moisture–hydraulic conductivity curve is usually derived from statistical pore-size distribution models (e.g., Mualem, 1976). Once a set of mathematical formulations is selected, an associated set of parameters must be estimated for a practical application. In the soil physics community, these parameters are fitted to or estimated with field (in situ) data for retention and hydraulic conductivity or diffusivity using a nonlinear optimization technique, such as the Marquardt or Levenberg–Marquardt method preprogrammed in the RETC (RETention Curve) computer code distributed by the USDA Salinity Lab for such analysis (van Genuchten et al., 1991). This code can also be modified to account for more complicated flow processes, such as hysteric two-phase flow (Lenhard et al., 1991) or preferential flow (Germann, 1990).

The use of such techniques for a large region, say, of a GCM grid-square size, is impractical because it would require a vast amount of field data collected throughout the region. In the hydrological or atmospheric modeling community, an alternative approach has been to use the parameters tabulated as a function of texture. In all the modern LSMs, with the possible exception of the one by Abramopoulos et al. (1988), the soil moisture retention curve and the soil moisture–hydraulic conductivity curve are based on those in Clapp and Hornberger (1978). These formulations contain four parameters: the saturated soil water content (or soil porosity in the LSMs), the saturation suction (or retention), one shape factor and the saturation hydraulic conductivity. Clapp and Hornberger (1978) provided a table of mean values for each of these four parameters as a function of the 11 USDA (US Department of Agriculture) soil textural classes.

There has been little attention paid to the validity of these mean soil properties in the climate modeling community. The mean parameters were estimated from 1446 soil samples out of an initial set of over 1800 collected from 34 localities throughout the United States. There are large standard deviations for the saturated soil water content, the shape factor and the saturation suction, indicating the heterogeneous nature of soils. Ek and Cuenca (1994) recently studied how the natural variability of the shape parameter in the Clapp and Hornberger (1978) formulation affects the simulated surface fluxes and boundary layer development. They found that the results were very sensitive, especially when soil moisture was very low and for bare soil conditions.

Direct measurements of soil porosity, saturation soil suction, soil particle size distribution and saturation hydraulic conductivity are difficult to obtain. Soil moisture is also not widely and routinely measured. Soil classifications at regional and global scales have relied heavily on the dated UNEP atlas (e.g., Wilson and Henderson-Sellers, 1985; Zobler, 1986; Webb et al., 1993). The global distribution of the water table depth, the position of bedrock and frozen soil depths are all poorly known. These could be potentially important in studying soil–climate interaction in the context of global change. Future improvement of the soil hydrology in global climate models, therefore, may largely depend on the collection of these data and on how the effective soil parameters for GCM scales are derived.

2.2. Vegetation

In agricultural and forest meteorology, the detailed multi-layer canopy-scale process models have long been available to simulate vertical profiles of micro-meteorological entities (e.g., the mean field and flux) within and above the canopy. These models include the K theory (flux gradient theory) approach (e.g., van de Griend and van Boxel, 1989; Kondo and Watanabe, 1992, and references therein), the higher-order closure models (e.g., Shaw and Pereira, 1982; Meyers and Paw U, 1987) and particle trajectory (or Lagrangian) models (e.g., Raupach, 1989). The latter two approaches were developed to remedy the known problems associated with the K theory, such as its inability to account for the countergradient fluxes within the canopy air space and its inability to explain the secondary wind speed maxima observed in forest canopies.

Raupach (1991) indicated that six separate physical or biological components were needed to build a comprehensive canopy-scale process model. These six components are radiation physics, soil physics, interception, plant physiology, aerodynamics and turbulent transfer of scalars. The canopy models mentioned above usually focus on one or two of the six components. If a truly multi-layer canopy model were ever built, it would require a tremendous number of parameters, probably of the order of 100, many of which would be difficult to measure, even locally.

The canopy treatments in modern LSMs for use in atmospheric models are highly simplified. In particular, they use: (1) no more than two canopy layers (Dickinson et al., 1986; Sellers et al., 1986); (2) a big-leaf, big-stoma concept (Deardorff, 1978); (3) a two-stream approximation for shortwave transfer (Dickinson, 1983) or an even simpler approach; (4) K theory (Sellers et al., 1986) or a simpler approach for drag coefficients; and (5) bucket-type treatments for interception (Deardorff, 1978). Although they are more heavily parameterized compared to the above-mentioned multi-layer canopy models, their simulations of momentum, energy and vapor fluxes above the canopy can be similar to those from a detailed canopy process model. Pielke (1984) showed that the more expensive approach has not improved simulations of the resolvable dependent variables in the planetary boundary layer over those obtained from the best first-order representations. Based on both ''upward'' and ''downward'' influences, Raupach (1991) argued that the comprehensive, multi-layer canopy models are appropriate for the downward effect of a given climate on the vegetation microclimate, whereas the simpler models are useful in the upward effect of vegetation on climate. This view is also shared by Vogel et al. (1995), who compared a hierarchy of models for determining energy balance components over vegetation canopies. Watanabe (1993) compared the performances of K theory and a second-order closure model, and demonstrated small differences in the calculated fluxes above the canopy.

One of the keys to improving these upward influences in the LSMs is to derive effective vegetation parameters on the GCM scale from the in situ and remotely sensed data. What has been widely used by the GCM land modeling community is to assume that each grid box has one or several vegetation types that are prescribed from coarse grid data sets of land cover (e.g., Matthews, 1983; Olson et al., 1983; Wilson and Henderson-Sellers, 1985); each vegetation type is then related to a set of parameters through a look-up table. Recently, considerable progress has been made to update the vegetation parameters using field data (e.g., Kelliher et al., 1995) and to scale them up to the GCM scale using the satellite-derived land-cover data at 1 km × 1 km resolution (e.g., Loveland et al., 1991; Arain et al., 1997). Xue et al. (1996) found that the use of the satellite-derived land-cover data can improve simulations of the US weather and climate.

2.3. Snow

Snow is one of the most important variables affecting agriculture, hydrology, water resources and climate. Numerous observational studies have indicated that snow cover correlates with temperature (Namias, 1985), circulation pattern (Cayan, 1996) and monsoon rainfall (Hahn and Shukla, 1976). GCMs have projected 1.5–4.5°C warming in a doubled CO_2 climate, and much of the simulated warming occurs at high latitudes (Houghton et al., 1996). The existing numerical models for snow display a wide range of complexities. The most complex treatments may be those of Anderson (1976) and Jordan (1991), which are primarily

Table 1
List of BATS default parameters applicable to type A (HAPEX-MOBILHY; crop/mixed farming and soil index 5), type B (ABRACOS; tropical rain forest and soil index 10), type C (Yershov; short grass and soil index 6), type D (Tulun; short grass and soil index 8), type E (Uralsk; short grass and soil index 2), type F (Kostroma; short grass and soil index 12), type G (Khabarovsk; short grass and soil index 12) and type H (Ogurtsovo; short grass and soil index 7).

Symbols in the table are defined as follows: θ_s, soil porosity; s_w, fraction of θ_s at which permanent wilting occurs; ϕ_s, minimum soil suction (m); K_s, maximum hydraulic conductivity ($\times 10^{-6}$ m s^{-1}); B, Clapp and Hornberger ''B'' parameter; Z_u, depth of top soil layer (m); Z_r, rooting depth (m); Z_t, total soil depth (m); f_{root}, fraction of total roots in top soil layer; k_T, ratio of soil thermal conductivity to that of loam; r_{smin}, minimum stomatal resistance (m s^{-1}); z_{0c}, canopy roughness length (m); d, canopy zero displacement height (m); D_c, interception capacity per unit projected area (mm); A_{v0}, maximum value of vegetation cover fraction; S_v, seasonal range of vegetation cover fraction; LAI_{max}, maximum leaf area index; LAI_{min}, minimum leaf area index; SAI, stem area index; f_{vis}, light sensitivity factor used in calculating the dependence of stomatal resistance on visible solar flux (m^2 W^{-1}); D_f, inverse square root of leaf dimension (m$^{-1/2}$); $\alpha_{c,vis}$, canopy visible albedo; $\alpha_{c,nir}$, canopy infrared albedo; I_{clr}, soil color index.

Parameter	A	B	C	D	E	F	G	H
θ_s	0.45	0.60	0.48	0.54	0.36	0.66	0.66	0.51
s_w	0.300	0.487	0.332	0.419 (0.230)	0.119	0.542 (0.332)	0.542(0.300)	0.378
ϕ_s	− 0.20	− 0.20	− 0.20	− 0.20	− 0.03	− 0.20	− 0.20	− 0.20
K_s	8.9	1.6	6.3	3.2	80.0	0.8	0.8	4.5
B	5.5	9.2	6.0	7.6 (4.5)	4.0	10.8 (6.0)	10.8 (5.5)	6.8
Z_u	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1
Z_r	0.5	1.5	1.0	1.0	1.0	1.0	1.0	1.0
Z_t	1.6	5.0	5.0	5.0	5.0	5.0	5.0	5.0
f_{root}	0.3	0.8	0.8 (0.2)	0.8 (0.05)	0.8 (0.2)	0.8 (0.05)	0.8 (0.1)	0.8 (0.2)
k_T	1.1	0.8	1.0	0.9	1.5	0.7	0.7	0.95
r_{smin}	120	150	200 (100)	200 (85)	200 (100)	200 (85)	200 (100)	200 (110)
z_{0c}	0.06	2.00	0.05 (0.10)	0.05 (0.20)	0.05 (0.10)	0.05 (0.15)	0.05 (0.20)	0.05 (0.10)
d	0.0	18.0	0.0	0.0	0.0	0.0	0.0	0.0
D_c	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1
A_{v0}	0.85	0.90	0.80 (0.90)	0.80 (0.95)	0.80 (0.90)	0.80 (0.95)	0.80 (0.90)	0.80 (0.90)
S_v	0.6	0.3	0.1	0.1	0.1	0.1	0.1	0.1
LAI_{max}	6.0	6.0	2.0 (3.0)	2.0 (3.0)	2.0	2.0 (3.5)	2.0 (3.0)	2.0
LAI_{min}	0.5	5.0	0.5	0.5	0.5	0.5	0.5	0.5
SAI	0.5	2.0	4.0	4.0	4.0	4.0	4.0	4.0
f_{vis}	0.02	0.06	0.02	0.02	0.02	0.02	0.02	0.02
D_f	10.0	5.0	5.0	5.0	5.0	5.0	5.0	5.0
$\alpha_{c,vis}$	0.10	0.04	0.10	0.10	0.10	0.10	0.10	0.10
$\alpha_{c,nir}$	0.30	0.20	0.30	0.30	0.30	0.30	0.30	0.30
I_{clr}	2	3	2	2	2	2	2	2

oriented to the internal processes of snow. These models solve the nonlinear energy transfer equations and take into account the densification of snow and the retention and transmission of liquid water. These models are not suitable for use in GCMs because of computational limitations. The snow schemes used in GCMs include those by Dickinson et al. (1986), Verseghy (1991), Pitman et al. (1991), Loth et al. (1993), Marshall et al. (1994), Lynch-Stieglitz (1994), Pollard and Thompson (1995), Douville et al. (1995), and Bonan (1996). The multi-layer snow models of Loth et al. (1993) and Lynch-Stieglitz (1994) were shown to realistically simulate the profiles of snow density, temperature and water equivalent within the snowpack for specific locations. Walland and Simmonds (1996) found that the sub-grid-scale variations of topography have a large impact on snow cover in their GCM. Foster et al. (1996) intercompared the values of snow cover and snow mass simulated from seven GCMs with those derived from three remotely sensed data sets. Their results show that several of the models consistently underestimate snow mass, but that other models over-estimate the mass of snow on the ground. The models did a better job simulating snow conditions in the winter and summer than in the spring and fall. Further improving these snow models in GCMs requires adequate testing against the field data (Schlosser et al., 1997), explicit inclusion of the sub-grid-scale effect of topography (Walland and Simmonds, 1996; Arola and Lettenmaier, 1996) and accurate parameterization of snow masking over the vegetated surface (e.g., Donald et al., 1995; Yang et al., 1997).

The data for verification of snow models are available only at some stations. To be useful, the data should include snow albedo, depth, density, water content, temperature, thermal properties, and other site characteristics for vegetation and soil. For climate studies, long-term snow data and meteorological variables should both be available, especially the radiation components. Because these are not generally available, estimates need to be made based on screen level air temperature and humidity, and some measurements of cloud data. Cloud height and cloud base temperature are also not readily available. Therefore, the accuracy of the estimated radiation components will determine, to a large extent, the reliability of the modeled snow characteristics.

3. Validation of BATS for different surfaces

The current land surface models for GCMs, e.g., BATS (Dickinson et al., 1986, 1993) and SiB (Sellers et al., 1986), are necessary simplifications of the process-oriented models. No comprehensive evaluation of these LSMs took place prior to PILPS (Henderson-Sellers et al., 1993, 1995). We focus on BATS in this study and test its soil moisture simulations for three distinctive surface types using the field data: soya crop (HAPEX-MOBILHY), tropical forest (ABRACOS) and mid-latitude grass vegetation (six stations over the former Soviet Union, hereinafter referred to as the Russian data). The Russian data are also useful to test the snow sub-model in BATS. First, we used the default set of parameters in BATS to identify the deficiencies. Then, a series of experiments were performed to explore the model's sensitivity to altered values of parameters for vegetation, soil and snow (if present). Finally, some results from coupling BATS to a GCM are presented to illustrate the model's performance and limitations.

3.1. Testing BATS with the HAPEX-MOBILHY data

The HAPEX-MOBILHY data used in this study were collected at Caumont (SAMER No. 3, 43.68°N, 0.10°W). Detailed information concerning the site can be found in Andre et al. (1986) and Shao and Henderson-Sellers (1996). The vegetation type is a soya crop field, and the soil is loam. The BATS scheme has a vegetation class as crop/mixed farming and a soil type class as loam, both approximately corresponding to the features of the site. Because the default BATS parameters are those currently being used in the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM) integrations (Kiehl et al., 1996), it is interesting to examine the performance of BATS when the default parameters are applied to this site. Values of these parameters are given in Table 1.

Because the meteorological forcing data are available for only 1 year (from midnight 1 January to midnight 31 December) and the soil moisture measurements for the same year are available only starting from 7 January, the model was run to reach equilibrium with its initial soil moisture arbitrarily set to fully saturated on the initial 1 January. The equilibrium was defined as in Yang et al. (1995). To achieve this, the 1-year forcing data were looped through as many times as possible; we refer to this type of integration method as periodic forcing. The spin-up time is 4 years.

We examine how errors, if any, in forcing variables can contribute to the uncertainties in the partitioning of total run-off and evapotranspiration, and what the effects of the applied ''periodic forcing'' may be. To evaluate the model's sensitivity to measurement errors in forcing variables, we have considered, for each time step, a ± 2°C change in air temperature, a factor of 1.1 change in specific humidity, a factor of 2 change in wind speed, a $\pm 10\ \mathrm{W\ m^{-2}}$ change in incident shortwave or longwave radiation, a factor of 2 change in precipitation, and a ± 5 mb change in surface pressure. The prescribed change of radiation falls within the range of observational errors, while the changes for other variables are generally greater than are the ranges of observational errors (cf. Leese, 1993), but are typical of the differences between their observed values and those simulated by GCMs. One

Table 2
Sensitivity of the BATS model in terms of equilibrium annual surface run-off, base flow (or drainage) and evaporation to imaginary errors imposed, at every time step, in meteorological forcing variables with the HAPEX-MOBILHY data.

Symbols in the table are defined as follows: Control, the run with the standard forcing variables; T, air temperature (K); q, specific humidity (kg kg^{-1}); V, wind speed (m s^{-1}); $S\downarrow$, downward shortwave radiation (W m^{-2}); $L\downarrow$, downward longwave radiation (W m^{-2}); P_r, precipitation (kg m^{-2} s^{-1}); P_s, surface pressure (mb). Values in parentheses denote percentage difference, which is defined as [(Case − Control)/Control] × 100%.

Run	Surface run-off (mm year^{-1})	Base flow (mm year^{-1})	Evaporation (mm year^{-1})
Control	154.3	39.9	662.1
$T+2$	134.3 (− 13.0%)	23.5 (− 41.1%)	698.5 (5.5%)
$T-2$	193.9 (25.7%)	92.7 (132.3%)	569.3 (− 14.0%)
$q\times 1.1$	183.4 (18.9%)	74.9 (87.7%)	598.0 (− 9.7%)
$q\times 0.9$	133.0 (− 13.8%)	21.6 (− 45.9%)	701.7 (6.0%)
$V\times 2$	131.9 (− 14.5%)	19.4 (− 51.4%)	704.9 (6.5%)
$V\times 0.5$	166.6 (8.0%)	58.1 (45.6%)	631.6 (− 4.6%)
$S\downarrow +10$	144.9 (− 6.1%)	30.6 (− 23.3%)	680.7 (2.8%)
$S\downarrow -10$	163.5 (6.0%)	50.8 (27.3%)	643.6 (− 2.8%)
$L\downarrow +10$	143.0 (− 7.3%)	28.8 (− 27.8%)	684.4 (3.4%)
$L\downarrow -10$	165.4 (7.2%)	53.4 (33.8%)	637.4 (− 3.7%)
$P_r\times 2$	534.8 (246.6%)	353.4 (785.7%)	824.3 (24.5%)
$P_r\times 0.5$	31.5 (− 79.6%)	1.0 (− 97.5%)	395.7 (− 40.2%)
P_s+5	155.9 (1.0%)	41.8 (4.8%)	658.8 (− 0.5%)
P_s-5	153.1 (− 0.8%)	38.6 (− 3.3%)	664.6 (0.4%)

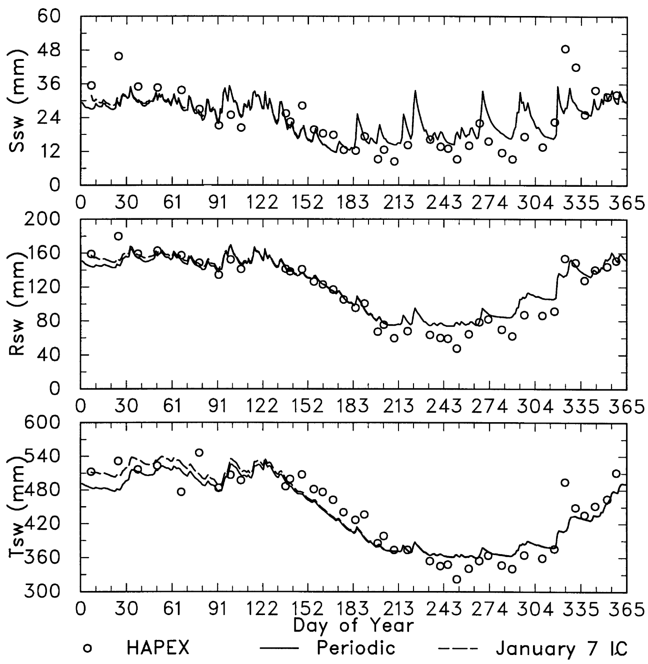

Fig. 1. Daily averages of soil water content in mm for three layers as simulated by the BATS model compared with HAPEX data (weekly measurements). The same layers apply to both simulated (lines) and observed (circles). The simulated results are shown for two runs, one with default parameters and periodic forcing (solid lines), the other also with default parameters but with the model starting on 7 January using the observed soil moisture contents from then up to 31 December (dashed lines).

Table 3
Soil moisture statistics for BATS output at equilibrium year compared to observations in the HAPEX-MOBILHY data.

Symbols in the table are defined as follows: S_{sw}, surface soil moisture content; R_{sw}, root-zone soil moisture content; T_{sw}, total column soil moisture content. The control is for a run with the BATS default parameterizations, while the β run denotes a run that is the same as the control, but the β factor is defined by eqn (1). Both RMS errors and correlation coefficients are calculated using observations, as shown in Fig. 1.

Run	S_{sw}	R_{sw}	T_{sw}
RMS errors (mm)			
Control	7.11	12.89	27.16
β	6.70	11.73	27.70
Correlation coefficients			
Control	0.76	0.95	0.93
β	0.83	0.95	0.92

run is performed for each single change of one variable.

The results at equilibrium are given in Table 2. Over all the cases considered, annual evaporation varies over a wide range of 400–825 mm, i.e. from − 40% to 25% different from the control. This range is determined by the factor of 2 ''errors'' imposed in precipitation. Without the precipitation cases, the range is reduced to 570–705 mm, or − 14% to 7% different from the control. The increase in evaporation (6%) in the $T + 2$ experiment is smaller than the decrease of evaporation (14%) in the $T - 2$ experiment. This asymmetry is due to the limiting effect of soil moisture and other stress factors on evaporation in parameterizations of evaporation common in current LSMs (Qu et al., 1998). The asymmetry also appears in the cases of changing specific humidity, wind speed and precipitation amount, but is not seen in the cases of changing downward solar radiation, downward longwave radiation and surface pressure. If a change of 10% in evaporation is used as a threshold to determine whether the ''errors'' in the forcing variables are tolerable or not, we see that the intolerable errors would be a 2 K cold bias in air temperature and factor of 2 changes in precipitation, while a 10% increase in the specific humidity is right at the threshold. If we use a change of 5% in evaporation as a threshold, the intolerable errors include the previously mentioned terms plus a 2 K warm bias in air temperature, a 10% decrease in the specific humidity and a factor of 2 increase in wind speed, while a factor of 2 decrease in wind speed is right at the threshold. However, the changes of 10 W m^{-2} in solar and longwave radiation, and of 5 mb in surface pressure, would be tolerable for both thresholds. As far as the quality of the forcing data used in this study is concerned, the measurement errors are unlikely to be as serious as those imposed above, and there would be small errors in all, rather than just one, of the forcing variables. Hence, their effects on the model's simulations should be small.

To address the question of periodic forcing, another run was performed by integrating the model from 7 January, when the soil moisture data were available for initialization and the integration was up to 31 December. The resulting soil moisture profiles are compared with those from the simulation with the periodic forcing (Fig. 1). The BATS scheme, with its default parameters, is capable of producing realistic simulations of soil moisture in three layers: surface layer, rooting zone and total column. The wet period during the winter, the drying period during the growing phase (days 120–210) and the dry period (days 210–300) are well captured. When using the measured soil moisture contents as the initial condition, there is a slight improvement for the first few months, suggesting that there is a slight difference between the antecedent precipitation and the precipitation used in the periodic forcing, and that the observed initial soil moisture reflects the influence of the antecedent precipitation. In the later part of the year, the differences in soil moisture between the two runs are indistinguishable. These results show that BATS equilibrates to a state independent of the initial condition and that if the initial condition is changed and the model is forced with the same periodic forcing, the same equilibrium state will be achieved. This suggests that if the interannual variations of precipitation are small, the (off-line) periodic forcing procedure might be an effective method to overcome the problem of lacking soil moisture measurements for initialization, provided that the meteorological forcing is available for 1 year and that the modeled soil moisture profile at equilibrium is realistic.

The soil moisture simulations can be improved using a different parameterization for the β factor, i.e. the ratio of actual to potential evaporation at the

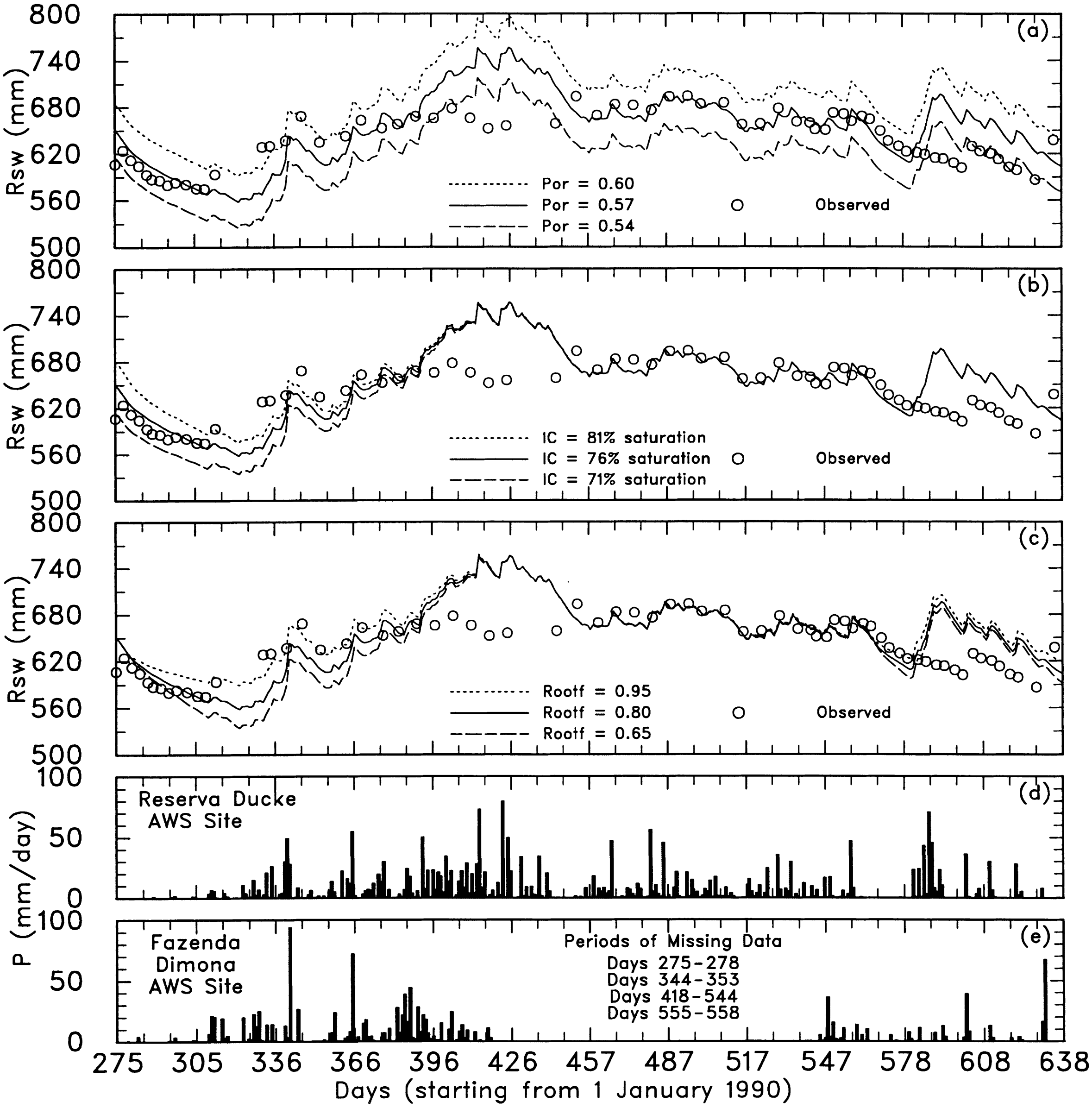

Fig. 2. Daily averages of soil water content in mm in the root zone (1.5-m layer) for the Reserva Ducke forest site as simulated by the BATS model (lines) compared with the measurements (circles) taken from a forest site near Fazenda Dimona (at an interval of 3–8 days). The model's sensitivity to porosity (a), initial soil moisture content (b) and top soil layer root fraction (c) is illustrated. Daily accumulated rainfall is also shown for the Reserva Ducke forest site (d) and the Fazenda Dimona pasture site (e).

soil surface. In BATS, β is obtained by a demand–supply approach (Dickinson et al., 1993), in which the diffusion-limited maximum evaporation (i.e. supply) is a nonlinear function of soil moisture content within both surface and root-zone soil layers, the thickness of both surface and root-zone soil layers, and soil hydraulic properties (θ_s, ϕ_s, K_s, B and diffusivity). According to a recent review by Mihailovic et al. (1995), β is more commonly represented by a much simpler function of surface soil moisture content only.

We test that of Deardorff (1978) with the form:

$$\beta = \min(1, S_{sw}/S_{sw,fc}), \quad (1)$$

where S_{sw} is surface soil moisture content (mm), and $S_{sw,fc}$ is the field capacity at the surface soil layer (mm) and its estimate for the site is 32 mm (Shao and Henderson-Sellers, 1996). As shown in Table 3, the new run leads to better simulations of surface and root-zone soil moisture (i.e. smaller RMS errors and higher correlation) than the control run. The improvements are most notable in the dry seasons (days 150–300), during which the overestimation of surface and root-zone soil moisture in the control (cf. Fig. 1) is reduced considerably. Consequently, daily mean evaporation increases by up to 40 W m^{-2} in the new run during that period. We interpret the improvement in the simulations from using eqn (1) as primarily due to the direct use of the observed field capacity, which is not explicitly a parameter in the default BATS formulations, but which could be estimated in BATS from the other given soil hydraulic parameters (see next section). Therefore, the BATS estimated field capacity may be different from that observed. Because the field capacity and other soil hydraulic parameters are equally difficult to specify in GCM scales, we will not use eqn (1) as a permanent change in the BATS formulations.

Further evaluation of BATS with the HAPEX-MOBILHY data can be found in Yang and Dickinson (1996), in which the model's vegetation scheme and the run-off formulations are tested.

3.2. Testing with the ABRACOS data

The ABRACOS data (Shuttleworth et al., 1991; Wright et al., 1996) that we used are taken from the Manaus study area, in which both forest and pasture sites are instrumented. The Reserva Ducke site (2.95°S, 59.95°W) is an area of protected primary forest 25 km northeast of Manaus, while the Fazenda Dimona site (2.32°S, 60.32°W) is a 10-km^2 clearing characterized by pasture grasses at 100 km north of Manaus (Wright et al., 1996).

While the meteorological forcing data were collected from the automatic weather station (AWS) at each site for more than 3 years, we used only the first year (2 October 1990–30 September 1991 for the Reserva Ducke site, and 29 September 1990–28 September 1991 for the Fazenda Dimona site), because only they were available at the time of paper preparation. The soil moisture was measured at irregular intervals of 3–8 days close to the AWS using a neutron probe soil moisture meter, except that for the Manaus forest data it was recorded in primary forest close to the Fazenda Dimona site rather than at the Reserva Ducke site. The default parameters for forest as currently used in the NCAR CCM are given in Table 1.

The model with the default parameters reproduces the gross features of seasonal variations of root-zone soil moisture, but is consistently too wet (Fig. 2a). A reduction in porosity from 0.60 (default) to 0.57 significantly improves the simulations. A further reduction of porosity to 0.54 makes the soil layer generally too dry. For all three runs, root-zone soil moisture is consistently ''overestimated'' around February and September 1991, corresponding to heavy rainfall (Fig. 2d). This overestimation may not be a problem of the model, but is more likely due to the mismatch of the soil moisture measurement site near Fazenda Dimona and the meteorological forcing measurement at Reserva Ducke. An examination of the rainfall data at the Fazenda site (Fig. 2e) reveals that there were much smaller rainfall amounts around February and September 1991.

The impacts of slight changes in initial soil moisture are illustrated in Fig. 2b, in which the runs use porosity at 0.57 but other parameters remain at default values. A 5% change in the initial soil moisture had a considerable impact on the first 4 months of the subsequent simulations.

Fig. 2c shows the effects of specification of root fraction in the top soil layer. The model is sensitive to this parameter only during the dry seasons. Because the root fraction and rooting depth are closely linked, they need to be specified with care. We use the default value of rooting depth of 1.5 m to directly compare with the soil moisture measurements that were limited to 2 m before 17 October 1991. Observations (Wright et al., 1996) suggested that the rooting depth for the Amazonian tropical rain forest can be much larger than 1.5 m and even extend to 8 or 16 m deep.

3.3. Testing with the Russian data

Data for meteorological observations, snow cover

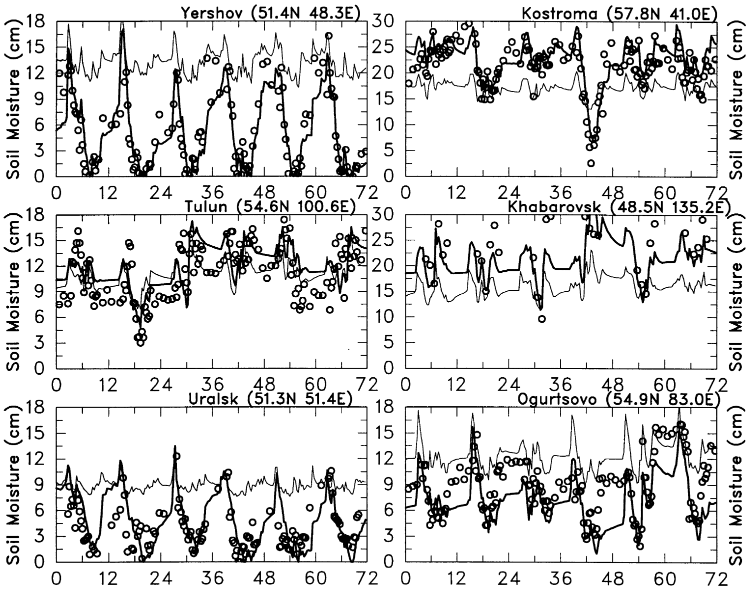

Fig. 3. Simulations and observations of root-zone (1-m layer) available soil moisture in cm (i.e. actual soil moisture minus soil moisture at wilting point) for the six stations for the entire 6 years. The soil moisture measurements (shown by circles) were made every 10 days in the warm season and each month during the winter. The simulations (shown by solid lines) are for two runs, one with default parameters (thin lines) and the other with modified parameters (thick lines). The abscissa labels indicate the beginning of the months for the period of 1978–1983.

and soil moisture in the former Soviet Union (FSU) are described in Robock et al. (1995). As used here, they cover the period from 1978 to 1983. The hydrometeorological stations were located on plots with grass vegetation. Each of the plots was a flat piece of land with an area ≥ 0.10 hectare, and the soil type was representative of the main soil type and landscape of the region. Soil moisture was observed every 10 days in the warm season and each month during the winter (Robock et al., 1995). Snow depth data were collected along snow courses, transects of 1–2 km in the vicinity of these stations, on the 10th, 20th and last day of each month during the winter, and were averaged (Yang et al., 1997). The default parameters for each station (1) are assigned according to the following considerations. Vegetation parameters are specified according to the default values for short grass in the BATS category; soil parameters are assigned to ensure that the model field capacity is close to observations, as indicated by the maximum values shown in Fig. 3. The model field capacity (or soil plant-available water-holding capacity), W_f, for 100-cm soil may be computed as follows:

$$W_f = 100(\theta_f - \theta_w), \tag{2}$$

where θ_f is the volumetric soil water content at the field capacity and θ_w the volumetric soil water content at the wilting point. Both are computed following the Clapp and Hornberger (1978) formulation and assuming the drainage by gravity, K_r, is 1.5 mm day^{-1}, as

$$\theta_f = \theta_s(K_r/K_s)^{1/(2B+3)} \tag{3}$$

and

$$\theta_w = \theta_s(\phi_s/\phi_w)^{1/B}, \tag{4}$$

where the symbols are defined in Table 1 and it is assumed that $\phi_w = -150$ m.

For the 12 BATS soil classes from sand (1) to clay (12) (Dickinson et al., 1993), W_f ranges from 10 to 20 cm (Fig. 4), which is adequate to describe the variations of the observed available soil moisture contents for Yershov, Tulun, Uralsk and Ogurtsovo. This is, however, not the case for Kostroma and

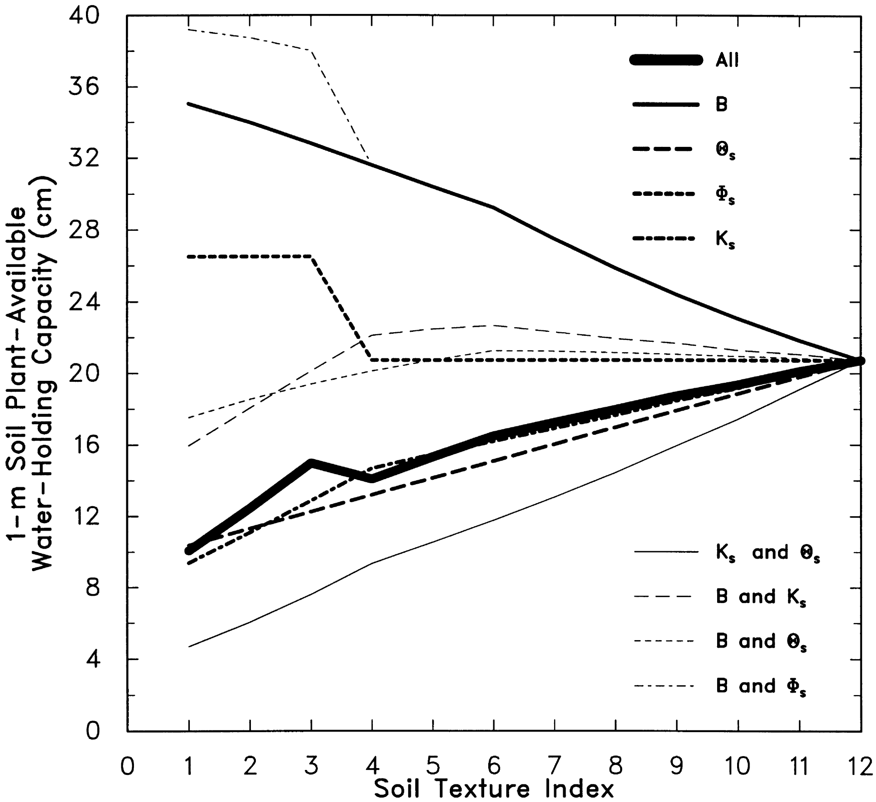

Fig. 4. The field capacity as a function of soil texture index from 1 (sand) to 12 (clay). The symbol ''All'' indicates that all the parameters (B, θ_s, ϕ_s and K_s) vary with the index, while ''B'' means that only B varies with the index; the other symbols follow the same rule.

Khabarovsk, because the observed maximum range is close to 30 cm. We sought a reasonable way to increase W_f while maintaining the values of the key soil properties (θ_s, K_s, ϕ_s and B) within the given 12 BATS classes. Therefore, we used the values at class 12 and varied one, two or three of the four parameters across the 12 classes to determine how W_f would change with each combination. Fig. 4 illustrates a spread of the nine typical cases, including the default case, in which all four parameters change as the class changes. The most straightforward way to find a value that is close to 30 cm is by varying B. Thus, we can choose a suitable value of B and the resulting θ_w for Kostroma and Khabarovsk (see the values in parentheses in Table 1).

For this site, we used a version of BATS which had the improved parameterizations of snow density and snow cover fraction, as described in Yang et al. (1997). We also applied the wind correction to the winter precipitation measurements, following Yang et al. (1997), and assumed the rain–snow transition temperature to be 0°C. The model was run to reach equilibrium with the given initial soil moisture. This was achieved by looping through the first year forcing data a number of times (typically 10 years or less; see Yang et al., 1995), after which the entire 6 years of data were used to drive the model. Only the results from the last 6 years were analyzed.

The model with default parameters gives soil too wet for Yershov, Uralsk and Ogurtsovo, but too dry for Kostroma and Khabarovsk (3). The observations show that the available soil moisture can drop to zero during the summer for Yershov and Uralsk, suggesting that the evapotranspiration might be too low with the default parameters. Therefore, the vegetation parameters for a generic short grass in CCM3/BATS were adjusted. The key vegetation parameters that were changed are A_{v0} (increased), LAI_{max} (increased),

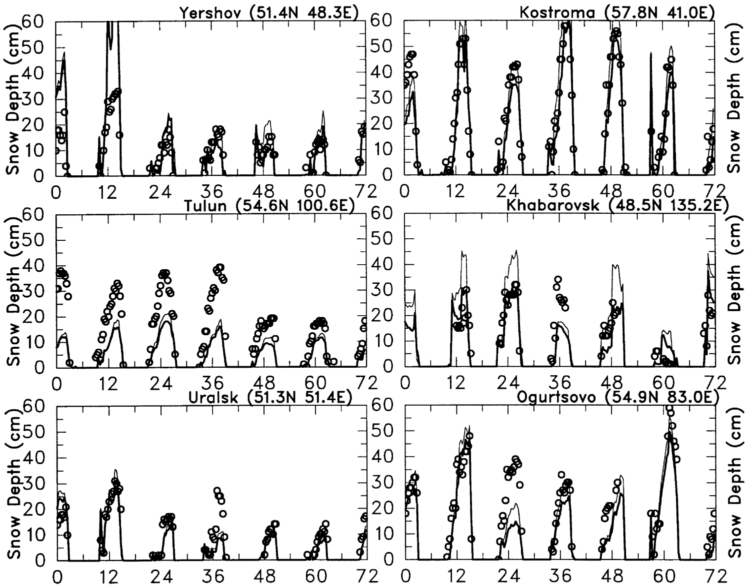

Fig. 5. Simulations and observations of snow depth in cm for the six stations for the entire 6 years. The snow depth measurements (shown by circles) were made from snow courses every 10 days during the winter. The simulations (shown by solid lines) are for two runs, one with default parameters (thin lines) and the other with modified parameters (thick lines). The abscissa labels indicate the beginning of the months for the period of 1978–1983.

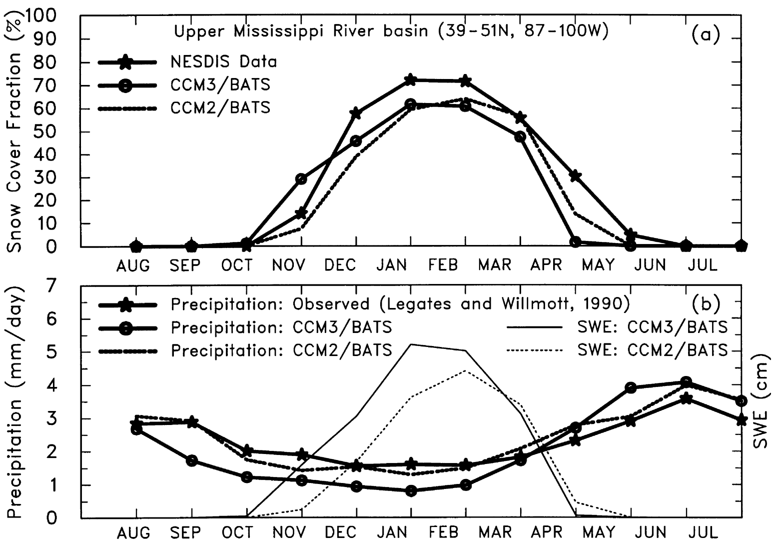

Fig. 6. Comparison of climate model simulations and long-term observations for snow cover fraction (a) and precipitation (b) over the upper Mississippi River basin (39–51°N, 87–100°W). The models are the CCM2/BATS and the CCM3/BATS. The sources for observations are NESDIS for snow (1973–1994 average) and Legates and Willmott (1990) for precipitation. The SWE from the two versions of CCM is also shown in (b).

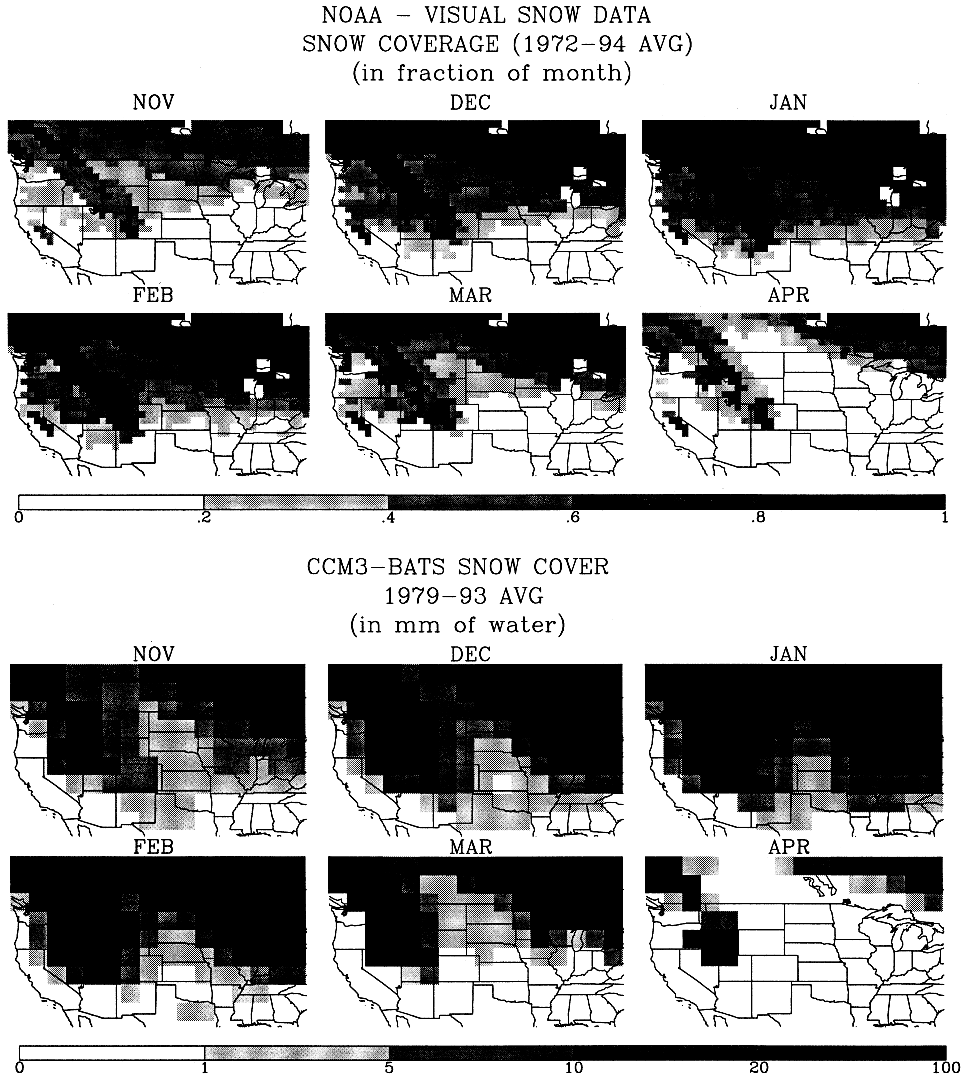

Fig. 7. Preliminary comparison of snow cover (in mm of water) from the CCM3/BATS over the USA/Canada regions and snow cover (in fraction of a month), which is averaged from the satellite pentad data (Robinson et al., 1993).

z_{0c} (increased), r_{smin} (decreased) and f_{root} (decreased). The simulations from the modified parameters are significantly improved except for Tulun, where the improvements are modest. Our results are qualitatively consistent with those from Xue et al. (1997), in the sense that they also adjusted parameters, but they were only for soil hydraulic properties (K_s, B and the logarithm of soil water potential at wilting point); they did not report the values of vegetation parameters in their study.

Yang et al. (1997) presented a detailed comparison of the modeled and observed snow variables (snow depth, snow water equivalent or SWE, surface temperature and surface albedo) for all six stations. Fig. 5 shows the time series of only snow depth as an example. The model gives good simulations with both types of parameters and for all the stations except for Tulun. The underestimation of snow depth in Tulun may be because the observations were taken in a clearing in a forest, which may present biological factors in the observations (Robock et al., 1995). For Khabarovsk, the modified parameters result in much better simulations than do the default parameters.

3.4. Results from the NCAR CCM coupled with BATS

The off-line evaluations and tests of BATS against observations as presented above show that BATS is capable of reproducing soil moisture and snow reasonably well compared with field data. Other validation studies using other field data show that BATS can realistically simulate seasonal variations of surface heat fluxes and that the simulations can be significantly improved by use of field-specific parameters (cf. Unland et al., 1996). These results indicate that the BATS parameterizations may be capable of describing the primary land surface processes, and so one would expect BATS coupled with a GCM to capture the general features of the surface hydroclimatological variables. To illustrate this, we show the modeled precipitation and snow mass or extent from BATS as linked to the NCAR GCMs (CCM2 and CCM3) at T42 resolution (approximately 2.8° × 2.8°, or 300 km × 300 km), and compare it with available observational data. Fig. 6a shows that the simulations of time and magnitude from both versions of the model reproduce the NESDIS data fairly well (Ropelewski, 1995, personal communication). However, the peak model values are lower than those observed, and the ablation occurs earlier by about 1 month. The CCM2/BATS did a better job of simulating precipitation for this region than did the CCM3/BATS throughout the year (Fig. 6b), but it gave a lower SWE for most of the snow season than did the CCM3/BATS. The latter arises because CCM2 has excessive surface solar radiation and warmer-than-observed surface air temperatures (Kiehl et al., 1996), which contribute to a larger loss of snow mass. Compared with the satellite snow extent data, the CCM3/BATS can capture the broad pattern of snow distribution, but it fails to reproduce the details of snow lines and some regional features (Fig. 7).

4. Conclusions

This paper first reviews the treatments of the three basic land surface components: soil, vegetation and snow. Although the detailed process-oriented models of each component are available, they may not be suitable for application in GCMs, which require coarse resolution LSMs having two to 10 layers of soil, one to two layers of vegetation and one to five layers of snow. Although necessarily simpler than the process-oriented models, these LSMs must not compromise basic processes. Therefore, their effectiveness and robustness must be tested by comparing with field data. Long-term, high-quality field data are crucial for improving LSMs in GCMs.

Second, the paper focuses on validation of one LSM, the BATS model, for three different surfaces (crop, grass and forest). The simulations of soil moisture and snow cover are emphasized. Spin-up is discussed. The impact of the periodic forcing has been examined using the HAPEX-MOBILHY data. BATS produces similar simulations of soil moisture when using either periodic forcing or the observed soil moisture data to initialize the model. These results suggest that if the interannual variations of precipitation are small, the (off-line) periodic forcing procedure might be effective in overcoming the problem of lacking soil moisture measurements for initialization, provided that the meteorological forcing is available for 1 year and that the modeled soil moisture profile at equilibrium is realistic.

The impacts of errors in the forcing variables on the spread of the partitioning of total run-off and evapotranspiration are examined. If we use a change of 10% in evaporation as a threshold to determine whether the ‘‘errors’’ in the forcing variables are tolerable or not, then intolerable errors are a 2 K cold bias in air temperature and factor of 2 changes in precipitation, and a 10% increase in the specific humidity corresponds to the evaporation threshold of 10%. If we use a change of 5% in evaporation as a threshold, besides the above, the intolerable errors include a 2 K warm bias in air temperature, a 10% decrease in the specific humidity and a factor of 2 increase in wind speed, while a factor of 2 decrease in wind speed is right at the threshold of 5%. However, even at 5%, changes of 10 W m^{-2} in solar and longwave radiation, and of 5 mb in surface pressure, are tolerable.

According to our tests using the HAPEX-MOBILHY data, the simpler soil moisture content approach for the evaporation efficiency gives slightly better simulations of soil moisture than does the physically based demand–supply approach. This is because the simpler approach uses the observed field capacity directly, which is not explicitly a parameter in the default BATS formulations, but which can be estimated in BATS from the other given soil hydraulic parameters. Because the field capacity and other soil hydraulic parameters are equally difficult to specify in GCM scales, we will not use this simpler approach as a permanent change in the BATS formulations.

The current framework of BATS soil hydrology, vegetation and snow schemes adequately reproduces observed soil moisture profiles for the three surfaces considered, and captures the seasonal evolution of snow mass. The simulations can be enhanced when site-specific information on surface parameters is available. Because of the realism of the overall framework of BATS, its linking with CCM leads to reasonably realistic simulations of surface hydroclimatological variables.

BATS should still be tested with as many field data as possible, and the optimum vegetation and soil parameters used in BATS should be systematically evaluated. Such carefully developed models are needed to allow derivation of effective parameters at GCM scales from remote sensing land-cover types and the optimum parameters.

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