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MODELING LAND SURFACE PROCESSES IN SHORT-TERM WEATHER AND CLIMATE STUDIES

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Land exchanges momentum, energy, water, aerosols, carbon dioxide and other trace gases with its overlying atmosphere. The land surface influences climate on local, regional and global scales across a wide range of timescales. This review concentrates on the rapid (i.e., seconds to seasons) biophysical and hydrological aspects of land surface processes. This paper provides the historical development of land surface models designed for short-term weather and climate studies, ranging from the early, simple "bucket" models to recent sophisticated soil-vegetation-atmosphere transfer schemes. Major research issues are reviewed by grouping into datasets, coupling to atmospheric models, component processes, and sub-grid-scale variability and scaling. Significant problems remain to be addressed, including the difficulties in parameterizing hillslope runoff, fractional snow cover, stomatal resistance, evapotranspiration, and sub-grid-scale variability and scaling. However, further progress is expected as the results of large-scale field experiments and satellite datasets are exploited.

1. Introduction

Land covers about 30% of the Earth's surface. The land surface consists of soil, vegetation, snow, glaciers, inland water, mountains, animals, human beings, their shelters, and much more. Land surface processes, in principal, refer to the exchanges of heat, water, CO_2, and other trace constituents among these components. In particular, land-atmosphere interaction refers to the exchanges of momentum, energy and mass (water, aerosols, and other important chemical constituents) between land surfaces and the overlying atmosphere. Land and oceans are coupled through river runoff and breezes that develop near the coast.

Unlike other surfaces of the Earth, land plays a distinctive role in weather and climate. Land has considerable heterogeneity such that bare soil, rock, short grass, tall grass, trees and snow patches can all coexist in one small area (Figure 1). This surface variability not only determines the microclimate but also affects the mesoscale atmospheric circulation (Giorgi and Avissar 1997; Pielke 2001). Land provides humans with a habitable place, which explains why over land there has been more historical documentation of routine observations of climatic variables such as temperature, clouds, and precipitation. Due to much less storage and negligible horizontal transport of heat over land, the diurnal, synoptic and seasonal variations of temperature are greater than those over water surfaces. The land orography not only exerts thermal and dynamic influences on weather and climate, but also has a direct effect on the surface hydrologic cycle (e.g., hillslope evaporation and runoff, Figure 1). The surface roughness over land is much greater than that over the oceans and affects greatly the heat and water transfer between the land surface and the overlying atmosphere. Land surfaces are more changeable than ocean surfaces. Such examples include urbanization, agricultural use, irrigation, deforestation, desertification, and bush fire as well as large natural seasonality in vegetation greenness and snow cover. Land surfaces store large quantities of carbon (more than twice that in the atmosphere). About half of the current anthropogenic emissions of CO_2 are being absorbed by the ocean and by land ecosystems, but there is a concern over how the terrestrial carbon sink is sensitive to climate and whether the sink will become a source due to global warming.

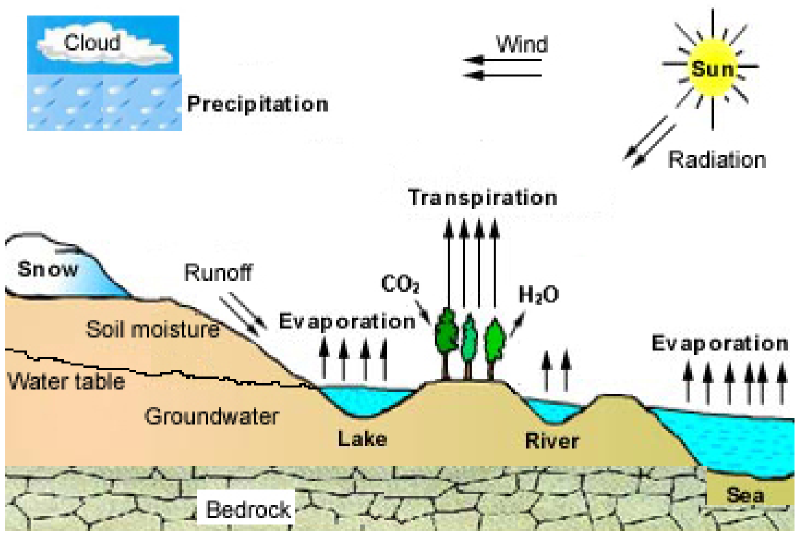

Figure 1. Schematic representation of the coupling between land, the atmosphere and oceans through the hydrologic cycle.

Not only should the land surface play an important role in weather and climate from the above intuitive reasoning, the evidence is increasing that the influence of the land surface is significant on local, regional and global climate on timescales from seconds to millions of years. On timescales of seconds to hours, land-atmosphere interaction is dominated by the rapid biophysical and biogeochemical processes that exchange momentum, energy, water, carbon dioxide and other chemical constituents between the land surface and the atmosphere (Jarvis 1976; Farquhar and Sharkey 1982; Dickinson 1983, 1992; Sellers 1992; Garratt 1993; Giorgi and Avissar 1997; Pielke 2001; Pitman 2003). On timescales of days to seasons, land-atmosphere interaction occurs through changes in the store of soil moisture (Yang 1995; Chen et al. 2001; Koster and Suarez, 2001), changes in snowpack (Robock et al. 2003), changes in carbon allocation, and vegetation phenology (e.g., budburst, leaf-out, senescence, dormancy) (Foley et al. 1996; Lu et al. 2001; Dickinson et al. 1998, 2002). Liu (2003) reviews monthly and seasonal variability of the land-atmosphere system. On timescales of years to centuries, vegetation structure and function (e.g., disturbance, land use, stand growth) is strongly determined by climate influences, primarily through temperature ranges and water availability (Foley et al. 1996). During the early to mid-Holocene period (6,000 years ago), important interactions and feedbacks between climate, vegetation, soil, snow, and lake may have taken place (Pielke et al. 1998). During the late Pleistocene period, glacial-interglacial cycles probably involve changes in the geographical distribution of vegetation, soils, surface albedo, and biogeochemical cycling in response to orbitally-induced insolation variations. On even longer geological timescales spanning hundreds of millions of years, the Earth's climate has been tightly coupled to atmospheric CO_2 levels through the carbonate-silicate cycle and/or the organic carbon cycle (Pielke et al. 1998).

This paper concentrates on land surface processes on timescales of a year and less. These timescales also are focused by international programs of Global Energy and Water cycle Experiment (GEWEX), GEWEX Americas Prediction Project (GAPP), and GEWEX Asian Monsoon Experiment (GAME), whose goals are to predict climate and hydrology on intraseasonal to interannual timescales. This article reviews the development of land surface models (LSMs) designed for use in three-dimensional weather and climate models with a focus on the rapid (biophysical) and intermediate (out to around a year; biogeochemical, hydrological, and phenological) processes (Figure 2). Understanding these processes is also central to an effective coupling with long-term biogeochemical cycles and vegetation dynamics because an accurate modeling of the latter two processes depends on credible simulations of canopy temperature, soil temperature and soil moisture.

2. Early Land Surface Models

2.1. *Bucket Model*

While biophysically realistic land surface processes are clearly important in weather and climate modeling, they were not included in atmospheric general circulation models until the late 1980s (Dickinson et al. 1986; Sellers et al. 1986; Abramopoulos et al. 1988).

In early GCMs, the land surface parameterization schemes are represented in a very crude way. Global soil was assumed to have a fixed water holding capacity of 15 cm (Manabe 1969). At each land grid square and each time step, the "bucket" is filled with precipitation and emptied by evaporation. The excess above its field capacity or a critical value is termed runoff. The evaporation rate is a product of the coefficient (the β function) and the potential evaporation. The coefficient, commonly called "soil wetness", or "moisture availability", is assumed to be a linear function of soil moisture content.

The empirical basis for the bucket parameterization is diurnally averaged data. It is most useful in GCMs that use diurnally averaged solar heating and therefore may not be appropriate for climate models with a diurnal cycle of solar radiation. Sato et al. (1989) found that the constant moisture availability in the bucket model can make the estimate of evapotranspiration several times too large compared to the more realistic Penman-Monteith equation (Monteith 1981). Sellers and Dorman (1987) found that evapotranspiration changes nonlinearly with the soil moisture especially when the soil is drying and when soil wetness drops below about 0.5.

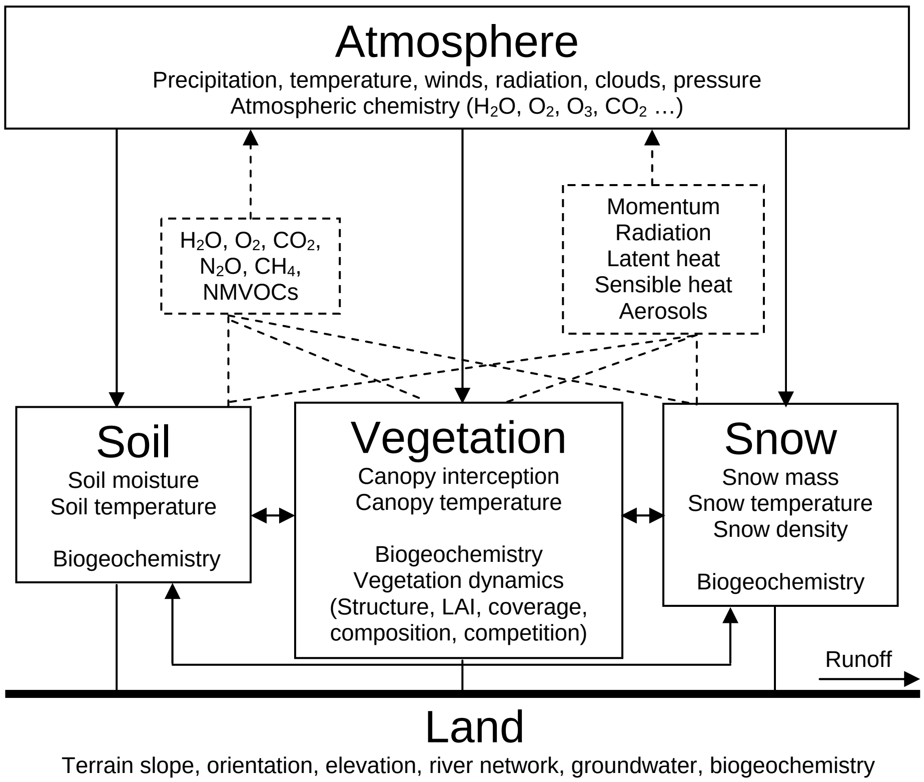

Figure 2. Schematic representation of the components and processes in a land-surface model. The framework may be used for studies of weather, climate, air quality, and water quality. The arrows indicate interactions between different components (solid boxes). NMVOCs are nonmethane volatile organic compounds (e.g., isoprene (C_5H_8) monoterpenes ($C_{10}H_{16}$)). More components may be added to represent glaciers, lakes, wetland, and urban areas.

The bucket model ignores the complex processes of soil water movement (e.g., capillary and gravitational processes) and the uptake of water by roots in the presence of vegetation (Yang 1995). In addition, the bucket model fails to represent the β function in terms its dependence on the soil moisture content through aerodynamic, stomatal and soil surface resistances (Yang, 1995).

Evapotranspiration may be expressed in the electrical analogue form by using the resistance formulation, whereby

$$E = \beta\rho[(q_s(T_g) - q_r]/r_a \,, \tag{1}$$

where ρ = air density, $q_s(T_g)$ = saturation mixing ratio at ground surface temperature T_g, q_r = mixing ratio at a reference height, r_a = aerodynamic resistance, β = soil wetness. The β function in (1) is defined as

$$\beta = W/W_k \,, \tag{2}$$

where W and W_k are soil moisture content and its critical value, respectively. The above relation is the well-known linear β-function for the Manabe bucket model. The bucket model predicts no evaporation when the soil is dry and $\beta = 0$, and has a potential evaporation when soil is wet and $\beta = 1$. This, in part, explains why over desert and semi-arid regions, the bucket model gives too little evaporation and hence too high a temperature; conversely over rainforest regions, it gives too high evaporation and too low a temperature (Sato et al. 1989).

We can also make β dependent on the aerodynamic and surface resistances,

$$\beta = r_a / (r_a + r_s), \tag{3}$$

where r_a = aerodynamic resistance, and r_s = surface resistance (a function of plant stomatal resistance and soil resistance). Substituting (3) into (1), we obtain a variant of the Penman-Monteith equation (Monteith, 1981). It shows that evapotranspiration is subject to soil and plant biophysical controls. Other conditions being equal, the Penman-Monteith equation gives lower evapotranspiration than the bucket model. Vegetation is present over most land surfaces. Thus using (3), in principal, gives a more realistic estimate of E than using (2).

2.2. *Sensitivity Studies Using Early Land Surface Models*

Although earlier climate models have not included the above plant biophysical processes, they are still able to reveal qualitatively and comparatively the importance of land surface processes to climate. Meanwhile, due to this simplicity, they make cause and effect relationships easier to describe.

Charney et al. (1977) first used a GCM to study the albedo effects on the initiation and maintenance of drought in the semi-arid regions. Other GCM studies have examined the modeled sensitivities to surface roughness, soil wetness and emissivity (Pielke et al., 1998). The surface hydrological variables, temperature, heat fluxes and circulations are shown to be very sensitive to the prescribed anomalies of the surface components. The experiments reveal that the deforestation issue may be better studied using GCMs with more elaborate land surface models that treat soil and canopy effects in a more realistic way.

3. Advanced Land Surface Models

Deardorff (1978) first proposed an advanced land surface model suitable for use in GCMs. In his model, Deardorff considered one-layer vegetation in addition to a two-layer soil. The evaporation from the soil layer and the wet canopy, the interception, and the transpiration from the dry parts of the canopy were considered. This formulation opened the way for the following developments toward constructing advanced land surface models in GCMs.

Over the past two decades, active researches on land-atmosphere interactions have led to more than two dozen LSMs constructed for use in GCMs (e.g., Henderson-Sellers et al. 1993; Slater et al. 2001). There is much similarity and overlap among these models (Figure 2). Two well known land surface parameterizations are discussed below. The emphasis is placed on the structure of the models, requirements of GCMs, and sensitivity studies. The structure part is discussed in the order of vegetation, soil, and snow sub-models.

3.1. *Biosphere-Atmosphere Transfer Scheme (BATS)*

The development of the Biosphere-Atmosphere Transfer Scheme (BATS) is well documented in Dickinson et al. (1986, 1993, 1998, 2002). Originally, BATS was designed for use in the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM). But its philosophy, as well as many aspects of its physical parameterizations, have been adopted in other LSMs (e.g., Noilhan and Planton 1989; Dai et al. 2003).

BATS has three soil layers and one vegetation layer. There are eight prognostic variables: leaf temperature, surface soil/snow temperature, subsurface soil/snow temperature, surface soil water, root-zone soil moisture, total soil water, snow cover amount measured in terms of liquid water content, and canopy water store. There are 18 surface types, with 15 types for vegetation which are based on Matthews (1983) and Wilson and Henderson-Sellers (1985). The soil type data are based on Wilson and Henderson-Sellers (1985). For each vegetation type, there are about 27 derived parameters which determine the morphological, physical and physiological properties of vegetation and soil.

Vegetation in BATS is assumed to be a flat, porous and uniform layer. It may cover the whole grid square of a GCM. The foliage is assumed to have zero heat capacity, and photosynthetic and respiratory energy transformations are neglected. The vegetation temperature is obtained by solving a vegetation energy balance equation,

$$R_n(T_f) = L\,E_f\,(T_f) + H_f\,(T_f), \tag{4}$$

where R_n, E_f, H_f are net radiation, canopy evapotranspiration, and sensible heat flux, respectively. T_f is the leaf temperature, and L is the latent heat of vaporization.

The rate of change of water store per unit land-surface area is calculated as

$$dW_{dew} \,/\, dt = \sigma_f\,P - E_f + E_{tr,} \tag{5}$$

where W_{dew} = water store on canopy surface, σ_f = fractional vegetation cover, P = precipitation rate, E_f = evapotranspiration rate, E_{tr} = transpiration rate. An accurate simulation of E_{tr} largely depends on the way stomatal functioning is described. The stomatal resistance takes a form of

$$r_s = r_{smin} f_R\, f_S\, f_{VPD}\, f_M, \tag{6}$$

where r_s = stomatal resistance, r_{smin} = minimum stomatal resistance, f_R = a factor dependent on solar radiation (visible), f_S = a seasonality factor dependent on leaf temperature, f_{VPD} = a factor dependent on VPD, f_M = a factor dependent on soil moisture potential and distribution of roots. In (6), r_{smin} is defined to be average value for the whole canopy in earlier versions of

BATS, whereas it is the minimum stomatal resistance at the top of the canopy in Dickinson et al. (1998).

When calculating f_R, a four-level canopy is simply considered to take account of different amounts of radiation received by leaves. Earlier version of BATS (Dickinson et al. 1993) calculated the leaf area index (LAI) as a function of temperature between prescribed maximum and minimum values, while Dickinson et al. (1998) simulates the growth and loss of the green foliage by describing leaf CO_2 assimilation in addition to leaf water use. The linkage between carbon assimilation and the reciprocal of stomatal resistance (i.e., stomatal conductance) is described by a derivative of that given by Ball et al. (1987), hereafter referred to as the Ball-Berry equation, that is

$$g_s = m(A_n/C_s)F_eP + g_0, \tag{7}$$

where g_s is stomatal conductance for water vapor transfer, g_0 is a prescribed minimum stomatal conductance, m is a slope parameter (equal to 9 for C_3 plants), A_n is the net carbon assimilation, C_s is the carbon dioxide partial pressure adjacent to the leaf, P is atmospheric pressure, and F_e is a humidity-dependency stress factor. Dickinson et al. (2002) include the effects of nitrogen cycling in (7).

A bulk canopy stomatal resistance is given by r_s/LAI, where LAI is the effective leaf area index used to account for the attenuation of radiation as light passes through the canopy and the coincident decrease in plant surface which is actively transpiring.

The soil and subsurface temperatures are obtained using the force-restore method (Dickinson 1988). Soil moisture contents are computed for three overlapping soil layers: the upper layer, the root zone and the total active layer. All share the same incident precipitation, drip from foliage, snowmelt, surface evaporation and surface runoff, while they have different transpiration rates because of different distribution of roots. Fluxes between soil layers are parameterized.

Snow and soil share the same thermal balance equations while the effects of snow on modifying thermal properties, albedo and surface fluxes are described by a simple roughness and snow masking relation. The effects of refreezing of melt water are ignored.

The requirement of BATS for inclusion in GCMs is that the host GCM contains diurnal variation of insolation. The input atmospheric variables are incident shortwave and longwave radiation at the surface, precipitation rate, temperature, water vapour mixing ratio and wind velocity at the lowest model level.

There are numerous sensitivity studies using BATS carried out in standalone (off-line) mode and coupled mode. The off-line studies include (1) the model's overall performance against the observations for different land types and under different forcing conditions (e.g., Unland et al. 1996; Yang et al. 1997, 1999; Sen et al. 2000), (2) sensitivity to the inevitable errors or uncertainties of input parameters for a given biome (Gupta et al. 1999), (3) sensitivity to the initialization of its prognostic variables (Yang et al. 1995), and (4) sensitivity to the aggregation methods employed to convert from finer resolution to the coarser grids of GCMs (Shuttleworth et al. 1997; Burke et al. 2000). BATS has been used in global climate models to explore the regional climatic impacts of tropical deforestation (e.g., Dickinson and Henderson-Sellers 1988), and the impacts of doubled stomatal resistance on the water resources of the American Southwest (Martin et al. 1999). BATS also is used in high-resolution regional climate models (e.g., Giorgi et al. 1999).

3.2. *Simple Biosphere (SiB) Model*

The development of the Simple Biosphere Model (SiB) is well documented in Sellers et al. (1986, 1996a,b). The philosophy of the model design is to model the vegetation itself and let the vegetation determine the ways in which the land surface interacts with the atmosphere. Specifically, when vegetation is present, it plays an important role in radiation absorption (i.e., high absorptivity in the visible wavelength interval and moderate reflectivity in the near-infrared region), biophysical control of evapotranspiration (e.g., through stomatal resistance), momentum transfer (e.g., by roughness length), soil moisture availability (i.e., by the depth and density of the vegetation roots), and insulation (i.e., by vegetation shelter).

The original SiB model has 3 soil layers and 2 vegetation layers. There are 8 prognostic physical-state variables, 3 temperatures (one for the canopy vegetation, one for both the ground cover and the soil surface, and one for the deep soil layer; 2 interception water stores (one for the canopy, and one for the ground cover); and three soil moisture stores. The revised SiB model has changed the original two-layer vegetation canopy structure to a single layer and incorporated a patchy snowmelt treatment (Sellers et al. 1996a).

In SiB, the global vegetation is classified into 12 ecotypes. They are originally based on Kuchler (1983) that recognizes 32 natural vegetation communities, and the land use data base of Matthews (1983). The 12 ecotypes are stored at 1°×1°. For each vegetation type, there are about 54 parameters for both canopy and ground cover, which determine the morphological, physical and physiological properties of the biome. All these parameters, together with prognostic variables of SiB and the atmospheric boundary conditions, are used to determine the fluxes between the surface and the atmosphere. The vegetation phenology is described by use of satellite data (Sellers et al. 1996b).

In the original SiB model (Sellers et al. 1986), there are two vegetation layers representing two morphological groups. The top layer consists of trees or shrubs while the ground layer is for grasses and other herbaceous plants. Either or both or neither may exist in one grid square of model land. Unlike BATS which neglects canopy heat storage, SiB assumes it is a function of leaf area index and intercepted water on the canopy.

In the revised SiB model (Sellers et al. 1996a), there is only one canopy layer. Some of the original vegetation classes are combined to reduce the number of distinct vegetation classes from 12 to 9. A canopy photosynthesis submodel is incorporated. This submodel makes explicit calculation of the photosynthetic CO_2 flux between the atmosphere and the land surface. The leaf photosynthesis-conductance model used in the new model is similar to that used in Dickinson et al. (1998), with, however, somewhat different implementation. SiB includes description of C_4 photosynthesis in addition to C_3 photosynthesis. The photosynthetic rate of the canopy as a whole is estimated from that of the uppermost leaves by multiplying by a factor that allows for the absorption of photosynthetically active radiation through the canopy. The canopy conductance then is estimated using the Ball-Berry equation (7), with the humidity stress factor set equal to relative humidity. Canopy transpiration thus is related directly to the whole-canopy carbon assimilation via the canopy conductance, but transpiration itself may feed back on the canopy conductance by influencing the canopy environment. The net CO_2 flux is assumed to be the difference between the soil respiration and the net carbon assimilation rate.

The soil model has three soil layers: an upper thin soil layer, a root zone and an underlying recharge layer. The topsoil layer is chosen to be thin to ensure there can be a significant rate of withdrawal of water by direct evaporation into the air when the pores of the soil are at or near saturation. The root zone layer contains all the roots for the two vegetation layers. The recharge layer is included to account for water transfer by gravitational drainage and hydraulic diffusion.

Although there are three governing equations for the three moisture stores in the three soil layers, there is only one equation for the soil temperature: deep soil temperature. This equation, together with the equation for the ground surface temperature, is formulated in the framework of the force-restore method (Deardorff 1977).

In SiB, the snow model is very simple compared to the sophisticated vegetation and soil models. The snow depth is explicitly predicted though it is very crude. There is no explicit treatment of snow temperature; rather it is included in ground surface temperature for a combination of three surface types: ground vegetation, soil and snow. Simple modifications of ground albedo, roughness length, the heat capacity of the canopy or ground, latent heat flux, and runoff are proposed to include snow effects (Sellers et al. 1986, 1996a).

SiB requires a GCM with diurnal variation of insolation. The atmospheric forcings include the incident radiative flux, the precipitation, air temperature, vapor pressure and wind components at the lowest model level. The incident radiative flux must be partitioned into five components: visible or PAR (< 0.72 μm) direct beam radiation, visible or PAR (< 0.72 μm) diffuse radiation, near infrared (0.72–4.0 μm) direct beam radiation, near infrared (0.72 – 4.0 μm) diffuse radiation, thermal infrared (8.0–12.0 μm) diffuse radiation, The precipitation includes large scale and convective parts.

Numerous studies have been conducted using SiB in off-line mode and coupled mode. The off-line studies include (1) the model's overall performance against the observations for different land types and under different forcing conditions (e.g., Sellers and Dorman 1987; Sellers et al. 1989; Sen et al. 2000), and (2) production of global biophysical land-surface dataset (Dorman and Sellers 1989; Los et al. 2000). SiB has been linked with GCMs to study the implementation strategy (Sato et al. 1989; Randall et al. 1996), explore the impacts of tropical deforestation on regional climate (Nobre et al. 1991), compare radiative and physiological effects of doubled atmospheric CO_2 on climate (Bounoua et al. 1999), simulate terrestrial surface CO_2 concentrations and fluxes (Denning et al. 1996), assess δO^{18} in atmospheric CO_2 (Ciais et al. 1997), assess sensitivity of climate to changes in vegetation growth as measured by NDVI (Bounoua et al. 2000), and examine the effects of vegetation on the diurnal temperature range (Collatz et al. 2000).

Xue et al. (1991) described a simplified version of SiB model (SSiB) by reducing the number of input parameters to 26 from the original 54. Among those 28 eliminated parameters, 17 are due to change of model structure (i.e., reducing the vegetation to one layer from the original two storeys) and the remaining 11 are from the simplification of parameterizations (e.g., in radiation fluxes, stomatal resistance and aerodynamic resistances). The simplified version is shown to reproduce the original results quite closely but with the computational cost reduced by about 55%.

4. Major Issues of Land-Surface Modeling

From a brief review of BATS and SiB as two examples of current LSMs, several general conclusions can be drawn. *First*, these models by design may be regarded as one-dimensional models, i.e., only layers in the vertical (z) direction are considered. While they are to be used ultimately in 3-dimensional atmospheric models, they ignore the horizontal interactions of land surface processes between adjacent grid squares. Processes that occur in the soil-vegetation-snow-atmosphere system of one grid square are not affected by what happen in systems of the neighboring grid squares. For example, the runoff is not modeled to take this into account. *Secondly*, vegetation is treated as a "big leaf", i.e., (linearly) scaling from a size of a normal leaf up to a grid square of 10 km × 10 km to 100 km × 100 km so that a single stoma is associated with the big leaf. *Thirdly*, only three land components (soil, snow and vegetation) are treated explicitly whereas land ice and lakes are neglected. *Fourthly*, while carbon fluxes and plant growth are included, the number and areal coverage of vegetation types within a grid are prescribed.

While some progress has been made to address some of the above issues, e.g. topographic-based runoff (Famiglietti and Wood 1994), vegetation dynamics (Foley et al. 1996), and effects of lakes in LSMs (Bonan 1995), it is likely that they will remain active research topics in next five to ten years. Meanwhile, further research is required to resolve uncertainties in several aspects discussed below.

4.1. *Datasets*

The dataset of vegetation and soil is an indispensable and important part of LSMs. However, the chosen number of vegetation and soil types, the assignment of the dominant type for each grid cell, and the derived secondary parameters for each type are still in "trial-and-error" stage. The sensitivity of LSMs to the chosen procedure has not yet been adequately explored. In particular, the sensitivity of the schemes to the aggregation procedure used to transform the data from the original high resolution (say, 1km×1km) to the required host model resolution has just begun to receive attention (e.g., Shuttleworth et al. 1997; Bonan et al. 2002b).

A wide variety of approaches has been used to specify vegetation and soil data. Matthews' (1983) vegetation dataset has been previously used in generating the data for BATS and SiB. However, the application of this dataset varies from scheme to scheme. For example, BATS also utilizes the Wilson and Henderson-Sellers vegetation dataset. The SiB model mainly uses the Kuchler (1983) vegetation dataset. Nowadays, LSMs have used satellite-derived land cover classifications and vegetation seasonality (Los et al. 2000; Bonan et al. 2002a,b).

The soil datasets are different in all the schemes although FAO/UNESCO (1974) is the main data source for most of the schemes. These digital archives of global vegetation and soil types are generally obtained by digitizing published global maps of vegetation and soil (Matthews 1983; Wilson and Henderson-Sellers 1985). Their approach was to choose one or a few main maps as main references for the globe, while the other maps were used to complement these for particular regions. Therefore, the data quality is heterogeneous across the globe. In addition, since their sources are generally based on past publications, their data cannot reflect recent rapid changes of land surface features (e.g., tropical deforestation).

As for the secondary parameters (e.g., roughness length, zero-plane displace height, leaf area index, canopy height, minimum stomatal resistance), most of them can be estimated by numerical and field experiments (Sellers and Dorman 1987; Noilhan and Planton 1989; Sellers et al. 1989; Sen et al. 2001), while some of them can be inferred from satellite observations (Sellers et al. 1996b; Zeng et al. 2000; Schaudt and Dickinson 2000). However, many have to be inferred by "intelligent guessing" as guided by the literature (Dickinson et al. 1993; 2002). Dorman and Sellers (1989) undertook an interesting study to provide a mutually consistent global climatology of surface albedo, surface roughness and the minimum stomatal resistance on a 1°×1° grid by running their SiB submodels with prescribed PAR and wind velocity. This global dataset has been updated recently by Los et al. (2000). Their dataset may be used in GCMs which do not have biophysically based LSMs.

Satellite imaging has been used to obtain integrated and consistent information about land-surface conditions. There are an increasing number of studies that have used satellite-derived vegetation types and/or seasonality in the coupled LSMs and GCMs (Bounoua et al. 2000; Buermann et al. 2001; Bonan et al. 2002b). Wei et al. (2001) used AVHRR-derived surface albedos to evaluate the coupled LSMs and GCM simulations. Tsvetsinskaya et al. (2002) related MODIS-derived surface albedo to soils and rock types over arid regions.

The past twenty years have also seen a number of important field experiments, e.g., ARME (Shuttleworth et al. 1984), BOREAS (Sellers et al. 1997), FIFE (Sellers et al. 1992), and HAPEX-MOBILHY (Andre et al. 1996) (see Appendix A for the meanings of these acronyms). These field observations need to be integrated with remotely sensed data to produce a new generation of global land surface parameters.

4.2. *LSM and GCM Compatibility*

When developing a land surface model for a GCM, one has to make certain compromises between its realism and its suitability for use in a climate model. Scientists in plant and soil sciences have developed their own complex framework which may be comparable to the climate models in terms of complexity. However, their framework must be somewhat simplified before being used in climate models. On the other hand, land surface processes have to be adequately treated to capture their importance to climate.

As discussed earlier, the minimum resolution requirements of LSMs are at least two soil layers and one canopy layer. Physically, the two soil layers are necessary since the top thin layer responds faster with the diurnal cycle while the deeper layer controls the seasonal changes (Deardorff 1977, 1978). More layers of vegetation are desirable for realistic calculations of radiation transfer and momentum exchange, but a larger number of parameters must be prescribed and calibrated (Sellers et al. 1989; Gupta et al. 1999). Most LSMs have only one vegetation layer.

Owing to the diurnal variations of incident radiation flux and its partition of sensible and latent fluxes, LSMs require GCMs to include a diurnal cycle. Most LSMs need the incident shortwave radiation in two parts split at 0.7μm. Compared to the other LSMs, SiB requires a more detailed description of the incident radiation flux from GCMs. The five components of the fluxes must be provided for use in the radiative transfer calculations in the canopy. In LSMs, the radiation transfer is generally calculated every time step while in GCMs radiation is computed every 3 hours for shortwave radiation and every hour for longwave radiation to save CPU time.

Generally, GCMs provide the atmospheric forcings for the LSMs such as incident radiation at the surface, precipitation, temperature, water vapour mixing ratio and the wind components. The modeled variables such as surface solar radiation, mixing ratio and precipitation may be questionable at present because of possibly inadequate treatments in cloud, aerosol properties and convection. As the rainfall intensity strongly affects the rainfall interception and re-evaporation by the canopy, the nature of the precipitation (frontal or convective) and its sub-scale variability determine the mutual requirements of GCMs and LSMs (Pitman et al. 1990).

The height of the lowest GCM model level is an important factor for LSMs. In calculating the turbulent transfers of momentum, sensible and latent fluxes between the surface and the lowest model level, the fluxes are assumed to be independent on the height above the surface. This approximation can hold only over flat and homogeneous surfaces (below 100m). Most GCMs to date treat the PBL in a very simple fashion. Only one or two model layers are located within the PBL, and the PBL processes are heavily parameterized with large-scale prognostic variables. On the other hand, most mesoscale models pay greater attention to computation of PBL processes. A lowest model level < 100 m seems necessary, and improved resolution PBL models may be needed in order for the GCM to be fully compatible with some LSMs.

4.3. *Surface Temperature*

The thermal balance at the land surface is important to the near ground climate. Two schemes are commonly used: the slab model and the force-restore method. A slab model assumes a thermal uniform layer with fixed thermal properties and heat conduction equations are solved for the temperature using a finite difference method or finite element method. Its accuracy is increased when the number of layers is increased. Due to the computational cost, the number of layers cannot be very high. The Community Land Model (CLM: Dai et al. 2003, Bonan et al. 2002a) has ten soil layers, while many LSMs have three layers.

The force-restore method is formulated from an analytical solution of the soil heat conduction equation under assumptions of periodic forcings and homogeneous medium

(Deardorff 1977, 1978; Dickinson 1988). One of the two prognostic variables interacts (rapidly) with the forcing term, and the other responds (slowly) with the storage term. This efficient scheme has been used in many LSMs including BATS and SiB.

The above two assumptions in the force-restore method are not always valid. The periodic solar forcing is often disturbed by other factors (e.g., clouds and trees), and heterogeneity rather than homogeneity is common (snow, soil, soil moisture, roots, etc) within a soil layer. Therefore, a number of approaches have been adopted to modify the original force-restore method. Dickinson (1988) takes account of the contribution from snow and soil moisture to the heterogeneity, and proposes a generalized force-restore method with a seasonal cycle included, which has been added in an updated version of BATS. CLM solves the heat diffusion equation explicitly for soil temperatures.

For canopy temperature calculations, the degree of complexity of schemes used is much lower than that adopted for calculating canopy albedo. For example, in SiB, a reasonably realistic two-stream approximation has been used to calculated albedo and five components of solar flux are considered. In contrast, a single temperature is assigned to the whole canopy with a single prescribed canopy heat capacity. This limitation will influence the calculation of stomatal resistance since it is a function of vapor pressure deficit that is in turn dependent on leaf temperature. The ground has one temperature for ground cover and bare soil, and each soil layer is assumed to be isothermal with a single temperature.

Similar approaches are also used in BATS for calculating vegetation temperature except that the heat capacity is assumed to be zero. The scheme used in Abramopoulos et al. (1988) in calculating canopy temperature is slightly more elaborate. It differentiates canopy temperature for wet and dry surfaces. It calculates surface temperatures and soil temperatures for shielded and bare portions of land, separately. Wang and Leuning (1998) developed an efficient one-layered, two-leaf canopy model which calculates the fluxes of sensible heat, latent heat and CO_2 fluxes separately for sunlit and shaded leaves.

4.4. *Soil Moisture and Canopy Interception*

Prognostic variables for describing soil water movements vary. For example, in BATS, they are soil moisture content measured in terms of water equivalent depth d_w, in meters. In Abramopoulos et al. (1988), they are mass of liquid water per unit lateral area in the soil layer, which is a product of water density and the depth, i.e., $d_w\ \rho_w$, (kg m^{-2}). In ISBA (Noilhan and Planton 1989) and CLASS (Verseghy 1991), they are measured with actual volumetric soil moisture X, (m^{3}m^{-3}). In SiB, they are soil moisture wetness W which is a ratio of actual volumetric soil moisture (X) in a layer to its value at saturation (X_s) or porosity. W is dimensionless. In the UKMO LSM (Warrilow et al. 1986), they are products of water density and actual volumetric soil moisture, and are also called soil moisture concentration, in units of kg m^{-3}. All these are essentially the same and related through

$$W = X/X_s = (X\rho_w)/(X_s\rho_w) = (d_w/D)/X_s = d_w/D_w = (d_w\rho_w)/(D_w\rho_w), \qquad (8)$$

where D is the depth of soil layer (m) and D_w is the soil moisture capacity (m).

Under a series of assumptions, e.g., a spatially homogeneous soil layer with no horizontal water movement and no melting or freezing within it, vertical movement of water soil will follow Darcy's law (noting that the actual form of the equation is dependent upon the direction of z and that of the water flux). This flux-gradient has been used in all the LSMs discussed in previous sections. Therefore, all of them treat more or less the same processes when calculating soil water. They include, for instance, the forcing (throughfall, canopy drip and snowmelt), surface soil evaporation, surface runoff, capillary and gravitational drainage, and transpiration by the canopy. Only in Abramopoulos et al. (1988), are the surface slope and the interstream distances taken into account in order to calculate the underground runoff and soil moisture content. Some LSMs have included explicit treatment of the frozen soil water budget (Verseghy 1991; Dai et al. 2003).

LSMs calculate the canopy vegetation water store with similar governing equations. BATS uses the same method proposed by Deardorff (1978) to calculate the wet fraction of a canopy, i.e., the 2/3 power law,

$$L_w = (W_{dew}/W_{max})^{2/3}, \tag{9}$$

where L_w = fractional area of leaves covered by water, W_{dew} = water store on the surface of canopy, W_{max} = maximum water the canopy can hold .

In SiB, it is prescribed as

$$L_w = W_{dew} / W_{max} \tag{10}$$

under the condition of the saturation vapor pressure at canopy temperature which is less than the vapor pressure in canopy air space. L_w = 1 otherwise. In earlier version of BATS, W_{max} = $0.2\sigma_f$LAI (mm); in SiB (Sellers et al. 1989), $W_{max} = 0.1\sigma_f$LAI (mm)

4.5. *Evapotranspiration*

In general, surface evapotranspiration may be categorized into three components: soil evaporation, canopy evaporation and transpiration. They all look similar at first sight but are in fact quite different. They can, however, all be written as a product of potential evaporation and a coefficient in the form of (1). The fact that they are dissimilar is mostly because of the coefficient and partly due to the potential evaporation. In the case of soil, the coefficient is controlled by the soil wetness and the soil properties (Abramopoulos et al. 1988). In the case of canopy evaporation, the coefficient is determined by the amount of liquid water on surfaces of the canopy and on the plant morphology, e.g., vegetation cover and leaf area index (Rutter and Morton 1977; Sellers et al. 1986). In the case of canopy transpiration, the coefficient depends upon plant physiology, morphology and the environmental conditions (Sellers et al. 1986). The potential evaporation from the canopy surface is different from that from a soil surface partly because the aerodynamic resistances are different and partly because the temperatures at the soil surface and canopy surfaces differ. In the following we will discuss how they are parameterized in LSMs.

Soil evaporation

Soil evaporation is the water flux from the soil surface to its overlying atmosphere. It is limited by diffusion. Therefore in BATS, evaporation from the soil surface is given by the minimum of potential evaporation and maximum moisture flux through the wet surface that the soil can sustain. This diffusion-limiting maximum moisture flux is parameterized from a multi-layer soil model. However, in SiB, a different approach is used. A soil surface resistance is calculated explicitly and then added to the aerodynamic resistance. The resistance formulation of bare soil evaporation takes the form similar to (3).

Canopy evaporation

Canopy evaporation or wet-canopy evaporation is sometimes called interception loss since it is that portion of the rainfall that is held on canopy surfaces as liquid water and which then evaporates to the atmosphere without reaching the soil moisture store (Rutter and Morton 1977). Researches on interception loss from forests can be represented by two directions: one is guided by Rutter and his co-workers who used a complex numerical model; the second is represented by Gash (1979) who used an analytical model, although conceptually similar to the Rutter model.

In both groups of models, the evaporation from a saturated canopy during rainfall is estimated from the Penman-Monteith equation, i.e., a product of a wetting function and the potential evapotranspiration (Deardorff 1978). The wetting function has been discussed before, and the potential evapotranspiration is the evaporation which would occur from a totally wet canopy assuming the water flux comes freely from the water at the surface of canopy (i.e., canopy surface resistance is zero). Both groups of models were tested successfully against measurements (Rutter and Morton 1977).

Many LSMs simplified Rutter's approach to calculate the evaporation from the canopy surface water store. For example, these LSMs use one prognostic variable only for the whole canopy surfaces (including leaves and stems/trunks). In the resistance formulation of evaporation, the only resistance term is aerodynamic resistance. Therefore, it is essentially equal to potential evaporation rate as for a wet soil surface, except where the resistance is replaced by a bulk boundary layer resistance for the canopy leaves, which is different from the common approach of micrometeorologists and forest scientists who have typically an aerodynamic resistance above canopy under neutral conditions (Shuttleworth et al. 1984).

Canopy transpiration

Canopy transpiration, sometimes called dry-canopy evaporation, is a physiological process associated with water transfer from the soil through roots, stems, branches and leaves (Sellers 1992). Most LSMs use the same formulation of transpiration, i.e., Penman-Monteith's combination equation (Monteith 1981), but more closely related to a soil-plant-atmosphere model for transpiration proposed by Federer (1979).

In general, the resistance term is a sum of the bulk stomatal resistance for the canopy and the bulk boundary layer resistance for the canopy leaves, although in Federer's original model, the aerodynamic resistance was approximated by an equation for a thermally neutral atmosphere above the canopy, which is similar to that used by forest scientists for calculation of interception loss. The transpiration is then limited by the supply of water from the roots and atmospheric conditions of demand. For example, the canopy cannot transpire when there is dew forming onto its surface, nor can it when the soil moisture potential drops below the wilting point. However, the parameterizations of stomatal resistance, the dry fraction of transpiring canopy and roots-limiting factor are different in the various LSMs. In Federer's approach, the roots-limiting factor was formulated according to Cowan (1965), but this is somewhat simplified in all the LSMs. Recently, a database of vertical root profiles was developed from the literature with 475 profiles from 209 geographic locations (Schenk and Jackson, 2002), and it was incorporated in CLM (Dai et al., 2003).

4.6. *Stomatal Resistance*

Stomatal resistance is a biophysical subgrid scale process which is difficult to parameterize while retaining sufficient generality for use in climate models. Although numerous factors determine leaf stomatal resistance, there are only five environmental conditions: solar radiation, temperature, vapour pressure deficit, leaf water potential (soil water potential), and ambient carbon dioxide, which are commonly parameterized. Various schemes exist in the literature (e.g., Jarvis 1976; Farquhar and Sharkey 1982; Sellers et al. 1996a; Dickinson et al. 1998, 2002). In the LSMs, a simple strategy is used to scale stomatal resistance from a leaf to a grid. After the stomatal resistance (r_s) of a leaf is calculated, the bulk stomatal resistance (r_c) is derived by assuming all the leaves of the canopy are parallel and then applying Ohm's law. A quantity over a grid square is thus obtained by multiplying vegetation cover fraction (σ_f) to scale up from a canopy to a grid square. In general, there are some common uncertainties in the above formulations for bulk stomatal resistance. For example, the coefficients and parameters that determine the stress factors are not readily available. They must be determined from complex physiological experiments and more advanced theories need to be established.

4.7. *Canopy Drip*

In all LSMs, the calculation of retained water on the canopy surface (including leaves and stems/trunks) is analogous to the bucket model for soil moisture content though a more general drip formulation is developed by Massman (1980). A universal water holding capacity on the canopy surface is equivalent to the soil field capacity. The water level falls because of canopy evaporation as discussed above and rises because of intercepted rainfall or dew formed onto the surface. The excess amount of water beyond the maximum water storage is called canopy drip, which is equivalent to the soil surface runoff.

In most LSMs, the intercepted rainfall is simply a proportion of incident precipitation above the canopy according to the calculated vegetation cover fraction. In SiB, this is calculated in an analogous way to the exponential attenuation of radiation through the canopy when the flux is vertical and the leaves are black (Sellers et al. 1986), and the drip is formulated considering the sub-grid scale precipitation (Sato et al. 1989). A similar approach is used in the updated version of the UKMO LSM (Warrilow et al. 1986). This concept is very important because the drip, unlike the surface runoff, is still a contributory factor in the hydrological cycle, i.e., a water input to the soil water budget. Therefore, its magnitude will affect the infiltration and evaporation at the surface. Pitman et al. (1990) has demonstrated its importance by prescribing this process in a version of BATS following Warrilow et al. (1986) and Shuttleworth (1988).

4.8. *Runoff*

Runoff, as a part of the hydrological cycle, is important in hydrology and climatology though it is viewed differently in both sciences. It is often perceived in hydrology as a direct response to precipitation (i.e., overland flows in the form of sheets, rivulets, streams and rivers, and/or near-surface flows guided by underlying impermeable layers or subground air channels) with evapotranspiration being a residual; conversely, it is often treated in climatology as the residual after evapotranspiration requirements have been satisfied (Dickinson 1992). In the latter case, the determination of surface evapotranspiration is based on well-established theories by micro-meteorologists and plant scientists.

Hydrologists and climatologists also have different interests in the scales of runoff. Hydrologists are primarily concerned with the local or catchment scale runoff, while climatologists consider runoff at a much broader scale (Shuttleworth 1988; Dickinson 1992). Recently, both views have begun to converge and there is an emergence of the so-called macrohydrology or global-scale hydrology (Shuttleworth 1988).

Warrilow et al. (1986) have given a useful introduction to runoff generating mechanisms. These are summarized as follows: (1) Horton runoff, (2) Dunne runoff, (3) saturation through flow, occurring at less permeable levels in the soil which having become saturated enable moisture to move parallel to the soil horizons in a downslope direction rather than vertically and (4) deep percolation to the water table and thus to surface drainage.

Horton runoff or infiltration excess overland flow, is generated due to the excess of precipitation intensity over soil infiltration capacity at a point. It accounts for only a small fraction of the surface runoff contribution to streamflow. Dunne runoff or saturation excess overland flow is caused due to the occurrence of precipitation over saturated and impermeable surfaces. It is largely responsible for the rapid response of streams to precipitation.

The runoff process is difficult to model due to its complex nature. Nevertheless, there is still a rich coverage in the literature on modelling runoff process over a small scale (e.g., Maidment 1993). L'vovich (1979) provided global maps of annually-averaged surface and ground-water runoff. In contrast, runoff in early generation of climate models simply is overland flow. In more advanced LSMs, it also includes the gravitational drainage. In BATS, surface runoff is parameterized as

$$R_s = (\rho_w / \rho_{wsat})^4 G \text{ , if } T_g > 0°\text{C} \tag{11a}$$
$$R_s = (\rho_w / \rho_{wsat})\, G \text{ , if } T_g < 0°\text{C} \tag{11b}$$

where ρ_w is the averaged soil density toward the top layer, ρ_{wsat} is the saturated soil water density, G is the throughfall plus snowmelt and canopy drip minus soil evaporation. The total runoff is equal to the sum of the surface runoff and the subsoil drainage. In SiB, runoff is defined as precipitation excess plus gravitational outflow from the lowest soil moisture store.

Considerable progress in runoff research has been made recently that includes hillslope processes in LSMs. In particular, a topography-related parameterization of runoff (Beven and Kirkby 1979) has been incorporated in LSMs (Famiglietti and Wood 1994; Stieglitz et al. 1997; Koster et al. 2000; Yang and Niu 2003; Niu and Yang 2003). More research is needed to quantify the role of microtopography at ~2 m horizontal grid and macropores in soil moisture and runoff simulations.

Runoff is treated as a diagnostic variable in all the LSMs and generally regarded as the excess of water in the soil reservoir. This excess amount plays no further part in the model's hydrological cycle, though in CLASS, an attempt has been made to save surface runoff (overland flow) as ponded water between time steps (Verseghy 1991). Since there are no rivers or lake levels allowed explicitly in climate models, runoff cannot be used to increase lake levels or strengthen the river flows. Input of land runoff water has just begun to be used to alter the salinity of the ocean in the oceanic GCMs. In addition, those simplifications may be in accordance with omitted treatment of sub-grid scale distribution of topography and horizontal water flow processes, though surface slope and interstream distance has only been considered in the GISS LSM (Abramopoulos et al. 1988).

4.9. *Snow*

Of all the large-scale surface features, snow cover exhibits the largest spatial and temporal fluctuations, ranging from 7% to 40% in the Northern Hemisphere during the annual cycle. Associated with these fluctuations are variations in the surface albedo and radiation balance as well as water vapor input to the atmosphere through sublimation and evaporation and water input to the soil and river systems through melt. Therefore, snow cover represents an important component in the Earth's climate system (Yang et al., 1999).

The snow energy balance equations in BATS and SiB are solved following the force-restore approach, which uses a composite snow-soil energy budget and a single snow mass layer. Recent studies have shown that such approach is inadequate to simulate snow mass and snow temperature during melt and that the results are improved with a multi-layer snowpack model (e.g., Lynch-Stieglitz 1994; Jin et al 1999; Boone and Etchevers 2001; Yang and Niu, 2003).

LSMs simulate co-existence of snow-covered canopy, snow-covered bare ground, snow cover underneath the canopy, snow-free canopy, and snow-free bare ground (e.g., Verseghy 1991; Dickinson et al. 1993). The parameterization of snow cover fraction has been shown to play an important role in modulating surface radiation balance and hence snow mass balance through surface albedo (Yang et al. 1997; Slater et al. 2001; Wei et al. 2001). Further research is needed to provide an accurate parameterization of snow masking over the vegetated surface and in the mountainous regions.

4.10. *Sub-grid Scale Variability*

Sub-grid parameterization within climate models "will always remain the Achilles' heel of numerical climate simulation" no matter how fine the model's resolution is (Entekhabi and Eagleson 1989). Currently there are three main approaches to account for the sub-grid scale variability.

Component approach

In this approach, each grid-square of land is regarded to be a combination of basic components such as soil, vegetation and snow. Temperature and water content equations are established explicitly for each component. More components can be included to increase the realism. Interactions among components and from components to their common overlying surface are explicitly included. This approach has been used in BATS and SiB.

Tile approach

In a grid square which consists of a single component, there still exists heterogeneity. For example, Wetzel and Chang (1987) reported that for a grid square with a spacing of 100 km or greater and comprising a soil surface only, the expected sub-grid scale variability of soil moisture may be as large as the total amount of potentially available water in the soil.

Avissar and Pielke (1989) have discussed an approach which accounts for the non-uniform surfaces of a grid-square. In this approach, the heterogeneity of sub-grid surface is explicitly considered and the subdivision can be made so fine that each sub-cell (or patch) can be approximated as being homogeneous. It is assumed that there are no interactions among sub-cells of different properties, and only processes occurring between them and the overlying atmosphere are treated. The processes of each sub-cells are now assumed to be additive; a bulk (global or total) quantity (a flux or temperature) is obtained by summing according to the areal weights of each sub-cell.

This approach has been incorporated into the meso-scale numerical model to account for land surface heterogeneities (e.g., random distribution of topographical variations, land use diversity) for regional meteorology studies (Avissar and Pielke 1989). They found strong contrasts in sensible heat fluxes because land surface heterogeneities may cause circulation as strong as sea breezes. However, their application in climate models is limited because climate models have much coarser resolutions with which the number of subgrid areas has to be very large. The larger number is difficult to realize because of computer limitations. In addition, there are no adequate data to validate the parameterization. Nevertheless, the composite fluxes from a grid square are used in recent LSMs (e.g., Bonan et al. 2002a,b).

Statistical approach

In this approach (Giorgi and Avissar, 1997), some key surface variables such as the grid-square area-means are assumed to have subgrid variance and follow a probability density function (pdf). Precipitation, especially convective, was the first such variable to be studied seriously. The grid size in a climate model is large enough to include a number of convective storms. Consequently there must be great spatial variability of distribution of precipitation over a grid square. However, GCMs assume precipitation uniformly covers the grid square. Precipitation intensity is thus underestimated and this in turn leads to an unrealistic estimation of canopy drip, soil moisture, infiltration, runoff and evapotranspiration. Eagleson et al. (1987) conducted theoretical and observational studies on the relationship between convective storms and the spatial distribution of surface wetting for a catchment area. Warrilow et al. (1986) assumed that precipitation falls on a proportion, μ, of the grid area, and within μ, the local precipitation rate P_l is represented by a pdf of an exponential form as

$$f_p(P_l) = (\mu/P) \exp(-\mu P_l/P), \tag{12}$$

where P_l = local point precipitation rate, P = grid square area-averaged precipitation, a prognostic variable from climate models, $f_p(P_l)$ = probability distribution function, μ = fraction of grid-square receiving precipitation.

Assuming a point runoff rate is

$$R_l = \max(P_l - F, 0), \tag{13}$$

one has

$$R = P \exp(-\mu F/P), \quad F_e = P\,[1 - \exp(-\mu F/P)], \tag{14a, b}$$

where R_l , R = local point and area-averaged runoff rate, respectively. F_l, F = effective and maximum surface infiltration rates. A similar approach was adopted to simulate canopy interception of precipitation by Shuttleworth (1988).

The above schemes have been incorporated into BATS for sensitivity tests over the Amazon rain forest region by Pitman et al. (1990). They found that surface variables in terms of evaporation and runoff are very sensitive to precipitation regimes. Sato et al. (1989) have used a similar exponential pdf for precipitation and canopy interception in SiB and made GCM studies with this version of SiB. Further research is required to compare these two methods at regional scales using observational data.

Entekhabi and Eagleson (1988) further generalized Warrilow et al.'s approach, using a gamma pdf for soil moisture

$$F_s(s) = [\lambda^{\alpha}/\Gamma(\alpha)]s^{\alpha-1}\exp(-\lambda s), \quad \lambda, \alpha, s > 0, \tag{15}$$

where s = soil wetness, λ , α = parameters determining the variance of the mean $E(s)$.

For precipitation, an exponential form of pdf is used, which is a simplification of Γ-pdf when α = 1. Based on both P_l and s and the deterministic equations describing basic soil moisture physics, they then derived a number of grid-square averaged dimensionless quantities including surface runoff ratio (surface runoff to grid square mean precipitation), infiltration rate, bare soil evaporation efficiency (ratio of actual to potential evaporation) and transpiration efficiency.

In their derivations, two components (Horton and Dunne as defined before) of surface runoff are considered. The derived dimensionless quantities are useful to determine the key variables such as runoff, infiltration, bare soil evaporation and transpiration for a grid-square. For example, given a grid-square mean precipitation which is supplied from a climate model, the actual runoff of a grid square is obtained as a product of the runoff ratio and the grid-square mean precipitation. Entekhabi and Eagleson (1989) found that by having such formulations, runoff and evapotranspiration rates are very sensitive to the fraction of surface wetting and the spatial variability of the soil moisture. However, whether they are valid in LSMs or in GCMs has not yet been investigated.

5. Summary

A comprehensive review is given to the history, development, current status and future of biophysical and hydrological aspects of LSMs. It is pointed out that an improved understanding of these processes is critical for an accurate prediction of climate variability on intraseasonal to interannual timescales. The improved understanding also is essential for an effective coupling of these schemes with biogeochemical cycles and vegetation dynamics in GCMs to simulate the impacts of land-cover change and the impacts of increasing CO_2 on climate. Significant problems remain to be addressed, including the difficulties in parameterizing hillslope runoff, fractional snow cover, stomatal resistance, evapotranspiration, and sub-grid-scale variability and scaling. However, further progress is expected as the results of large-scale field experiments and satellite datasets are exploited.

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