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Land Surface Interaction

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Land and Climate Modeling

Historically, climate models have evolved out of general circulation models (GCMs) of the atmosphere, developed by meteorologists, and radiative convective models, developed by atmospheric physicists. These areas are my background. Land was thought of simply as a lower boundary condition to be put in as simply as possible. This was an implicit message in all of my academic training and was the basis for my first attempts to be involved in climate modeling. Conversely, many other disciplines have studied the role of soils and vegetation over land in great detail but traditionally with a very local focus, and with the atmosphere and radiation prescribed from observations rather than as part of the model.

When GCMs were originally developed, requirements for accuracy and realism in the physical descriptions were much lower than they have become today. I recall conversations two decades ago where a model simulation of the tropical tropopause 40 degrees warmer than reality was regarded as satisfactory. Now only a few degrees is of considerable concern, a discrepancy small enough that it could result as readily from errors in the data themselves or their interpolation or the model layer structure as from errors in the model treatment of radiation. Except for the basic fluid equations, physical treatments in the original GCMs were all generally quite simple, represented by a single equation and/or describable in one or two sentences. Land has been no exception. S. Manabe in 1969 first moved from the viewpoint of land as a boundary condition to land as an interactive part of the system with his "bucket" hydrology model; all land points had a water-holding capacity with any surplus going into runoff, and otherwise the storage was determined by the balance between precipitation or snowmelt and evapotranspiration. Surface albedos have been simply specified numerical values, often the same value for all land points, and the aerodynamic drag coefficients (determined in principle by surface roughness) have had one value for land and one for ocean.

Demands for better answers and increasing maturity of the science are driving climate models to much more detailed treatments of physical processes, much greater emphasis on validation of the assumed process physics, and in general greater complexity. These demands are shifting the requirements for training and the research emphases of individual scientists. Initial work with GCMs required a state-of-the-art knowledge of available methods for solving the model fluid equations and procedures for handling the large data sets generated by the models. Other than that, modelers needed only the ability to copy and implement the one-line descriptions of the various model physical processes. There was no urgent requirement to validate or improve upon these descriptions, or to obtain observational data for model input or validation. Thus, it was possible for a single highly skilled individual with the help of a couple of programmers to develop a GCM and carry out a research program with it. This traditional approach is still viable for programs emphasizing improved understanding of atmospheric processes. However, the task of developing models with adequate realism and validation to meet the challenge of global change requires scientists to specialize in particular aspects of a model, although still retaining at least some familiarity with the overall model structure and behavior. Indeed, we now need to develop teams or other collaborative groupings of scientists to focus on the development of critical and still poorly developed aspects of climate models. Two particular examples that come to mind are the treatment of cloud-radiative interactions and land processes within models.

Suppose you have decided to participate in a team effort to improve the treatments of land in climate system models with both the objective of relating this effort to development of this area as part of global change models and the immediate task of improving model projections of climate change. What must you know to get started? The issue is primarily the treatment of energy and water fluxes, since these have strong direct interactions with climate models. At the same time, you must develop a framework for treatment of fluxes of carbon dioxide and other trace gases from the land surface. Currently, climate modelers start with scenarios of trace gas increases and do not attempt to make them interactive with the climate model. Practically, at present the future human contributions and natural feedbacks are too poorly known for there to be any practical improvement possible in model projections with inclusion of these feedbacks, but from a viewpoint of better understanding the system, they must be explored. These notes are an attempt to provide the appropriate background material for someone who would like to start research in the treatment by climate models of surface water and energy fluxes.

Our past knowledge base consists of two distinct threads of effort: (1) model sensitivity studies that give us better insight into the role of land processes and help sort out the relative importance of various factors, and (2) consideration of the detailed processes that must be modeled to represent land in climate models. To the extent that we are guided by the requirements of modeling the overall system, these two threads must proceed in parallel. In practice, we are working with a system of infinite complexity, and the level of abstraction vs. details required must be guided by overall modeling experience. For example, a decade ago we did not know enough about the treatment of land processes in climate models to be able to argue that we had to consider soils and vegetation as separate components. While there is still not general agreement as to this claim, a good case can be made that it is true. We can also show that the distinction between a forest and a grassland has a noticeable impact on a climate model, at least in the tropics. However, we are a long way from being able to distinguish through modeling the implications of switching between a maple and an oak forest, or pine and spruce, as might be provided by ecological modeling of response to climate change. Thus current research on feedbacks of surface fluxes with vegetation is emphasizing primarily short and intermediate scales, that is, from model time steps of a few tens of minutes to the annual and interannual time scales.

Sensitivity Studies

A climate model with an atmospheric hydrologic cycle must on the average evaporate as much water from the surface as it precipitates from the atmosphere in order to conserve water and energy. However, the presence of runoff precludes such a balance over land alone. That is, the evaporative cooling that is a major determinant of summertime land temperatures depends not only on precipitation but also on how the surface apportions this input into evapotranspiration and runoff. This apportioning in turn depends on how net radiative energy is divided between latent and sensible heat fluxes. Past conventional wisdom has largely ignored potential feedbacks of land on the atmosphere, but with increasing understanding from modeling simulations of these exchanges, we can begin to appreciate what might happen. The most common approach has been to study the effect of an arbitrary, though perhaps physically motivated, change in surface energy or hydrological processes. The study of the effect of postulated large increases in surface albedo became popular following the suggestion of Charney et al. (1977) that this might represent a positive biophysical feedback. A large number of other such studies has been reviewed by Mintz (1984). Of these I mention only that of Shukla and Mintz (1982). They considered two scenarios, an earth with either perfectly wet or completely dry continents. Their July simulations showed that the latter would have continental surfaces on the order of 10°C warmer than if they were wet, and that large reductions in continental rainfall might occur. Physical changes of continental land surfaces disrupt global energy flows by relatively small amounts compared to increasing trace gases (Dickinson, 1986). Therefore, their effects on climate are likely to be most pronounced on a regional scale.

Current studies are attempting to formulate more realistic change scenarios, though perhaps still hypothetical. I now consider the three kinds of such scenarios.

The climate effects of tropical deforestation have long been a source of speculation. However, now that land surface models are beginning to plausibly address the role of vegetation in determining the surface fluxes of moisture and energy in GCM climate models, there is some hope of establishing at least qualitatively reasonable conclusions. The Amazon region contains about half of the world's tropical forests, and in the last decade humans have been rapidly removing these forests. Thus it is useful to focus the question of climate effects of tropical deforestation on this region. Water budget studies have established that about half of the precipitation in the Amazon is supplied by evapotranspiration from the forest. How might removal of the forest reduce the evapotranspiration? Would this reduction, in turn, reduce the amount of precipitation? What effects might changing surface fluxes and precipitation have on global circulation patterns by analogy with the more thoroughly studied effects of anomalies over the tropical oceans? Several studies in the last few years have established that large decreases in evapotranspiration would result from forest removal. The most recent studies are also indicating comparable decreases in precipitation, possibly even greater in amount than the reduction in evapotranspiration.

Another important role of land surface models is in the exploration of the contribution of land to year-to-year variations in precipitation and temperature anomalies. The continental-scale drought of 1988 and the current long-term drought in California are practical examples, dependent to some extent on feedbacks between atmosphere and surface. The currently most popular hypothesis is that such anomalies are initially related to anomalies in patterns of rainfall over the tropical oceans related to the El Niño phenomenon. Studies are addressing the role of land interactions in this question, but it is too early for definitive results.

The third scenario currently being studied in which land interactions have an important role is that of global warming in response to increasing greenhouse gases. Model simulations that have produced midcontinental drying during the summer have found that the consequent feedbacks (lack of evapotranspiration) further amplified the warming. Interactive clouds have been found to have a yet further amplification effect: that is, the surface is further warmed by more solar radiation reaching the surface because of fewer clouds.

The Process of a Land Model

The atmospheric components of a GCM (Figure 1) provide the surface with fluxes of solar and thermal infrared radiation and precipitation in the form of rain or snow, and with near-surface values of wind vector, air temperature, and humidity. Water conservation is imposed by transferring the water applied to the surface either into storage by soil reservoirs or into loss by evapotranspiration or by runoff. The total radiation absorbed by the surface is balanced by emission of thermal infrared radiation, by the latent heat loss associated with the evapotranspiration or by fluxes of sensible heat, and by diffusion of thermal energy into the soil.

The original Manabe bucket model for the above processes was supposed to evaporate at the same rate as a wet surface (zero canopy resistance) during well-watered conditions, and to hold a maximum water level of typically 0.15 m, which corresponds to the available soil water, that is, the water in the rooting zone at some average field capacity minus that still present at some average wilting point. However, since it did not include the process of diffusion of water in soils or canopy resistance, its evaporation rates were unrealistic both for bare soil (after a very brief period at the rate of atmospheric demand) and for vegetated areas. Evaporation from most bare soils, in reality, is greatly reduced after the loss from the surface layer of about 1 cm of water.

Vegetation acts as a completely wet surface only during and immediately following precipitation when its foliage is wet. Otherwise, it has two important controls: (1) It can extract soil water from a greater depth than would evaporation from bare ground, and (2) it

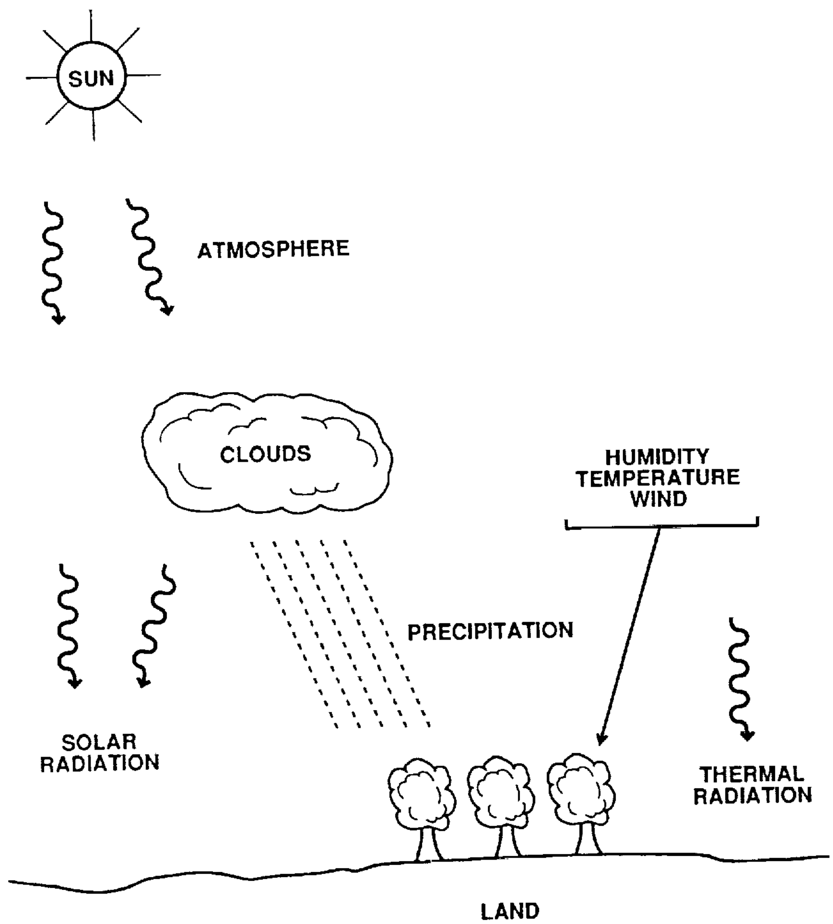

Figure 1. Sketch of the inputs needed at the surface from the atmospheric model (Dickinson et al., 1991).

retards the rate of evapotranspiration from the potential rate through resistance by the stomates to molecular diffusion of water.

The initial bucket model formulation, neglecting any reference to canopy resistance, is still widely used in GCMs for studies of climate change, and in particular has been used to address the question of effects of increasing carbon dioxide and other trace gases (e.g., Manabe and Stouffer, 1980; Washington and Meehl, 1984). The standard Goddard Institute for Space Studies (GISS) GCM has a more detailed but otherwise similar bucket model with an additional layer and water storage capacities depending on ecosystem type, and includes a rough approximation of the vegetative removal of deeper soil water by allowing infinite upward diffusion during the growing season in vegetated regions (Hansen et al., 1983).

Inclusion of geographic distributions of soil and vegetation properties in a model as shown in Table 1 allows increased realism. In particular the aerodynamic resistance of a shortgrass vegetation is over an order of magnitude less than that of forest vegetation, and this difference can have major effects on the nature of evapotranspiration in model simulations. Smooth surfaces have higher temperatures for the same atmospheric conditions because larger temperature and moisture differentials are needed to drive the fluxes required by the energy balance (Figure 2). These warmer temperatures reduce net radiation (by increasing longwave emission) by up to several tens of W/m^2. A comparable additional reduction in net radiation is implied by the higher albedo that usually applies to shorter vegetation.

For fixed net radiation and atmospheric conditions, the differences in evapotranspiration between tall and short vegetation can be obtained from the Penman-Monteith equation. For dry conditions, the relative effect of changing surface roughness depends on

Table 1a: Vegetation/land cover parameters used for South America

Parameter	**Land Cover/Vegetation Type***								
	1	**2**	**5**	**6**	**7**	**11**	**17**	**18**	**19**
Maximum fractional vegetation cover	0.85	0.80	0.80	0.90	0.80	0.10	0.80	0.80	0.80
Difference between maximum fractional vegetation cover and cover at temperature of 269 K	0.6	0.1	0.3	0.5	0.3	0.1	0.3	0.2	0.3
Roughness length (m) of vegetation	0.06	0.02	0.8	2.0	0.1	0.1	0.1	0.8	0.05
Depth of the total soil layer (m)	1.0	1.0	2.0	1.5	1.0	1.0	1.0	2.0	1.0
Depth of upper soil layer (m)	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1
Rooting ratio (upper to total soil layers)	3	8	10	12	8	8	5	10	10
Vegetation albedo for wavelengths <0.7 μm	0.10	0.10	0.08	0.04	0.08	0.17	0.08	0.06	0.08
Vegetation albedo for wavelengths >0.7 μm	0.30	0.30	0.28	0.20	0.30	0.34	0.28	0.24	0.30
Minimum stomatal resistance (s/m)	150	250	250	250	250	250	250	250	250
Maximum LAI	6	2	6	6	6	6	6	6	6
Minimum LAI	0.5	0.5	1.0	5.0	0.5	0.5	1.0	3.0	0.5
Stem (and dead matter) area index	0.5	4.0	2.0	2.0	2.0	2.0	2.0	2.0	2.0
Inverse square root of leaf dimension (m$^{-1/2}$)	10	5	5	5	5	5	5	5	5
Light sensitivity factor (W/m^2)	0.01	0.01	0.03	0.03	0.01	0.01	0.01	0.03	0.01

*See definitions in Table 1b.
From Dickinson and Henderson-Sellers, 1988.

Table 1b: Vegetation/land cover types used in the CCM

1. Crop
2. Short grass
3. Evergreen needleleaf tree
4. Deciduous needleleaf tree
5. Deciduous broadleaf tree
6. Evergreen broadleaf tree
7. Tall grass
8. Desert
9. Tundra
10. Irrigated crop
11. Semi-desert
12. Ice-cap/glacier
13. Bog or marsh
14. Inland water
15. Ocean
16. Evergreen shrub
17. Deciduous shrub
18. Mixed woodland
19. Impoverished scrub-grassland*

*Land type 19 was introduced especially for a deforestation study and is not part of the CCM land data set.

the magnitudes of the net radiation flux and the vapor pressure deficit. Under wet conditions the greater surface roughness of forests tends to enhance the evaporation loss resulting from interception, but how much depends on the spatial distribution of rainfall and on the interception model in use. In the limit of sparse or absent vegetation, evaporation will be determined by the treatment of diffusion of water through the soil.

To help summarize the common content of two canopy models developed for application in GCMs (i.e., the biosphere-atmosphere transfer scheme—BATS—Dickinson et al., 1986, and the simple biosphere model—SiB—Sellers et al., 1986), I consider the "lowest common denominator" that they both contain, then describe why various models introduce further complexity in various features. The following derivations (Dickinson et al., 1991) capture the essence of all the model treatments under full canopy conditions while leaving out many details such as treatment of fluxes into the ground and through leaf boundary layers.

In meteorological models, the upward flux F_X of a quantity X is generally represented with the aerodynamic expression

$$F_X = \rho_a C_D u (X_s - X_a) \qquad (1)$$

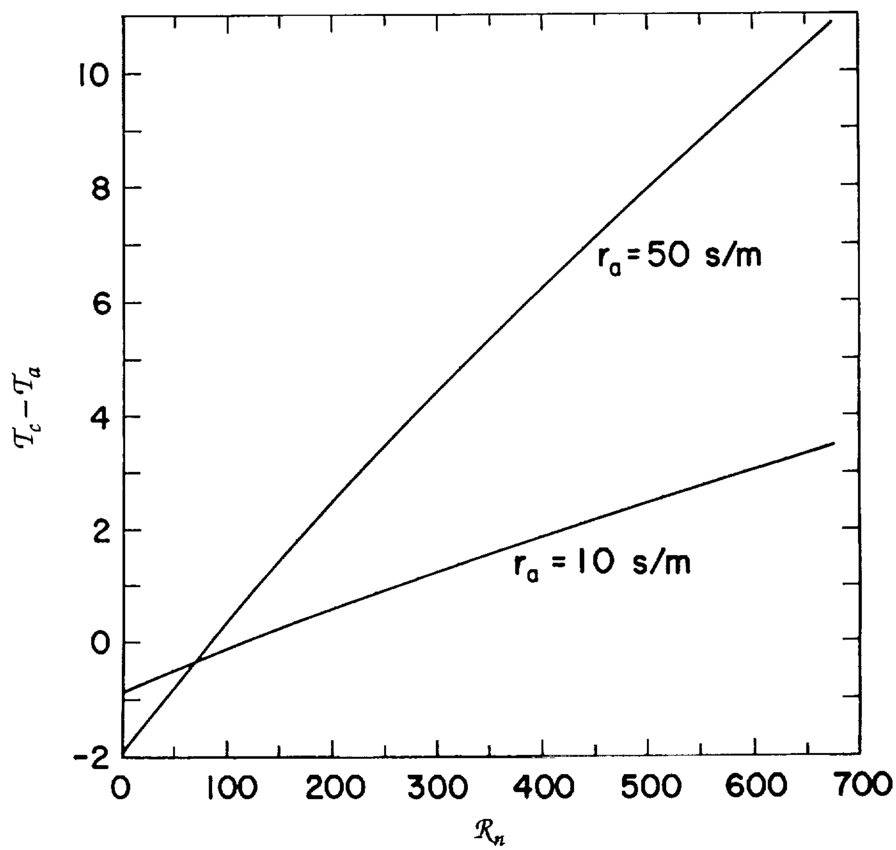

Figure 2. Difference between canopy (T_c) and atmospheric (T_a) temperature vs. net radiation (R_n) for two different surface resistances (r_a) calculated using the simple model described in this paper. Both net radiation and surface resistance are defined for isothermal conditions, that is, canopy and air temperature the same. Hence radiative and stability feedbacks are not included. A relative humidity of 0.7 and air temperature of 22°C are assumed; canopy resistance is 100 s/m (Dickinson et al., 1991).

Subscripts s and a refer to surface and overlying air concentrations of X, u is the magnitude of wind, ρ_a the air density, and C_D a nondimensional transfer coefficient. The factor C_D is from Monin-Obukhov similarity theory for the surface mixed layer of the atmosphere.

For better semblance to the current notation of micrometeorology, we introduce surface resistance $r_a = (C_D u)^{-1}$. Then the flux H of sensible heat is given by

$$H = \rho_a C_p (T_a - T_c)/r_a \tag{2}$$

where T refers to temperature, C_p the specific heat of air, and the subscript c refers to the surface being a canopy. For vegetated surfaces, inclusion of the diffusive resistance by stomates to evapotranspiration E is crucial and so must be included. The integrated effect of the resistances of individual leaves is the canopy resistance r_c such that

$$E = \frac{\rho_a (q_a - q_c)}{r_a + r_c} \tag{3}$$

where q refers to water vapor mixing ratio and q_c is determined for the internal leaf tissues, i.e., for saturated conditions. The fluxes defined by Equations (2) and (3) are illustrated in terms of resistances in Figure 3.

Equations (2) and (3) are constrained by the requirement that sensible plus latent energy flux be balanced by net radiation R_n given by

$$R_n = S_{\downarrow}(1 - \text{ albedo}) + R_{I\downarrow} - \varepsilon\sigma T_c^4 \tag{4}$$

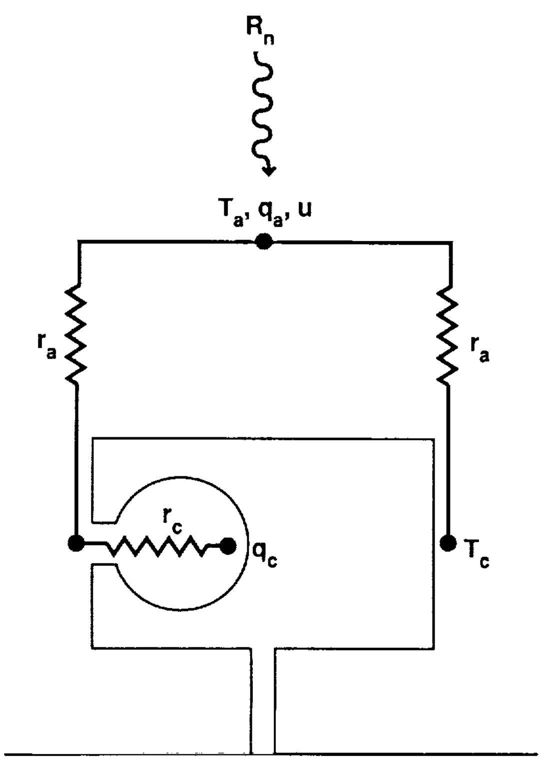

Figure 3. Schematic resistance diagram for the simple generic canopy model described here. See text for definitions of terms (Dickinson et al., 1991).

where $S_{\downarrow}$ = the incident solar energy, $R_{I\downarrow}$ = the downward thermal infrared energy minus any that is reflected, ε = surface emissivity, and σ = the Stefan-Boltzmann constant. The surface energy balance with soil heat flux neglected is written

$$H + \lambda E = R_n \tag{5}$$

where λ = latent heat of evaporation. More realistic models add to Equation (5) a soil heat flux term.

Observational data are conveniently analyzed by the Penman-Monteith approach of combining Equations (2) and (3) with (5), expressing q_c in terms of $T_a - T_c$ and saturated q_a (= q_a^{sat}). The resulting expression relates E to R_n, $q_a - q_c^{sat}$, r_a, and r_c. If all but one of these are measured, the remaining quantity, usually either r_c or E, can be inferred. The main difference between applying these principles to analysis of observational data and applying them to climate modeling is that for the latter none of these quantities can be assumed measured. Rather they must be determined from more basic data and model processes.

Net radiation as described by Equation (4) is conceptually relatively simple. The atmospheric model provides $S_{\downarrow}$ and $R_{I\downarrow}$ and the canopy model provides albedo (according to some combination of specified parameters and a canopy radiative model) and T_c, the latter of which is unknown until Equations (2) through (5) have been solved. As already mentioned, r_a is obtained from boundary-layer theory. The difference $q_a - q_c^{sat}$ is known from the atmospheric model, so the only real complexity is in the specification of r_c and the determination of T_c.

The leaves are assumed to contribute in parallel so conductances $1/r_s$ are averaged, i.e.,

$$r_c = \langle r_s \rangle / \mathrm{LAI} \tag{6}$$

where the wedge brackets denote an inverse average over the range of the canopy leaf area index (LAI), that is, equivalently, the resistances are summed in parallel.

We represent the dependence of r_s on model variables and for different ecosystems by a minimum value r_{smin} and a product of limiting factors, i.e.,

$$r_s = r_{smin}\, f_1(T) f_2(vpd) f_3(PAR) \ldots \tag{7}$$

where each of the fs has a minimum value of 1; f_1 gives a dependence on some characteristic temperature, the most obvious being that of canopy or root zone soil; f_2 a dependence on vapor pressure deficit $vpd = (q_c - q_{ca}) p_s / 0.622$ where q_{ca} is the water mixing ratio in the air outside the leaves and p_s the surface pressure; f_3 a dependence on the photosynthetically active radiation flux density (PAR). Additional dependences include water stress (crucial but implemented differently in different models) and nutrient stress (not yet included in existing models).

Water loss and CO_2 uptake by plants are obviously linked by their sharing of the stomates as their path of dominant diffusion resistance. Physiologists indicate that this linkage is active; that is, stomates act to maintain a constant ratio between water loss and carbon assimilation. Such an active linkage between transpiration and carbon assimilation would provide a basis for understanding the functional dependences in Equation (7). The calculation of carbon assimilation may be required to determine stomatal resistance, or at least may be feasible with little or no additional computation. Such a computation would couple the components of the carbon cycle with fast time scales to the climate model.

There is little guidance for current treatments of the temperature dependence term, beyond the recognition that optimality will generally be achieved in the range of 20–30°C and that stomates will cease functioning at temperatures of freezing, 0°C, and of rapid protein denaturation, ~50°C. Hence BATS makes a quadratic fit to these limits for f_1^{-1} (Figure 4a).

There is also no systematic basis for specifying a vpd dependence for r_s. However, many observations indicate a near linear dependence of f_2^{-1} on vpd, with stomatal closure in the range 0.03 to 0.05 p_s. Hence, SiB and the latent version of BATS assume $f_2^{-1} = 1 - vpd/c$ where $c \cong 0.04\ p_s$ (Figure 4b).

In SiB and BATS, a canopy light model is used to provide light levels at a given depth in the canopy, and hence average or integrate the f_3 component of r_s as indicated in Equation (6). This term differs superficially between BATS and SiB but is functionally the same (Figure 4c). In SiB, it is written

$$f_3 = 1 + \frac{a_2/c_2}{b_2 + PAR} \qquad (8)$$

where a_2, b_2, c_2 are adjustable constants. BATS uses

$$f_3 = \frac{1 + PAR/PAR_c}{r_{smin}/r_{smax} + PAR/PAR_c} \qquad (9)$$

where r_{smax} is the maximum (cuticular) resistance of green leaves, and PAR_c chosen as the light level where $r_s = 2r_{smin}$. Equations (8) and (9) are equivalent, provided

$$r_{smin} = c_2$$

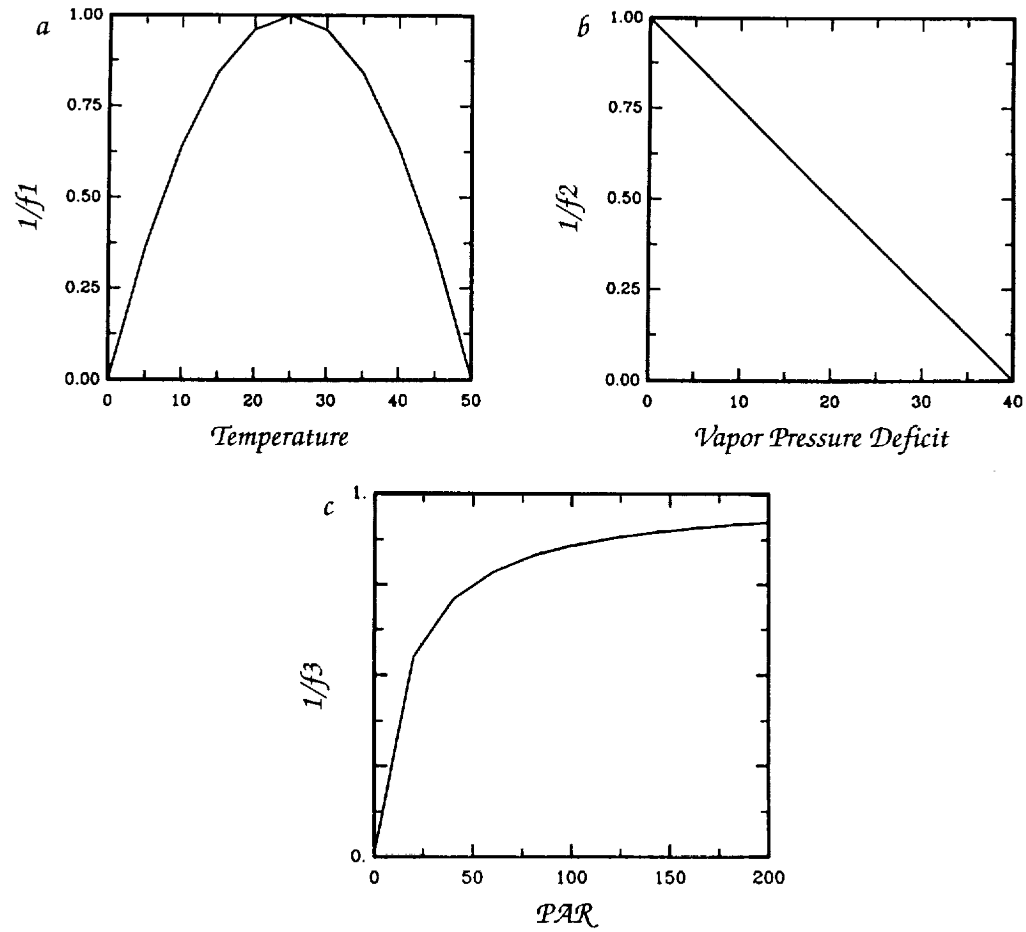

Figure 4. Environmental dependences of inverse of stomatal resistances (i.e., conductance) in BATS model: (a) dependence of conductance on temperature; (b) dependence of conductance on vapor pressure deficit, (c) dependence of conductance on photosynthetically active radiation flux density, or PAR (Dickinson et al., 1991).

$$r_{smax} = c_2 + \frac{a_2}{b_2}$$

$$PAR_c = b_2 + \frac{a_2}{c_2}$$

In W/m^2 of visible radiation, $PAR_c \approx 10$–50. Precise specification of this parameter is neither necessary nor practical. However, factor-of-two variations change r_c significantly, so ideally, an accuracy of better than ±20% is desirable.

For canopies with random leaf angle distributions exposed to direct radiation, the average *PAR* on a leaf surface is determined from

$$PAR = GrS_{\downarrow} \exp(-GD/\mu) \tag{10}$$

where D is the depth into the canopy in units of LAI, μ = cosine of solar zenith angle, $G \sim 0.5$ is the average leaf projection in the direction of the sun, and r = ratio of PAR to total incident solar radiation. BATS uses Equation (10) with diffuse sky radiation accounted for through an additional term which assumes $\mu = 0.5$. SiB does likewise but allows in addition for effects of leaf orientation. Neither model attempts to account in detail for radiation scattered by the foliage, which for PAR has about 0.1 the intensity of the incident solar radiation. SiB uses elegant analytic solutions to Equation (9), whereas BATS uses somewhat simpler and probably equally effective numerical solutions.

Interception, the water from precipitation that evaporates from the canopy without reaching the soil, has similar one-layer parameterizations in all models. The interception parameterizations also provide dew or frost formation when the water vapor gradient from foliage to air reverses. The fraction of canopy surface covered by water has zero resistance r_c and hence, especially in forests, can rapidly evaporate back into the atmosphere. The parameterization choices are the water-holding capacity of the foliage W_{sc}, the fraction of the incident precipitation that is intercepted, and the fraction of the foliage that is covered by water when it has less water than full capacity. After the canopy reaches capacity, all additional precipitation is put into throughfall.

BATS and SiB now use W_{sc} = 0.1 LAI (in mm) for water capacity. This quantity is tuned to give observed canopy interception losses and is somewhat lower than the observed storage of water by leaves. BATS has a "stem area index" surface, as well as LAI, that is also wetted. SiB determines a cross section for interception depending on LAI, similar to that for radiation, whereas BATS assumes all precipitation over vegetation is first captured by foliage. SiB assumes the fraction of canopy surface wetted is the ratio of canopy water to W_{sc}, whereas BATS and a new model being developed at GISS use a 2/3 power. In reality, the fractional wetting is very dependent on the hydrophilic properties of the leaves. Some leaves are partially wetted with smooth water films, while on others, water droplets form. Thus, for application to specific sites and vegetation, the present models may be unrealistic. The values of W_{sc} in SiB and BATS are inferred from two years of Amazon measurements (Shuttleworth, 1988).

Model-Specific Parameterizations

The components of vegetation resistance summarized in the previous section either are treated similarly in essentially all of the vegetation resistance models or else are assumed constant. While details are debatable, we can agree that there is an appropriate functional form consistent with observational information, and that the prescribed constants in the different models are not drastically different. In this section I discuss those components that diverge more drastically in the different land parameterizations.

Water Stress

The models differ in their treatment of the effect of water stress on stomatal resistance, in part because of the lack of reliable quantitative information on the subject. SiB assumes that this contribution to r_s depends on leaf water potential, with the leaf water potential being related to soil water potential through the effect of soil and root diffusion resistance to the water movement. These processes have been represented through more detailed mechanistic models, and some observational information is available for individual sites. However, there is little or no basis for specifying the necessary parameters over large areas.

BATS uses a simplified version of the approach illustrated in Figure 5. Each ecosystem is characterized by a maximum transpiration rate under well-watered conditions, presumed to be determined by root and soil resistance. This maximum rate is reduced below field capacity according to the difference between the soil water suction (negative potential) for wilting and that computed for existing soil moisture. The latter is obtained as an average over the model soil layers, where the average is weighted with the root surface density in each layer. This averaging requires only an estimate of the relative distribution of roots because their absolute contribution is subsumed in the assumed maximum transpiration rate. The most crucial parameter in the BATS treatment of root resistance is this maximum rate, which in principle could be specified from remote sensing.

A similar but even simpler concept is used in the new GISS model being developed, i.e., the contribution to transpiration is reduced in each soil layer from that computed by the unstressed canopy model by the ratio of the total water potential to that for wilting.

Within-Canopy Resistances

SiB and BATS attempt to include the bulk effect of boundary-layer resistances across leaf surfaces. Given a local wind in the canopy, this resistance for heat and moisture can be inferred from laboratory studies. Its accurate specification is limited by the knowledge of the wind distribution within the canopy. BATS simply esti-

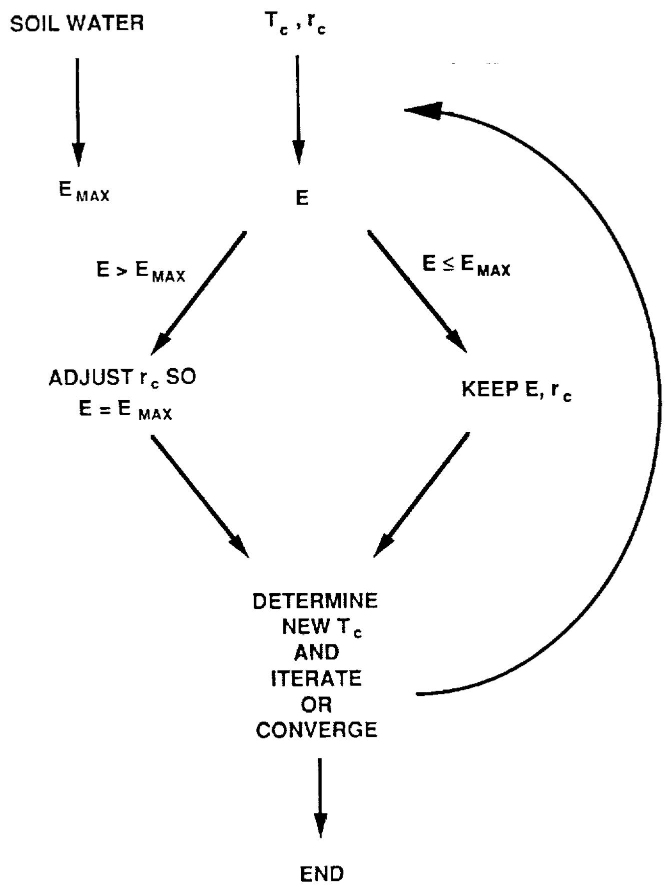

Figure 5. Schematic of the approach used by BATS to determine dependence on stomatal resistance on soil water. Atmospheric demand is determined and compared with the maximum that can be supplied by roots for given soil water. If the former exceeds the later, stomatal resistance is increased to reduce the demand to match supply (Dickinson et al., 1991).

mates this wind from the frictional velocity, whereas SiB bases it on a solution for eddy diffusion within the canopy.

Partial Vegetation

The treatment of partial vegetation in BATS helps illustrate some of the questions that must be faced in treating this issue. The fraction of vegetation covered by a model grid square is prescribed, with a seasonal variation determined from soil temperature. No specification is given as to the spatial scales to be associated with the bare soil fraction. Land classes range from desert (that is, all bare soil) and semidesert (that is, mostly bare soil) to various forest types that are mostly vegetation cover. Sensible and latent fluxes are computed separately for the bare ground and for the fraction of soil under vegetation. Both soil fractions are assumed to have the same temperature and moisture content but to differ in values of overlying wind and transfer coefficients, i.e., different values of r_a are determined for these two fractions. The details of these prescriptions are guided by their reasonableness in the limits of bare soil and full canopy. A more realistic treatment might determine separate soil temperatures and moisture for shade vs. sun and require information on the spatial scales of the vegetation and exposed soil areas.

In SiB, the effects of partial vegetation are incorporated directly into the radiation, momentum (turbulent), and energy transfer submodel. However, no account is taken of larger-scale heterogeneities. The new GISS model separates the grid box into vegetation-covered and bare soil components. Fluxes and soil temperatures are calculated for each of these surface types, and then area-weighted to interface to the first layer of the atmosphere.

Canopy Temperature

Calculation of canopy temperature has been the most difficult aspect of vegetation resistance models in GCMs, because implementing the calculation of surface temperatures can lead to worrisome inaccuracies and, at worst, to severe computational instabilities. Without a successful approach to this question, much greater errors in determination of evapotranspiration can be made than might result from inaccuracies in canopy resistance. In general, we must distinguish between two or more temperatures that are coupled through energy fluxes and must separately satisfy energy balance requirements. For example, SiB and BATS have separate soil and canopy temperatures, and the GISS model in addition distinguishes between bare ground and under-vegetation soil temperatures. These surfaces each have heat capacities, some of which may be zero.

Let T be the vector representing all the model surface temperatures and C a diagonal matrix where elements are individual surface heat capacities. To determine T, knowing its value at a previous time step, we must solve numerically an equation of the form

$$C\frac{\delta \mathrm{T}}{\delta t} - \mathrm{F}(\mathrm{T}) = 0 \qquad (11)$$

where F(T) is a vector whose individual elements represent the sum of energy fluxes into a given surface and t is time. Let superscript n refer to the value of T at the nth model time step. Time steps are Δt. We could first try the simplest solution to Equation (11), i.e.,

$$T^n = T^{n-1} + C^{-1}F\left(T^{n-1}\right)\Delta t \tag{12}$$

This may work with short enough time steps, a few minutes or less, and large enough heat capacities (i.e., provided Δt is small compared to the inverse of the largest eigenvalue of $C^{-1}\ \delta F/\delta T$), but otherwise it can be a prescription for numerical disaster giving wild and growing oscillations from one time step to the next. In the limit of small heat capacities, Equation (11) should approach a statement of energy balance at the present time level. An alternative solution likely to be more accurate and stable is hence

$$T^n = T^{n-1} + C^{-1}F\left(T^{n}\right)\Delta t \tag{13}$$

but this form appears to require already knowing the solution to use. However, its solution may be possible using a Newton iteration, i.e., by writing the *i*th component of $F = F_j$ as

$$F_j\left(T^{n,n+1}\right) = F_j\left(T^{n,i}\right) + \frac{\delta F_j\left(T^{n,i}\right)}{\delta T_k}\left(T_k^{n,i+1} - T_k^{n,i}\right) \tag{14}$$

where the second, *i*th, superscript refers to the number of the iteration, there is a summation over the *k*th subscripts, and T_k is the *k*th temperature. Equation (14) is substituted into Equation (13) for $F(T^n)$ and Equation (13) is solved for $T^{n,i+1}$, taking as a first guess the value of T at the previous time step, that is, $T^{n,1} = T^{n-1}$.

The SiB model specifies soil and canopy heat capacities and uses only the first-guess form of Equation (14). BATS, on the other hand, assumes zero heat capacity for the canopy and first iterates Equation (14) for canopy temperature to convergence, taking soil temperature as that from the previous time step. It then solves Equation (13) as a scalar equation for soil temperature, using the first derived canopy temperature and only the first guess from Equation (14). A further simplification, sometimes used, e.g., at the European Centre for Medium Range Weather Forecasts (ECMWF), is to assume all surfaces have the same temperature, so that Equation (13) is applied as a scalar to derive joint soil and canopy temperatures.

Preliminary examination of the errors from the SiB and BATS solutions for canopy temperature, including lack of conservation of energy, suggests that both approaches can reasonably control errors; SiB conserves energy even if its first iteration is inaccurate, whereas BATS only does so with convergence of the temperature iteration. The BATS solution may be more accurate for canopy-dominated transpiration, but the SiB approach requires less numerical computation and so may be preferable, considering all of the other uncertainties in the parameterizations.

Present Experience with a Land Model Coupled to a GCM

Here we will report on the latest simulation studies with the community climate model version 1 (CCM1) of the National Center for Atmospheric Research (NCAR) coupled to the BATS land surface treatment. Shortcomings are emphasized more than successes to point to where further progress is needed.

Over the last year, we have developed two slightly different versions of the CCM1 tuned to give top-of-the-atmosphere fluxes of solar and longwave radiation that match those observed by satellite. Both start with the standard CCM1 with a diurnal cycle added and the solar radiation calculation done with a treatment developed by A. Slingo of NCAR and the British Meteorological Office. One version in addition calculates clouds with a scheme developed by A. Slingo, initially for the ECMWF model. During these studies, we discovered that previous models which assumed a diurnal average sun would absorb globally about 10 W/m^2 additional solar radiation when a diurnal cycle was added, hence requiring us to adjust the solar cloud treatment to reflect about that much more radiation to get the match with data from the Earth Radiation Budget Experiment (ERBE) satellite (since the treatment had previously been tuned for an average sum).

The bulk of our simulations have been done with the standard CCM1 clouds, and we have not noticed any major differences over land with the other code. The integration includes an eight-year control simulation with a lower boundary over ocean as the new sea surface climatology developed by K. Trenberth and collaborators. The first four years of this integration are used to specify monthly average ocean heat transport (flux correction) needed to balance net surface energy fluxes at each ocean grid point. Otherwise, we treat the ocean at each grid point as a 60-m slab of water for its temperature calculation.

Together with a thermodynamic sea ice model, we have begun to integrate this CCM1/BATS/flux-corrected slab-ocean model for 330 and 660 ppm of CO_2. In both cases we used a 30-m slab for one year to speed up convergence and, for the doubled CO_2, initialized with a uniform 3°C warming over all the ocean points.

The purpose of these integrations is to explore the behavior and feedbacks of a detailed land model (BATS) in a global climate change scenario, comparable to those that have been used previously with simple bucket models.

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