

## Evapotranspiration models with canopy resistance for use in climate models, a review

Robert E. Dickinson<sup>a</sup>, Ann Henderson-Sellers<sup>b</sup>, Cynthia Rosenzweig<sup>c</sup> and Piers J. Sellers<sup>d</sup>

<sup>a</sup>*National Center for Atmospheric Research, Boulder, CO 80307, USA*

<sup>b</sup>*Macquarie University, School of Earth Sciences, Sydney, N.S.W. 2109, Australia*

<sup>c</sup>*Columbia University and NASA/Goddard Institute for Space Studies, New York, NY 10025, USA*

<sup>d</sup>*NASA/Goddard Space Flight Center, Greenbelt, MD 20771, USA*

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### ABSTRACT

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The most advanced climate models are now incorporating effects of vegetation into their surface energy and hydrological balance formulations. These efforts are reviewed here with emphasis on the canopy resistance component through development of a simple generic canopy model that includes the relatively well-established processes common to several canopy submodels. We also review where the land surface treatments diverge, in particular in their consideration of water stress, partial vegetation, within-canopy resistances and computational algorithms for the determination of canopy temperature. The inclusion of canopy submodels in climate models may be of questionable utility without improvements in some aspects of atmospheric simulations, such as the modeled distributions of precipitation and incident surface radiation. Improved treatments of the heterogeneous nature of precipitation and land surface properties are also needed.

### INTRODUCTION

Land processes have long been represented as a component of general circulation models (GCM) (e.g. Manabe, 1969) to impose conservation of energy and water mass on the system, or failing that, at least a reasonable lower boundary condition. The atmospheric components of the models (Fig. 1) provide the surface with fluxes of solar and thermal infrared radiation and precipitation in the form of rain or snow, and 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 runoff. The total radiation absorbed by the surface is balanced by emission of thermal infrared radiation,

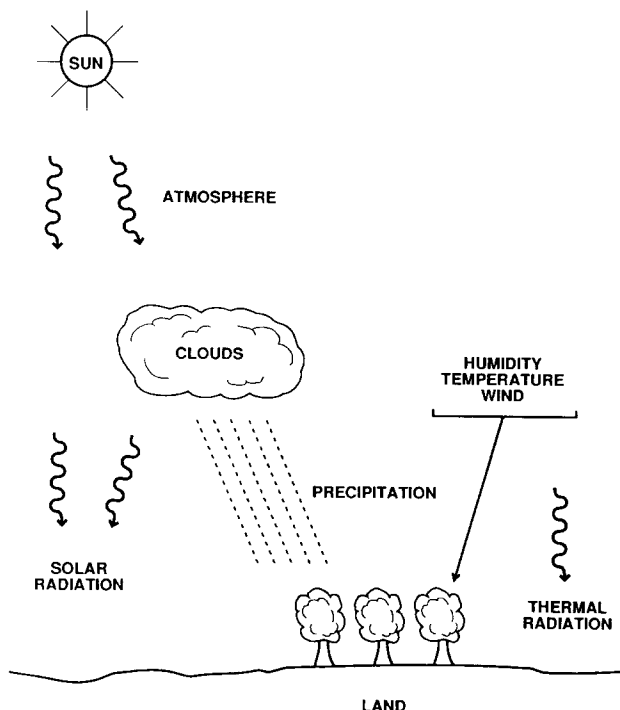


Fig. 1. Sketch of the inputs needed at the surface from the atmospheric model.

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 bucket model for the above processes (Manabe, 1969) is supposed to evaporate at the same rate as a wet surface (zero canopy resistance) during well-watered conditions. It holds a maximum water, typically 0.15 m, which corresponds to the available soil water, i.e. the water in the rooting zone at some average field capacity minus that still present at some average wilting point. However, because the bucket model does not include the process of diffusion of water in soils or canopy resistance, its evaporation rates are 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: (a) it can extract soil water from a greater depth than would evaporation from bare ground; (b) it retards the rate of evapotranspiration from the potential rate through resistance by the stomates to molecular diffusion of water. The standard bucket models are especially unrealistic during

dry conditions when they continue to evaporate at near potential rate. Observations indicate that if the bucket model were to use the temperature of an irrigated surface rather than that of the actual surface, it would provide a more realistic dependence on soil dryness, at least for short vegetation (e.g. Davies and Allen, 1973).

The initial bucket model formulation, neglecting any reference to canopy resistance, is still widely used in GCM for studies of climate change, and in particular 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).

The Manabe (1969) parameterization assumes that land properties are globally uniform. In reality, the aerodynamic resistance of a short grass 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, as shown in Fig. 2 (e.g. Dickinson and Henderson-Sellers, 1988). These warmer temperatures reduce net radiation (by increasing long-wave emission) by up to

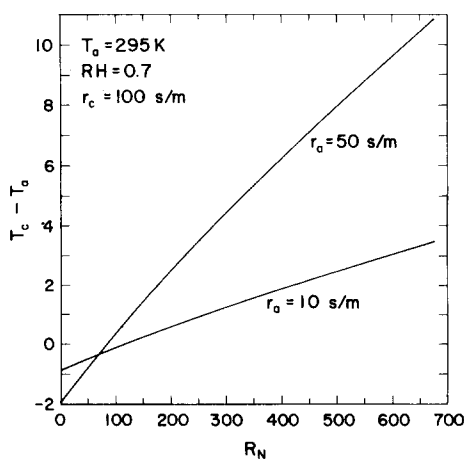


Fig. 2. Difference between canopy and atmospheric temperature vs. net radiation for two different surface resistances calculated using the simple model described in this paper. Both net radiation and surface resistance are defined for isothermal conditions, i.e. canopy and air temperature the same. Hence radiative and stability feedbacks are not included. A relative humidity of 0.7 and air temperatures of 22°C are assumed.

several tens of  $\text{W m}^{-2}$ . A comparable additional reduction in net radiation is implied by the higher albedo of 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 the magnitudes of the net radiation flux and the vapor pressure deficit (e.g. Jarvis et al., 1976; Rowntree, 1988). 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.

Jarvis and McNaughton (1986) reviewed models of canopies coupled to planetary boundary-layer models. The evapotranspiration inferred from these coupled models is relatively insensitive to canopy resistance and is approximated by the Priestley and Taylor (1972) expression. Shuttleworth (1988) found that the Priestley–Taylor formulation gives annual average evapotranspiration at an Amazon forest site within 5–10% of that observed, but is less accurate for individual months and days, e.g. giving an overestimate of 20–30% ‘on fine days’ and an underestimate by 20–70% ‘on wet days’.

Laval et al. (1984) have introduced into the French Laboratory for Dynamic Meteorology model (LMD) a version of the Priestley–Taylor formulation derived by Perrier (1982) that decreases the Priestley–Taylor coefficient ‘ $\alpha$ ’ with decreasing soil moisture. Because GCM already include planetary boundary layers and need reasonably accurate estimates of evapotranspiration on an hourly basis, most GCM groups have not found the Priestley–Taylor approach as attractive as the explicit inclusion of canopy models.

In sum, the earlier bucket and Priestley–Taylor models have been useful, but are not capable of representing the diurnal cycle and other important features of biophysical controls over land. The thrust of this paper is to summarize how more realistic modeling of biosphere–atmosphere interactions may be achieved simply enough to be incorporated into GCM.

#### DISCUSSION OF SIMPLE GENERIC CANOPY MODELS

To help summarize the common content of several canopy models developed for application in GCM (i.e. the ‘biosphere atmosphere scheme’ (BATS) Dickinson et al., 1986; Dickinson, 1989; the ‘simple biosphere model’ (SiB) Sellers et al., 1986; and the model under development at the Goddard Institute for Space Studies (GISS), Abramopoulos et al., 1988), we consider the ‘lowest common denominator’ ingredients that they all contain, then describe why various models introduce further complexity in various features. The following derivations capture the essence of all the model treatments under full-canopy conditions, while leaving out many details such as the 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 by the aerodynamic expression

$$F_x = \rho C_D u (X_s - X_a) \quad (1)$$

Subscripts s and a refer to surface and overlying air concentrations of  $X$ ,  $u$  is the magnitude of wind,  $\rho$  is the air density, and  $C_D$  is a non-dimensional 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

$$H = \rho c_p (T_a - T_c) / r_a \quad (2)$$

where  $T$  refers to temperature,  $c_p$  is 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

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

where  $q$  refers to specific humidity and  $q_c$  is determined for the internal leaf tissues, i.e. for saturated conditions. The fluxes defined by eqns. (2) and (3) are illustrated in terms of resistances in Fig. 3.

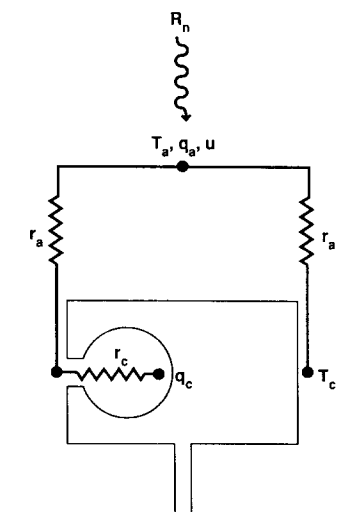


Fig. 3. Schematic resistance diagram for the simple generic canopy model described here.

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

$$R_n = R_s(1 - \text{albedo}) + R_{l\downarrow} - \epsilon\sigma T_c^4 \quad (4)$$

where  $R_s$  = the incident solar energy,  $R_{l\downarrow}$  = the downward long-wave radiation minus any that is reflected,  $\epsilon$  = surface emissivity and  $\sigma$  = Stefan-Boltzmann constant. The surface energy balance with soil heat flux neglected is written

$$H + \lambda E = R_n \quad (5)$$

where  $\lambda$  = latent heat of evaporation. More realistic models add a soil heat flux term to eqn. (5).

Observational data are conveniently analyzed by the Penman-Monteith approach of combining eqns. (2) and (3) with (5), expressing  $q_c$  in terms of  $T_a - T_c$  and saturated  $q_a (= q_a^{\text{sat}})$ . The resulting expression relates  $E$  to  $R_n$ ,  $q_a - q_a^{\text{sat}}$ ,  $r_a$  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 the 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 eqn. (4) is conceptually relatively simple. The atmospheric model provides  $R_s$  and  $R_{l\downarrow}$  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 eqns. (2)–(5) have been solved. As already mentioned,  $r_a$  is obtained from boundary-layer theory. The difference  $q_a - q_a^{\text{sat}}$  is known from the atmospheric model, so the only real complexity is in the specification of  $r_c$  and the determination of  $T_c$ .

All the authors' models represent  $r_c$  as a summation of the contributions of stomatal resistance,  $r_s$ , of individual leaves. The leaves are assumed to contribute in parallel so conductances  $1/r_s$  are averaged, i.e.

$$r_c = \langle r_s \rangle / L \quad (6a)$$

where the angled brackets denote an inverse average over the range of the canopy leaf area index,  $L$

$$\langle () \rangle = \frac{L}{\int^L dL / ()} \quad (6b)$$

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

$$r_s = r_{s\min} f_1(T) f_2(D) f_3(\text{PAR}) \dots \quad (7)$$

where each of the  $f$  values has a minimum value of one;  $f_1$  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  $D = (q_c - q_{ca})p_s/0.622$  where  $q_{ca}$  is the specific humidity 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 dependencies include water stress (crucial, but implemented differently in the different models) and nutrient stress (not yet included in existing models).

Water loss and  $\text{CO}_2$  uptake by plants are obviously linked by their sharing of the stomates as their path of dominant diffusion resistance. Physiologists (Farquhar and Sharkey, 1982; Ball et al., 1987) indicate that this linkage is active, i.e. 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 dependencies in eqn. (7). The calculation of carbon assimilation may be required to

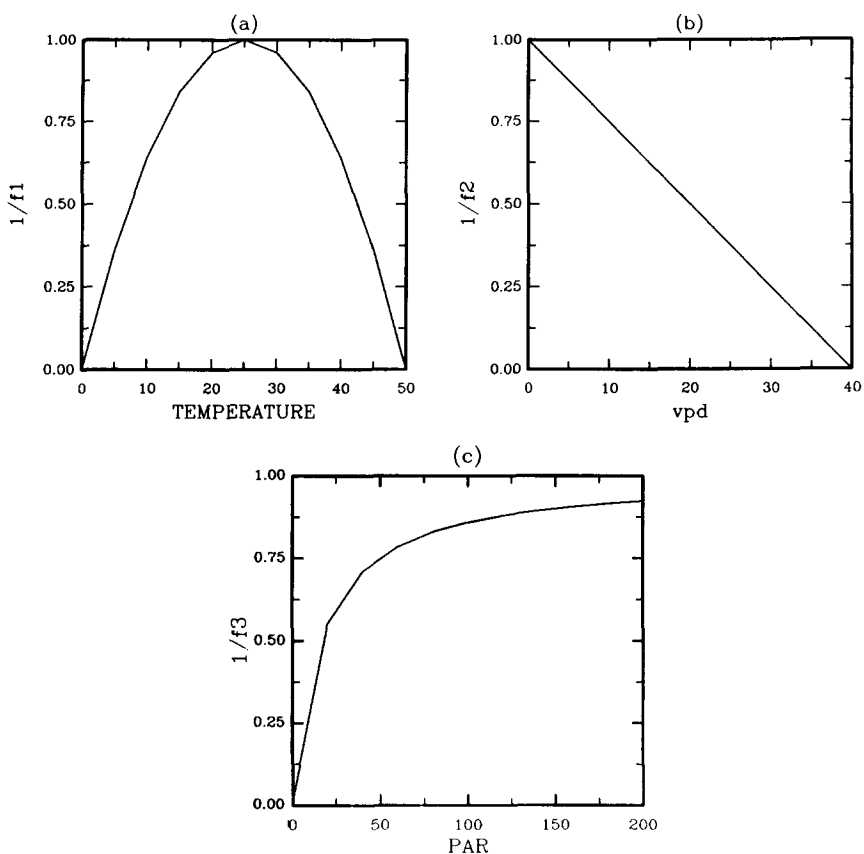


Fig. 4. Environmental dependencies of the inverse of stomatal resistances (i.e. conductance) in BATS model: (a) dependence of conductance on temperature; (b) dependence of conductance on vapor pressure deficit; (c) dependence of conductance on PAR.

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 20–30°C, that stomates will cease functioning at temperatures of freezing (0°C) and of rapid protein denaturation (about 50°C). Hence BATS makes a quadratic fit to these limits for  $f_1^{-1}$  (Fig. 4(a)). Since different species and different genotypes acclimatize their photosynthesis to different optimum temperatures, it is likely that they do the same for transpiration, i.e. a realistic  $f_1$  should reach minimum values at lower temperatures for cold climate ecosystems.

There is also no systematic basis for specifying a  $D$  dependence for  $r_s$ . However, many observations indicate a near linear dependence of  $f_2^{-1}$  on  $D$ , with stomatal closure in the range 30–50 hPa. Hence, SiB and the latest version of BATS assume  $f_2^{-1} = 1 - D/c$  where  $c \approx 0.04 p_s$  (Fig. 4(b)). The GISS model and earlier versions of BATS neglect this dependence.

The dependence of  $r_s$  on PAR (assumed equivalent to visible radiation for model calculation) is ignored in the GISS model by assuming an average  $\langle r_s \rangle$  in eqn. (6a). 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 eqn. (6a). This term differs superficially between BATS and SiB, but is functionally the same (Fig. 4(c)). In SiB, it is written

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

where  $a_2$ ,  $b_2$  and  $c_2$  are adjustable constants, and  $R_p$  is the PAR. BATS uses

$$f_3 = \frac{1 + R_p/R_{pc}}{r_{smin}/r_{smax} + R_p/R_{pc}} \quad (9)$$

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

$$r_{smin} = c_2$$

$$r_{smax} = c_2 + a_2/b_2$$

$$R_{pc} = b_2 + a_2/c_2$$

In  $W m^{-2}$  of visible radiation,  $R_{pc} \approx 10$ –50. Precise specification of this parameter is neither necessary nor practical. However, factor of two varia-



tions change  $r_c$  significantly so, ideally, an accuracy of better than  $\pm 20\%$  is desirable.

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

$$R_p = GrS_i \exp(-GL/\mu) \quad (10)$$

where  $L$  is the depth into the canopy in units of leaf area index (LAI)  $\mu = \cos$  of solar zenith angle,  $G \approx 0.5$  is the average leaf projection in the direction of the sun and  $r =$  ratio of  $R_p$  to total incident solar radiation. BATS uses eqn. (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  $R_p$  has about one-tenth the intensity of the incident solar radiation. SiB uses elegant analytic solutions to eqn. (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.

All the authors' models now use  $W_{sc} = 0.1 L$  (in mm) for water capacity. BATS has a 'stem-area index' surface as well as  $L$  that is also wetted. SiB determines a cross-section for interception depending on  $L$  similar to that for radiation, whereas the other models assume 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 GISS use a two-thirds power, following Deardorff (1978). 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 other leaves water droplets form. Thus, for application to specific sites and vegetation, the present models may be unrealistic. The  $W_{sc}$  in SiB and BATS are inferred from 2 years of Amazon measurements (Shuttleworth, 1988).

#### MODEL-SPECIFIC PARAMETERIZATIONS

The components of vegetation resistance summarized in the previous section are either treated similarly in all 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, we discuss those components that diverge most drastically in the different land parameterizations.

### *Water stress*

The models all 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 same approach, as illustrated in Fig. 5. Each ecosystem is characterized by a maximum transpiration rate under well-watered conditions, presumed to be determined by root and soil resis-

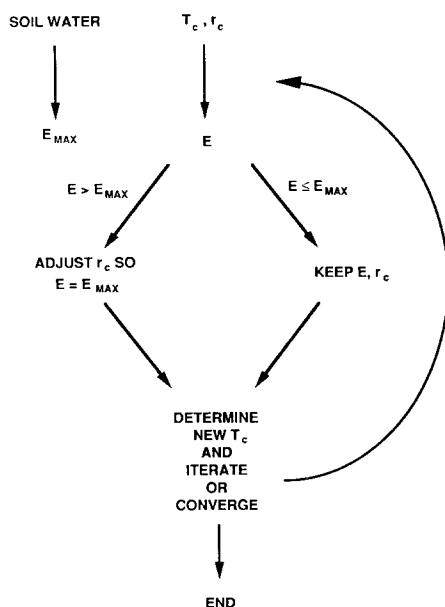


Fig. 5. Schematic of the approach used by BATS to determine the dependence of stomatal resistance on soil water. Atmospheric demand is determined and compared with the maximum that can be supplied by the roots for given soil water. If the former exceeds the latter, stomatal resistance is increased to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

tance. 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 weighed 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.

The GISS model uses a similar but even simpler concept, 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 estimates this wind from the frictional velocity, whereas SiB bases it on a solution for eddy diffusion within the canopy. GISS neglects any boundary-layer resistances 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 (i.e. all bare soil) and semidesert (i.e. 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  $r_a$  values 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 versus 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 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 weighed 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 GCM, 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 the determination of evapotranspiration can be made than might result from inaccuracies in canopy resistance. A general description of what is done by various models requires the following somewhat formal mathematical treatment. Details of the specification of heat capacities and soil heat flow cannot be given here. In general, all the authors' models 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

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

where  $F(T)$  is a vector whose individual elements represent the sum of energy fluxes into a given surface. Let the superscript  $n$  refer to the value of  $T$  at the  $n$ th model time step. Time steps are  $\Delta t$ . We have all first tried the simplest solution to eqn. (11), i.e.

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

This may work with short enough time steps, a few minutes or less, and large enough heat capacities, (i.e. provided  $\Delta t$  is small compared with 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, eqn. (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

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

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

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

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 eqn. (13) for  $F(T^n)$  and eqn. (13) is solved for  $T^{n,i+1}$ , taking as a first guess the value of  $T$  at the previous time step, i.e.  $T^{n,1} = T^{n-1}$ .

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

## CONCLUSIONS AND DISCUSSION

The addition of vegetation resistance into climate models (which is done with broad similarities in all the models discussed), along with other parameterizations of land surfaces, is a useful and necessary step. It improves the realism of calculated surface energy fluxes and hence the representation of surface climate as needed for studies of future climate change. However, inclusion of a canopy submodel demands improvements in other atmospheric aspects of the GCM to be able to achieve its full potential. In particular, with improved realism of surface treatments, we realize how sensitive simulations

of land climate can be to inaccuracies in modeled distributions of precipitation and cloud attenuation of incident surface radiation.

Assuming that the atmospheric simulations can be made accurate enough to warrant it, the general directions of further improvements in treatment of vegetation resistance are clear. We must learn how to represent the vertical fluxes from heterogeneous land surfaces. The emphasis of GCM land-surface modeling has swung from a virtually bare planetary surface to replication of a highly detailed, but largely homogeneous vegetation; the appropriate surface probably lies somewhere in the middle, with large-scale patches of simply described heterogeneous vegetation and bare soil.

On spatial scales of up to tens of kilometers, advective effects tend to average out atmospheric differences. However, on all scales down to plant level, the many different soil and vegetated surfaces are exposed to differing patterns of radiation and precipitation. The consequence is not only a mosaic of surface temperatures and local evapotranspiration, but also a system that is not described very well by mean properties because of the intrinsic non-linearities of the energy exchange properties. What specific first steps will be taken to improve the situation will differ between modeling groups. More systematic inclusion of multiple surface energy balances, as outlined in the preceding section, is one possible improvement. Another is to generalize the canopy representations to better give the statistical distributions of soil and canopy properties, as well as those of solar radiation and precipitation. This might be achieved by using present models, but integrating them over many points within each GCM grid cell.

Another direction of improvement is to incorporate parameter values more specific to given land and covers. All the models contain many default or generic parameter values where it has not been possible to resolve differences between different cover types. In particular, the parameters used in the leaf/canopy resistances of the models need clearer distinctions between different ecosystems. However, as discussed, this may be possible only through joint modeling of carbon uptake and water loss.

In sharpening the distinctions between different cover types, we need to improve the global specifications of the cover types used in the models, as well as treat more complex surfaces. The seasonal and intra-annual variations of the properties of a given cover type may, furthermore, have a larger impact on vegetation resistance than the variations between cover types. These variations require modeling the seasonal generation, growth, senescence and death of green leaf cover, responding to specific phenologies as well as temperature and water stresses.

The relationships between soil moisture and canopy resistance and, in particular, the range of soil water storage over which the canopies can function, are also likely to need improvement. Large-scale remote sensing approaches

may be the best way to gain this improvement, although little useful information from remote sensing has yet been incorporated into the models.

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