A Simple Biosphere Model (SiB) for Use within General Circulation Models

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ABSTRACT

A simple but realistic biosphere model has been developed for calculating the transfer of energy, mass and momentum between the atmosphere and the vegetated surface of the earth. The model is designed for use in atmospheric general circulation models.

The vegetation in each terrestrial model grid area is represented by two distinct layers, either or both of which may be present or absent at any given location and time. The upper vegetation layer represents the perennial canopy of trees or shrubs, while the lower layer represents the annual ground cover of grasses and other herbaceous species. The local coverage of each vegetation layer may be fractional or complete but as the individual vegetation elements are considered to be evenly spaced, their root systems are assumed to extend uniformly throughout the entire grid area. Besides the vegetation morphology, the physical and physiological properties of the vegetation layers are also prescribed. These properties determine (i) the reflection, transmission, absorption and emission of direct and diffuse radiation in the visible, near infrared and thermal wavelength intervals; (ii) the interception of rainfall and its evaporation from the leaf surfaces; (iii) the infiltration, drainage and storage of the residual rainfall in the soil; (iv) the control by the photosynthetically active radiation and the soil moisture potential, inter alia, over the stomatal functioning and thereby over the return transfer of the soil moisture to the atmosphere through the root-stem-leaf system of the vegetation; and (v) the aerodynamic transfer of water vapor, sensible heat and momentum from the vegetation and soil to a reference level within the atmospheric boundary layer.

The *S*imple *B*iosphere (SiB) has seven prognostic physical-state variables: two temperatures (one for the canopy and one for the ground cover and soil surface); two interception water stores (one for the canopy and one for the ground cover); and three soil moisture stores (two of which can be reached by the vegetation root systems and one underlying recharge layer into and out of which moisture is transferred only by hydraulic diffusion and gravitational drainage).

1. Introduction

General circulation models (GCMs), whether used for numerical weather prediction or climate simulation, require a determination of the fluxes of radiation, water vapor, sensible heat and momentum across the lower boundary of the atmosphere. Similarly, one-dimensional energy balance and mass transfer models, used for the local estimation of evapotranspiration and photosynthesis, utilize boundary conditions set some small distance above the surface within the atmospheric boundary layer. It may be said, therefore, that the recent interest in modeling land surface-atmosphere interactions has converged from two opposite directions—opposite, that is, in terms of scale and the given boundary conditions. On the one hand, climatologists and meteorologists are concerned with the large-scale radiation balance and the partition of surface net radiant energy into sensible and latent heat flux, as these processes are boundary conditions on a fairly sensitive dynamic system. Consequently, this group's principal aim is to integrate all of the small-scale physical processes, i.e., interception of radiation by plant leaves and transpiration of water (from soil to atmosphere via the plant's internal transport system), over the horizontal scales commensurate with general circulation models: 100–400 kilometers. On the other hand, biophysicists and ecologists are interested in the relationships extant between organism and environment—relationships that span length scales from the molecular level (in the case of photosynthesis) to several kilometers (when considering energy flow in ecosystems). For the most part, each group uses the findings of the other as boundary conditions to drive their own theoretical and numerical models; for example, meteorologists make use of simple empirical "β-functions" to parameterize the dependence of evapotranspiration on soil moisture while ecologists take climate as a given forcing in developing theories of speciation, adaptation, and the structuring of plant communities. It is only in the past few years that the two fields have come sufficiently close to allow studies of the interactions between the biosphere and the atmosphere. Hitherto, these have taken the form of simple sensitivity experiments conducted with general circulation models wherein the characteristics of the land surface were changed (often drastically) from their control conditions in order to allow an assessment of the magnitude of the feedback effects on the atmosphere. Mintz (1984) has reviewed 11 such experiments, conducted by different research groups, and has pointed out that in many respects the results are in agreement, i.e. the calculated large-scale fields of air temperature, humidity, wind, and precipitation are strongly affected by the drag, radiation absorption, and energy partition properties of the underlying surface. Similarly, a number of studies have been carried out where prescribed atmospheric conditions were used to drive Soil–Plant–Atmosphere Models (SPAMs) capable of simulating the effects of changing the vegetation type of an area. Spittlehouse and Black (1981), Sellers (1981) and Sellers and Lockwood (1981) demonstrated that the energy and water balance of a region could be highly dependent upon plant physiology and morphology; the former controls the transpiration rate while the latter has a marked effect on the interception loss rate (that portion of rainfall that is held on plant leaves as liquid water and then evaporates to the atmosphere without reaching the soil moisture store). Goudriaan (1977), Otterman (1981) and Kimes (1984) have shown how plant structure affects the radiation absorption of a region by the trapping of light in multiple reflections by plant elements. In such cases, where the hemispheric albedo of an area may have a significant climatic feedback effect, for example, in marginal desert areas or taiga-tundra boundaries, the presence or absence of vegetation may be a significant element in the local radiation climatology.

These and other research efforts have tended to be heavily weighted, in terms of physical realism, toward the specialty of the investigating group and so the nature of *two-way* interactions is still poorly understood. This paper describes a *Si*mple model of the *B*iosphere, SiB, suitable for operation within general circulation models (GCMs), which hopefully will increase our understanding of the interactions. The proposed model is intended to be as physically and biologically realistic as possible. Initially, the model will be physiologically reactive to atmospheric conditions but the type, density and health of the vegetation will be prescribed as functions of season and location. This initial version of the model, which is presented here, should provide more realistic fluxes of sensible heat, latent heat, and momentum over the continents than existing formulations.

Later studies will explore the possibility of making the biosphere model phenologically interactive with the atmospheric GCM, which will permit the simulation of the effects of anomalous climatic conditions on the surface and the prediction of atmospheric circulation patterns over periods on the order of a year. Formulation of governing equations determining the type, and hence the morphological and physiological attributes, of the vegetation would enable the combined biosphere–atmosphere model to yield predictions over longer time scales, on the order of decades or longer. Should the goal of a fully interactive biosphere–atmosphere model be achieved, simulation studies of complex global problems would become feasible. For example, the possible effects of the increase in atmospheric CO_2 concentration would best be investigated using coupled models of this nature.

2. Sensitivity of the atmosphere to the state of the land surface

The dependence of the large-scale atmospheric circulation and rainfall on land surface boundary conditions has been the subject of discussion and speculation for many years but only recently have GCMs been used to calculate the atmospheric response to prescribed changes in the land surface boundary conditions.

Charney et al. (1977) and subsequent researchers showed that changing the land surface albedo can produce significant changes in the large-scale atmospheric circulation and rainfall. Walker and Rowntree (1977), Shukla and Mintz (1982) and others have demonstrated that changing the available soil moisture may have large feedback effects on the continental climates.

Recently, Sud et al. (1986) have shown that changing the land surface roughness alters the convergence of the horizontal water vapor transport in the atmospheric boundary layer of a GCM and produces large changes in the distribution of the convective precipitation.

3. Modeling strategies

In almost all existing GCMs (see the review by Carson, 1981), the fluxes of radiation, heat (sensible and latent) and momentum across the lower boundary of the atmosphere are treated as independent processes. Usually, the surface fluxes are made to depend on an independently specified surface albedo, an independently specified surface roughness length or drag coefficient and an independently formulated dependence of evapotranspiration on soil moisture. The latter component has generally been conceptualized as a "bucket" in which the level of the water is lowered when evaporation exceeds precipitation, and is raised when precipitation is larger, up to the point at which the bucket overflows and produces "runoff." Over some broad range in the level of the water in the bucket, which varies from model to model, the rate of evaporation is taken as equal or nearly equal to that from a freely evaporating surface, and is reduced only when the water level is low. Although this approach places reasonable bounds on surface evaporation rates, it is not the way in which water vapor is transferred from the land to the atmosphere in the real world.

In reality, plants are not passive sponge-like structures as implied by the bucket model, which views the vegetation as a pervious sheet separating the soil from the atmosphere. Plants are alive and have evolved into complex organisms that regulate the passage of water and gas through their systems in an efficient manner to maximize their prospects of growth and survival. Modeling efforts directed at understanding the links between surface and atmosphere should take this into account and introduce some of the influences of the physiology and morphology of the vegetation to be found in a given region.

In view of the above, the adopted strategy in formulating SiB has been to model the vegetation itself and to let the vegetation determine the ways in which the land surface interacts with the atmosphere. These interactions may be summarized as follows:

(i) *Radiation absorption.* The spectral properties of leaves and multiple reflections between them make vegetation canopies highly absorbent in the visible [photosynthetically active radiation (PAR)] wavelength interval of 0.4–0.72 μm and moderately reflective in the near-infrared region (0.72–4.0 μm). In contrast, bare ground generally exhibits a gradual increase in reflectivity with wavelength over the interval of 0.4–4.0 μm.

(ii) *Biophysical control of evapotranspiration.* Plant metabolism is based on the photosynthetic reaction whereby shortwave radiative energy is used to combine water and atmospheric carbon dioxide into sugars and other organic compounds. To do this, plants must allow for the transfer of CO_2 from the atmosphere to the cellular sites of photosynthesis located inside the leaves. This necessitates the maintenance of an open pathway between the atmosphere and the saturated tissues inside the leaf, which leads to an inevitable loss of water vapor over the same route. Higher plants regulate the amount of gas exchange (and hence water loss) by means of valve-like structures on the leaf surface (stomates). These appear to react to the environmental conditions of PAR flux density, temperature, humidity and leaf water potential so as to maximize the ratio of CO_2 influx to water vapor efflux (see Farquhar and Sharkey, 1982) or to conserve moisture in times of water stress. Clearly, since the radiant energy absorbed by the surface is mainly divided between sensible and latent heat, any decrease in the evapotranspiration rate will be approximately balanced by a concomitant increase in sensible heat loss.

Vegetation canopies also intercept precipitation and can store the equivalent of several millimeters of water on the leaf surfaces. The evaporation of this intercepted water reduces the precipitation input into the soil and lowers the local Bowen ratio, sometimes to negative values.

(iii) *Momentum transfer.* Vegetation canopies usually present a relatively rough, porous surface to the planetary boundary layer airflow. The resultant turbulence enhances the transport of sensible and latent heat away from the surface while exerting a drag force which may be significantly larger than that produced by bare ground.

(iv) *Soil moisture availability.* The depth and density of the vegetation root systems determines the amount of soil moisture available for evapotranspiration.

(v) *Insulation.* The soil surface under a dense vegetation canopy intercepts less radiation and may also be aerodynamically sheltered. For these reasons, the energy available to the covered soil is small and the component terms of the soil energy budget (evaporation, sensible heat flux and ground heat flux) are correspondingly reduced.

In designing a model that takes account of the above factors, we are constrained by the need for a solution that will simulate the energy budget of all vegetation communities by a simple alteration of the model's input parameters. This criterion is made all the more necessary by the use of vector processor computers in GCM calculations which perform essentially the same mathematical operations on all of the model grid areas simultaneously.

4. The model structure

The morphological and physiological characteristics of the vegetation community in a grid area are used to derive coefficients and resistances that govern the momentum, radiation, and sensible and latent heat (water vapor) fluxes between the surface and atmosphere. All of these fluxes depend upon the state of the vegetated surface and the atmospheric boundary conditions.

Detailed descriptions of the methods used to transform the morphological and physiological characteristics of a given vegetation community into the model parameters are given in subsequent sections of this paper. This section is confined to the definition of the atmospheric boundary conditions, the morphological, physiological and physical parameters, and the prognostic variables and governing equations of SiB.

a. Atmospheric boundary conditions for SiB

The following atmospheric variables are used as upper boundary conditions by SiB.

1) AIR TEMPERATURE, VAPOR PRESSURE AND WIND SPEED (T_r, e_r, u_r)

All general circulation models provide grid area averaged values of air temperature, T_r, water vapor pressure, e_r, and wind speed, u_r, at some reference level, z_r, within the atmospheric boundary layer. These are the atmospheric physical-state variables that, in part, determine the fluxes of energy, mass and momentum across the lower boundary of the atmosphere.

2) FIVE COMPONENTS OF THE INCIDENT RADIATION [$F_{\Lambda,\mu(0)}$]

The SiB model requires the magnitudes of five components of the downward radiation flux, $F_{\Lambda,\mu(0)}$, through the bottom of the atmosphere (where Λ refers to wavelength interval, and μ to direction).

Visible or PAR (<0.72 μ) direct beam radiation, $F_{s,b(0)}$

Visible or PAR (<0.72 μ) diffuse radiation, $F_{s,d(0)}$

Near Infrared (0.72–4 μ) direct beam radiation, $F_{n,b(0)}$

Near Infrared (0.72–4 μ) diffuse radiation, $F_{n,d(0)}$

Thermal Infrared (>4 μ) diffuse radiation, $F_{t,d(0)}$

The incoming shortwave radiation is partitioned into two wavelength intervals, F_s and F_n, as the spectral properties of green leaves exhibit a sharp change at around 0.72 μm. The visible or photosynthetically active radiation (PAR), F_s, is absorbed strongly by the chlorophyll in the leaves while the near infrared radiation, F_n, is largely scattered. The incoming shortwave radiation is further partitioned into diffuse and direct beam components because the absorption and transmittance coefficients of vegetation canopies are highly dependent upon the angle of the incident flux.

3) PRECIPITATION RATE (P)

The grid area-averaged precipitation rate P is provided by all atmospheric GCMs. Most GCMs calculate two components of the precipitation: a grid area component and a convective component that represents nonuniform shower-type precipitation.

b. *Morphological parameters of SiB*

In SiB, the world's vegetation is divided into two morphological groups: trees or shrubs, which constitute the upper story canopy vegetation; and the ground cover, which consists of grasses and other herbaceous plants. As shown schematically in Fig. 1a, varying amounts and densities of the canopy vegetation and ground cover may coexist within a given grid area.

When we compare SiB with the physiognomic classification of the natural vegetation types of the world given by Kuchler (1949, 1983), we find that of his 32 vegetation types, 31 can be represented by SiB. (Only one small part of the globe, a region of about 500 km^2 in northeast India, is classified by Kuchler as having two layers of upper story vegetation, trees and shrubs—his vegetation type DBs, consisting of broadleaf deciduous trees and broadleaf evergreen shrubs. This vegetation type must be represented in some other way when using SiB.)

The root and soil structure of SiB is shown in Fig. 1b. The upper story vegetation consists of perennial plants with persistent roots assigned to a fixed depth taken to be the bottom of soil layer 2. The ground cover is made up of annual plants and may have a time-varying root depth that may not exceed the bottom of soil layer 2. There is an upper, thin soil layer (soil layer 1), from which 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. Beneath the root zone, there is an underlying recharge layer (soil layer 3) where the transfer of water is governed only by gravitational drainage and hydraulic diffusion.

The parameters that define the morphology of the Simple Biosphere are given in Table 1a.

c. *Physical and physiological parameters of SiB*

The physical properties of the vegetation and soil and the morphological and physiological properties of the vegetation govern the interception of radiation and the transfer of moisture through the soil–plant system. The relevant parameters used in SiB are given in Tables 1b and 1c.

d. *Prognostic physical state variables of SiB and their governing equations*

The SiB model has seven prognostic physical-state variables: two temperatures (one temperature for the canopy vegetation, T_c, and one temperature for both the ground cover and the soil surface, T_{gs}); two interception water stores (one for the canopy, M_c, and one for the ground cover, M_g); and three soil moisture stores (W_1, W_2, and W_3).

The governing equations for the two temperatures, T_c and T_{gs}, are:

Canopy

$$C_c \frac{\partial T_c}{\partial t} = Rn_c - H_c - \lambda E_c, \tag{1}$$

Ground

$$C_{gs} \frac{\partial T_{gs}}{\partial t} = Rn_{gs} - H_{gs} - \lambda E_{gs}, \tag{2}$$

where

T_c, T_{gs} = temperature, K
Rn_c, Rn_{gs} = absorbed net radiation, W m^{-2}
H_c, H_{gs} = sensible heat flux, W m^{-2}
E_c, E_{gs} = evapotranspiration rate, kg m^{-2} s^{-1}
C_c, C_{gs} = heat capacity, J m^{-2} K^{-1}
λ = latent heat of vaporization, J kg^{-1}.

(a)

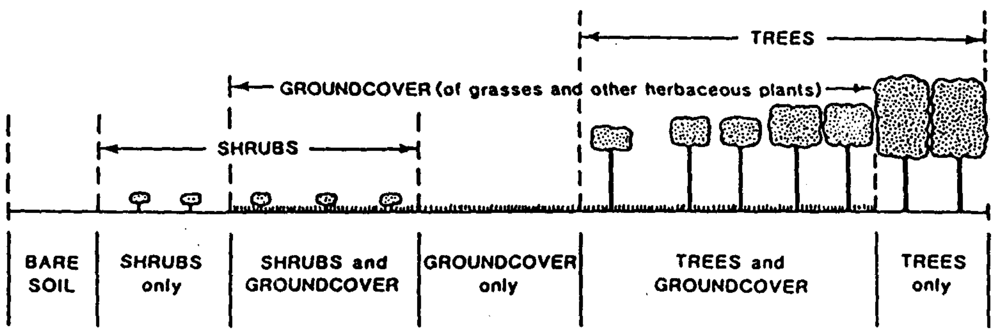

FIG. 1a. Vegetation morphology as represented in the Simple Biosphere (SiB).

The subscript c refers to the canopy, g to the ground cover, and s to the bare soil surface. The combined subscript gs refers to both the ground cover and bare soil surface.

The evapotranspiration from the canopy, E_c, has two components: (i) E_{wc}, evaporation of water from the wetted fraction of the canopy (canopy interception loss); and (ii) E_{dc}, transpiration of soil water extracted by the root system and lost from the dry fraction of the canopy.

The evapotranspiration from the ground cover and the surface soil layer E_{gs}, has three components: (i) E_{wg} and (ii) E_{dg}, which correspond to E_{wc} and E_{dc} for the ground cover; and (iii) E_s, direct evaporation of water from the surface soil layer.

The governing equations for the two interception water stores are

Canopy

$$\frac{\partial M_c}{\partial t} = P_c - D_c - E_{wc}/\rho_w, \tag{3}$$

(b)

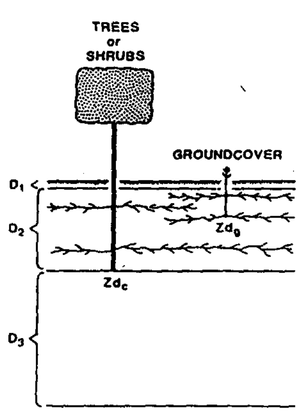

FIG. 1b. Root systems and soil layers of the Simple Biosphere (SiB). The parameters that define the vegetation and soil properties are listed in Table 1.

Ground Cover

$$\frac{\partial M_g}{\partial t} = P_g - D_g - E_{wg}/\rho_w, \tag{4}$$

where

M_c, M_g = water stored on the leaves, m
P_c, P_g = rate of precipitation interception, m s^{-1}
D_c, D_g = water drainage rate, m s^{-1}
E_{wc}, E_{wg} = rate of evaporation of water from the wet portions of the leaves, kg m^{-2} s^{-1}
ρ_w = density of water, kg m^{-3}.

The governing equations for the three soil moisture stores are

$$\frac{\partial W_1}{\partial t} = \frac{1}{\theta_s D_1}\left[P_1 - Q_{1,2} - \frac{1}{\rho_w}(E_s + E_{dc,1} + E_{dg,1})\right], \tag{5}$$

$$\frac{\partial W_2}{\partial t} = \frac{1}{\theta_s D_2}\left[Q_{1,2} - Q_{2,3} - \frac{1}{\rho_w}(E_{dc,2} + E_{dg,2})\right], \tag{6}$$

$$\frac{\partial W_3}{\partial t} = \frac{1}{\theta_s D_3}[Q_{2,3} - Q_3], \tag{7}$$

where

W_1, W_2, W_3 = soil moisture wetness in the three soil layers
$= \theta_i/\theta_s$
θ_i = volumetric soil moisture in layer i, m^3 m^{-3}
θ_s = value of θ_i at saturation, m^3 m^{-3}
D_i = thickness of ith soil layer, m
$Q_{i,i+1}$ = flow between i and $i+1$ soil layer, m s^{-1}
Q_3 = gravitational drainage from recharge soil moisture store, m s^{-1}
$E_{dc,i}$, $E_{dg,i}$ = canopy, ground cover abstraction of soil

TABLE 1. Parameters used by SiB. Subscript c refers to upper story or canopy vegetation; g to ground cover and s to soil. The combined subscript gs refers to ground cover and soil together.

Parameter	Definition
	a. Morphological parameters
V_c, V_g	Fractional area covered by the canopy, ground cover vegetation
$O(\xi, \theta)_c$, $O(\xi, \theta)_g$	Leaf angle distributions
z_2, z_1	Height of the canopy top, canopy bottom, m
Ld	Canopy leaf and stem area density (as a function of time), m^2 m^{-3}
L_{tg}	Ground cover leaf and stem area index (as a function of time), m^2 m^{-2}
z_{dc}, z_{dg}	Rooting depth (as a function of time), m
D_{dc}, D_{dg}	Root length density (as a function of time), m m^{-3}
D_1, D_2, D_3	Thickness of the three soil moisture storage layers, m
	b. Physiological parameters
N_c, N_g	Green leaf fractions of total leaf and stem area indices (as functions of time)
$(a, b, c)_c$, $(a, b, c)_g$	Light dependent stomatal resistance coefficients for a green leaf
$(T_l, T_h, T_0)_c$, $(T_l, T_h, T_0)_g$	Minimum, maximum and optimum temperatures for stomatal functioning, K
h_{5c}, h_{5g}	Parameter that governs the stomatal response of leaves to the atmospheric water vapor pressure, mb^{-1}
$(h_6, \psi_0)_c$, $(h_6, \psi_0)_g$	Parameters that govern the stomatal response of leaves to the leaf water potential, m^{-1}, m
$(r_{\rm plant})_c$, $(r_{\rm plant})_g$	Resistance imposed by plant vascular system, s
R_c, R_g	Root resistance per unit root length, s m^{-1}
	c. Physical parameters
$\alpha_{(\Lambda)c}$, $\alpha_{(\Lambda)g}$	Reflectances of leaves for wavelength interval Λ
$\delta_{(\Lambda)c}$, $\delta_{(\Lambda)g}$	Transmissivities of leaves for wavelength interval Λ
$\alpha_{s(\Lambda)}$	Soil reflectance for wavelength interval, Λ
Cd	Drag coefficient of a canopy leaf
P_s	Shelter factor of a canopy leaf
C_s	Heat/vapor transfer coefficient of a canopy leaf
Z_{gs} (or C_{Dgs})	Roughness length, m, (or drag coefficient) of the ground cover and soil surface (as a function of time)
C_c, C_{gs}	Heat capacity of the canopy and ground, J m^{-2} K^{-1}
θ_s	Soil pore space, m^3 m^{-3}
ψ_s	Soil moisture potential at saturation, m
K_s	Hydraulic conductivity of saturated soil, m s^{-1}
B	Parameter that relates volumetric soil moisture to soil moisture potential

moisture by transpiration from the ith soil layer, kg m^{-2} s^{-1}

P_1 = infiltration of precipitation into the upper soil moisture store, m s^{-1}

= minimum (P_0, K_s)

= 0, when $W_1 = 1$

K_s = hydraulic conductivity of saturated soil, m s^{-1}

P_0 = effective precipitation rate on soil surface, m s^{-1}

$= P - (P_c + P_g) + (D_c + D_g)$

P = precipitation rate above the canopy, m s^{-1}.

The precipitation excess, $(P_0 - P_1)$, joins the gravitational outflow from the lowest soil moisture store, Q_3, to produce runoff:

$$Q_r = P_0 - P_1 + Q_3 \tag{8}$$

where Q_r = runoff rate, m s^{-1}.

In the numerical solution of the prognostic equations for T_c and T_{gs}, we make use of the fact that the storage terms, involving C_c and C_{gs}, are small relative to the energy fluxes, Rn, H and λE. The heat capacity of the canopy, C_c, is of the order of the heat capacity of 0.2 mm of water per unit leaf area index. The ground (ground cover plus soil) heat capacity C_{gs} is defined by invoking the concept of a diurnal thermal skin depth and is taken to be thermally equivalent to 25–40 mm of water depending on soil wetness, following Arakawa (1972). These values make (1) and (2) "fast" response (or "quasi-diagnostic") equations so that changes in T_c and T_{gs}, even over a time step as short as an hour, can have a significant feedback on the magnitude of the energy fluxes.

The energy fluxes, Rn_c, Rn_{gs}, H_c, H_{gs}, λE_c, λE_{gs}, are explicit functions of (i) the atmospheric boundary conditions: T_r, e_r, u_r, $F_{\Lambda,\mu(0)}$ and P; (ii) the prognostic variables of SiB: T_c, T_{gs}, M_c, M_g, W_1, W_2, W_3; (iii) the three aerodynamic resistances: r_a, $\bar{r}_b$ and r_d; and (iv) the three surface resistances: $\bar{r}_c$, r_g, and $r_{\rm surf}$ (see Fig. 2). As will be shown later, these energy fluxes are used in an explicit backward-differencing scheme to calculate the changes in T_c and T_{gs} over a time step Δt. After that, the various moisture fluxes predicted by the model are used to update the moisture stores: M_c, M_g, W_1, W_2, W_3.

The resistance formulation allows one to describe the sensible and latent heat fluxes in the electrical analogue form, whereby

$$\text{flux} = \frac{\text{potential difference}}{\text{resistance}}.$$

The terms H_c, H_{gs}, and λE_c, λE_{gs} are determined in this way. In the case of the sensible heat fluxes, the potential differences are $(T_c - T_a)$ for the canopy and $(T_{gs} - T_a)$ for the ground, where T_a is the air temperature in the canopy air space. For the latent heat fluxes, λE_c and λE_{gs}, the corresponding vapor pressures are used (see Fig. 2 and Table 2). All the potential differences can then be expressed in terms of the tempera-

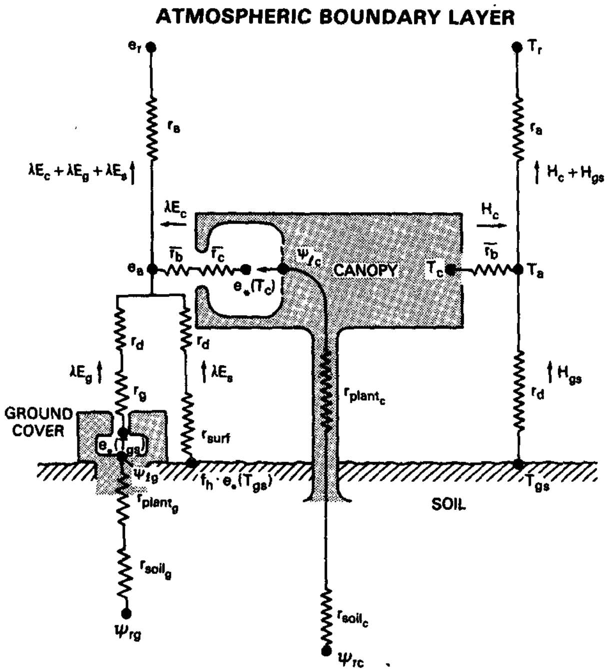

FIG. 2. Framework of the Simple Biosphere (SiB). The transfer pathways for latent and sensible heat flux are shown on the left- and right-hand sides of the diagram respectively. The treatment of radiation and intercepted water has been omitted for clarity. Symbols are defined in Table 2.

tures, T_c, T_{gs}; the moisture stores, M_c, M_g, W_1, W_2, W_3; and the atmospheric boundary layer reference temperature, T_r, and water vapor pressure, e_r, operating through the intermediate variables of T_a and e_a in the canopy air space.

The resistances are equivalent to the integrals of inverse conductances over a specified length. In the case of the aerodynamic resistances, r_a, $\bar{r}_b$, and r_d, the conductances correspond to the turbulent transfer coefficients for heat and water vapor. The surface resistances, $\bar{r}_c$, r_g and r_{surf}, are the additional resistances imposed on the transfer of water vapor from the saturated tissues within the leaves of the canopy and ground cover vegetation and from the water within the soil surface store, respectively. The three aerodynamic resistances, r_a, $\bar{r}_b$ and r_d, are derived from calculations involving the morphology of the vegetation and soil surface (see section 6), the wind speed at the reference height in the atmospheric boundary layer, u_r, and the temperature differences, $(T_c - T_a)$, $(T_{gs} - T_a)$ and $(T_a - T_r)$. The surface resistances, $\bar{r}_c$, r_g and r_{surf}, are calculated from a combination of plant physiological and morphological parameters, the state of the surface soil moisture store, the atmospheric boundary conditions and the values of the prognostic variables. These calculations and the role of the interception stores, M_c and M_g, are described in section 7. The flux of water by gravitational drainage and hydraulic diffusion in the soil is described in section 8.

The calculations described in sections 5 through 7 are used to reduce (1) and (2) to a pair of coupled differential equations in T_c and T_{gs}, for which solutions are obtained using the method described in section 9. Following this calculation, the other prognostic variables are advanced in time.

The prognostic variables and the atmospheric boundary conditions are listed in Table 2a, b. These are used to calculate the potential differences and resistances that determine the sensible and latent heat fluxes, which are listed in Table 2c.

5. Radiation fluxes

The interception, reflection, transmission and absorption of radiation by vegetation can be modeled in

TABLE 2. List of prognostic and forcing variables and variables associated with flux calculations in SiB.*

Variable	Definition
	a. Prognostic
T_c	Canopy temperature, K
T_{gs}	Ground temperature, K
M_c	Liquid water stored on canopy foliage, m
M_g	Liquid water stored on ground cover foliage, m
W_1	Wetness of surface store
W_2	Wetness of root zone
W_3	Wetness of recharge zone
	b. Atmospheric boundary conditions
u_r	Wind speed at reference height, m s^{-1}
T_r	Air temperature at reference height, K
e_r	Vapor pressure at reference height, mb
$F_{\lambda,\mu(0)}$	Incident radiative flux (spectral, Λ, and angular, μ, components), W m^{-2}
P	Precipitation, m s^{-1}

c. Fluxes

Flux	Potential difference	Resistances
H_c	$(T_c - T_a)\rho c_p$	$\bar{r}_b/2$
H_{gs}	$(T_{gs} - T_a)\rho c_p$	r_d
$H_c + H_{gs}$	$(T_a - T_r)\rho c_p$	r_a
λE_c	$(e_*(T_c) - e_a)\rho c_p/\gamma$	$f(\bar{r}_c, \bar{r}_b, M_c)$
λE_g	$(e_*(T_{gs}) - e_a)\rho c_p/\gamma$	$f(r_g, r_d, M_g)$
λE_s	$(f_h \cdot e_*(T_{gs}) - e_a)\rho c_p/\gamma$	$f(r_{surf}, r_d)$
$\lambda E_c + \lambda E_g + \lambda E_s$	$(e_a - e_r)\rho c_p/\gamma$	r_a

* T_a, e_a = air temperature, vapor pressure in canopy air space, K, mb;
ρ, c_p = density, specific heat of air, kg m^{-3}, J kg^{-1} K^{-1};
γ = psychrometric constant, mb K^{-1};
$\bar{r}_b$ = bulk boundary layer resistance, s m^{-1};
r_d = aerodynamic resistance between ground and canopy air space, s m^{-1};
r_a = aerodynamic resistance between canopy air space and reference height, s m^{-1};
$\bar{r}_c$ = bulk stomatal resistance of upper story vegetation, s m^{-1};
r_g = bulk stomatal resistance of ground vegetation, s m^{-1};
r_{surf} = bare soil surface resistance, s m^{-1};
f_h = relative humidity within pore space of surface soil layer;
$e_*(T)$ = saturation vapor pressure at temperature T, mb.

various ways (see the review of Dickinson, 1983), but a simple, economical method is appropriate to SiB.

The two-stream approximation, as described by Meador and Weaver (1980), Coakley and Chylek (1975), and Dickinson (1983), allows for the multiple reflection of light by leaves, an effect which can make a dense canopy an effective radiation trap. Here, we follow the proposal of Dickinson (1983) and let

$$-\bar{\mu}\frac{dI\uparrow}{dL} + [1-(1-\beta)\omega]I\uparrow - \omega\beta I\downarrow = \omega\bar{\mu}K\beta_0 e^{-KL}, \quad (9)$$

$$\bar{\mu}\frac{dI\downarrow}{dL} + [1-(1-\beta)\omega]I\downarrow - \omega\beta I\uparrow = \omega\bar{\mu}K(1-\beta_0)e^{-KL}, \quad (10)$$

where

- $I\uparrow, I\downarrow$ = upward and downward diffuse radiative fluxes, respectively, normalized by the incident flux
- μ = cosine of the zenith angle of the incident beam
- K = optical depth of the direct beam per unit leaf area = $G(\mu)/\mu$
- $G(\mu)$ = relative projected area of leaf elements in direction μ
- $\bar{\mu}$ = average inverse diffuse optical depth per unit leaf area
- μ' = direction of scattered flux
- β, β_0 = upscatter parameters for the diffuse and direct beams
- ω = scattering coefficient of phyto-elements
- L = cumulative leaf area index, $m^2\ m^{-2}$.

The value of $\omega\beta$ is obtained from the analysis of Norman and Jarvis (1975),

$$\omega\beta = \frac{1}{2}[\alpha + \delta + (\alpha - \delta)(\cos^2\bar{\theta})] \quad (11)$$

where

- $\omega = \alpha + \delta$
- α = leaf reflection coefficient
- δ = leaf transmission coefficient
- $\bar{\theta}$ = mean leaf inclination angle above the horizontal plane.

Following Dickinson (1983), the direct beam upscatter parameter is taken to be

$$\beta_0 = \frac{(1+\bar{\mu}K)}{\omega\bar{\mu}K}a_{s(\mu)}, \quad (12)$$

where the single scattering albedo, $a_{s(\mu)}$ is given by

$$a_{s(\mu)} = \omega\int_0^1 \frac{\mu'\Gamma(\mu,\mu')}{[\mu G(\mu') + \mu' G(\mu)]}d\mu', \quad (13)$$

here $\Gamma(\mu,\mu') = G(\mu)G(\mu')P(\mu,\mu')$, $P(\mu,\mu')$ = scattering phase function. The above analysis borrows heavily from the review of Dickinson (1983) in which the physical significance of the various parameters is discussed in some detail.

A number of simplifying assumptions are made to permit the calculation of some of the above parameters. First, isotropic scattering is assumed. This allows us to write the scattering phase function as

$$P(\mu,\mu') \propto \frac{1}{G(\mu')},$$

which is independent of the angle of the incoming beam. To satisfy the normalizing expression

$$\int_{-1}^{1} P(\mu,\mu')G(\mu')d\mu' = 1,$$

$P(\mu,\mu')$ must be equal to $[2G(\mu')]^{-1}$. This gives us

$$\Gamma(\mu,\mu') = \frac{1}{2}G(\mu'). \quad (14)$$

Using (14), it is a relatively simple task to obtain expressions for $a_{s(\mu)}$ for vertical, horizontal and spherically distributed leaves (see Sellers, 1985). For canopies with irregular leaf-angle distributions, the Ross (1975) χ_L function, which characterizes the deviation of the leaf angle distribution away from the spherical distribution, may be used to define $G(\mu')$ and $G(\mu)$ in (13). The full forms of the $a_{s(\mu)}$ functions for a range of canopy architectures may be found in Sellers (1985). The value of the average inverse diffuse optical depth per unit leaf area, $\bar{\mu}$, is obtained from

$$\bar{\mu} = \int_0^1 \frac{\mu'}{G(\mu')}d\mu'. \quad (15)$$

Using the above definitions and suitable boundary conditions, (9) and (10) can be solved to obtain estimates of the reflectance and transmission coefficients of the canopy. For example, if all of the incoming radiation is direct beam radiation, the boundary conditions will be (i) no downward diffuse radiation at the top of the vegetation

$$I\downarrow = 0, \quad \text{at} \quad L = 0, \quad (16a)$$

and (ii) reflectance below the vegetation

$$I\uparrow = \alpha_d I\downarrow + \alpha_b e^{-KL_t}, \quad \text{at} \quad L = L_t, \quad (16b)$$

where

- $\alpha_{b,d}$ = reflectance of underlying surface for direct beam, diffuse radiation
- L_t = local leaf area index of the vegetation.

The solutions to (9) and (10) are then

$$\left.\begin{aligned} I\downarrow &= \alpha_1 e^{-KL} + \alpha_2 e^{-\alpha_3 L} + \alpha_4 e^{\alpha_3 L} \\ I\uparrow &= \alpha_5 e^{-KL} + \alpha_6 e^{-\alpha_3 L} + \alpha_7 e^{\alpha_3 L} \end{aligned}\right\}, \quad (17)$$

where $\alpha_1, \alpha_2, \ldots \alpha_7$ are algebraic combinations of the coefficients in Eqs. (9) and (10) (see Sellers, 1985); $I\uparrow(0)$ may then be taken as the spectral albedo, and $I\downarrow(L_t)$

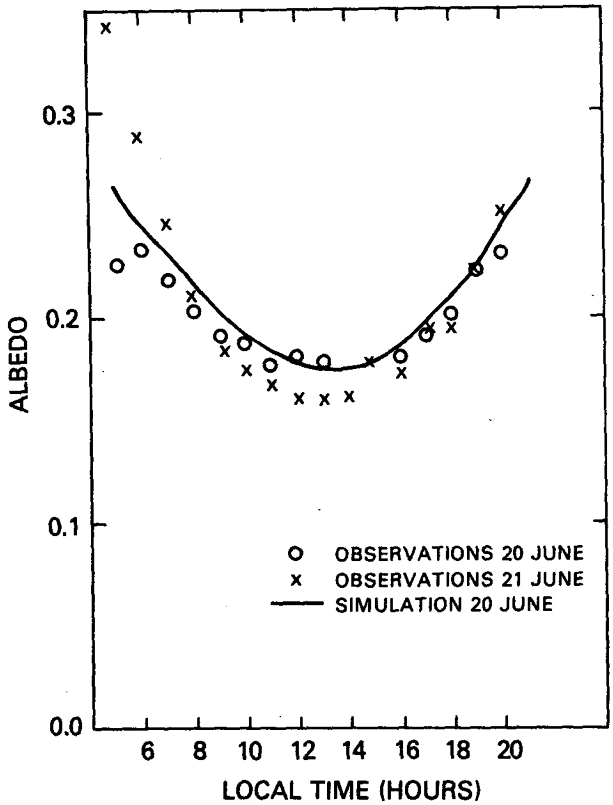

FIG. 3. Measured and simulated albedo values for shortwave radiation above a wheat crop. The data points were calculated from observations of incoming and outgoing shortwave radiation taken at Volkenrude, W. Germany, 20–21 June 1979 [see van der Ploeg et al. (1980)]. The simulation for the wheat crop at Ruthe (50 km away) for 20 June is also shown.

$+ e^{-KL_t}$ as the spectral transmittance of the vegetation for direct beam radiation. Figure 3 shows calculated global albedo values for a young wheat crop on a clear day in West Germany compared to two days of data obtained from van der Ploeg et al. (1980). For this trial, the data of Dickinson (1983) and Goudriaan (1977) were used to specify the values of the leaf transmittances and reflectances for the visible (<0.7 μm) and near-infrared (>0.7 μm) wavelength intervals. The incident solar radiation was split into visible and near-infrared direct and diffuse components according to the scheme of Goudriaan (1977; Fig. 1, p. 11). Both the magnitude of the global albedo (obtained by summing diffuse and direct components of the visible and near infrared reflectances) and its diurnal variation agree closely with the data.

Care must be taken to ensure that the local leaf area index L_t used in the above calculation, is related to the area-averaged leaf area index $\overline{L_t}$ by $L_t = \overline{L_t}/V$, where V_c, V_g represent the fractional cover of the canopy vegetation and ground cover. Also, mean optical properties of the phyto-elements must be used within each wavelength interval. Currently, these are determined by summing the weighted values for the live and dead leaves and the nonphotosynthesizing material.

The terms describing the effects of the direct incident flux on the canopy on the right-hand sides of (9) and (10) may be omitted, and suitable boundary conditions ($I\downarrow = 1$, at $L = 0$; $I\uparrow = \alpha_d I\downarrow$, at $L = L_t$) applied to carry out similar computations for the partition of the incident diffuse radiation (see Meador and Weaver, 1980). The full forms of the expressions for the single scattering albedos and the reflectance and transmission of radiation in plant canopies using the above scheme are summarized in a paper by Sellers (1985) for a number of leaf geometries.

The equation set (9)–(15) is used to describe the absorption of the different components of the incident radiation by the canopy and ground. First, the model is used to determine the spectral reflectance of the ground cover for diffuse and direct radiation so that the spectral reflectance of the ground, A, may be written generally as

$$A_{\Lambda,\mu} = I\uparrow_g V_g + (1 - V_g)\alpha_{s(\Lambda)}, \quad (18)$$

where

$A_{\Lambda,\mu}$ = ground spectral reflectance as a function of wavelength interval, Λ, and angle of incident radiation, μ

$\alpha_{s(\Lambda)}$ = soil spectral reflectance (assumed to be isotropic)

$I\uparrow_g$ = upward diffuse radiation flux above ground cover, normalized by the incident flux.

The value of $A_{\Lambda,\mu}$ must be determined for diffuse and direct fluxes to yield $A_{\Lambda,d}$ and $A_{\Lambda,b}$, respectively.

The radiation absorbed by the canopy and ground is then given by

$$F_{\Lambda,\mu(c)} = V_c[1 - I\uparrow_c - I\downarrow^c_{gs}(1 - A_{\Lambda,d}) - e^{-KL_{tc}}(1 - A_{\Lambda,b})]F_{\Lambda,\mu(0)}, \quad (19)$$

and

$$F_{\Lambda,\mu(gs)} = \{(1 - V_c)(1 - A_{\Lambda,\mu}) + V_c[I\downarrow^c_{gs}(1 - A_{\Lambda,d}) + e^{-KL_{tc}}(1 - A_{\Lambda,b})]\}F_{\Lambda,\mu(0)}, \quad (20)$$

where

$F_{\Lambda,\mu(0)}$ = incident radiant energy of wavelength interval Λ and direction μ, W m^{-2}

$F_{\Lambda,\mu(c)}$, $F_{\Lambda,\mu(gs)}$ = amount of $F_{\Lambda,\mu(0)}$ absorbed by the canopy and by the ground cover and soil, W m^{-2}

$I\uparrow_c$ = diffuse flux leaving top of canopy

$I\downarrow^c_{gs}$ = diffuse flux leaving base of canopy

$e^{-KL_{tc}}$ = direct flux penetrating the canopy = 0, when $\mu = d$

L_{tc} = local leaf area index of canopy, m^2 m^{-2}.

[NB: The diffuse terms in (19) and (20) are normalized by the incident flux, $F_{\Lambda,\mu(0)}$. Also, $A_{\Lambda,b}$ will vary with solar angle, while $A_{\Lambda,d}$ is constant for a given surface condition.]

The above calculations are carried out for the five components of incoming radiation: visible (diffuse and direct), near infrared (diffuse and direct), and thermal infrared (diffuse only).

Finally, the net radiation fluxes are obtained by subtracting the emitted thermal radiation from the absorbed radiation

$$Rn_c = \langle F_c \rangle - 2\sigma_s T_c^4 \cdot V_c\delta_t + \sigma_s T_{gs}^4 V_c\delta_t, \quad (21)$$

$$Rn_{gs} = \langle F_{gs} \rangle - \sigma_s T_{gs}^4 + \sigma_s T_c^4 V_c\delta_t, \quad (22)$$

where

$\langle F_c \rangle$, $\langle F_{gs} \rangle$ = sum of the five absorbed radiation components for the canopy, ground

σ_s = Stefan-Boltzmann constant, W m^{-2} K^{-4}

δ_t = canopy transmittance for thermal infrared radiation $= 1 - e^{-L_{tc}/\bar{\mu}}$.

Currently, the reflection of thermal infrared radiation is neglected as the emissivity of most natural surfaces is close to unity.

The effects of snow on the surface radiation budget are discussed in appendix B.

6. Aerodynamic resistances

Figure 2 shows how the SiB model conceptualizes the transfer of heat and water vapor from the canopy and ground to the atmospheric boundary layer via three surface resistances and three aerodynamic resistances. The calculation of the aerodynamic resistances is described in this section.

The turbulent transport of momentum and other quantities between the atmosphere and the surface is a complex process that does not lend itself to simple modeling (see Raupach and Thom, 1981). In SiB, eddy diffusion concepts are used to describe all such transfers, with the knowledge that these concepts fall short of complete physical realism.

The morphological parameters of SiB, given in section 4, are used to solve diffusion equations which describe the absorption of momentum by the canopy and ground.

Figure 4a shows how the canopy is represented as a block of constant density porous material sandwiched in between two constant stress layers. Extrapolation of the log-linear profile has been shown to yield underestimates of the turbulent transfer coefficient close to the top of plant canopies; we therefore assume that the log-linear profile, often used to describe the variation of wind speed with height within the constant stress layer near the surface, is valid only above a certain transition height, z_m, which is taken to be a function of the canopy morphology and below which an empirical adjustment to the profile must be made. Data from Garratt (1978) and Raupach and Thom (1981) are used to estimate the values of z_m and the ratio of the actual turbulent transfer coefficient at z_2 to that predicted from extrapolation of the log-linear profile.

Under neutral conditions, equations for the transfer of momentum above and within the canopy may be written as follows:

Above the canopy ($z > z_m$)

$$\tau = \rho u_*^2 = \rho \left[\frac{ku}{\ln\left(\frac{z-d}{z_0}\right)} \right]^2 \quad (23)$$

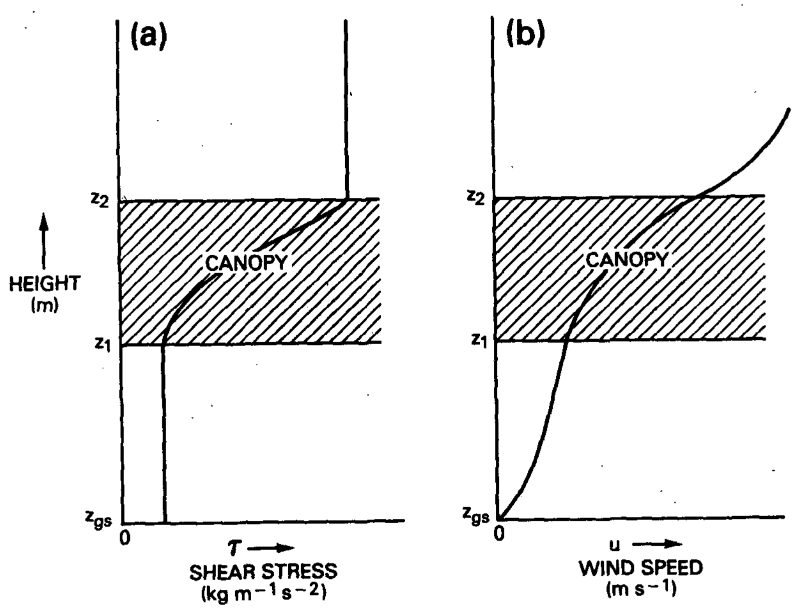

FIG. 4. Profiles of (a) shear stress and (b) wind speed above, within and below the canopy vegetation as represented by SiB.

where

τ = shear stress, kg m^{-1}·s^{-2}
ρ = air density, kg m^{-3}
u_* = friction velocity, m s^{-1}
k = von Karman's constant = 0.41
d = zero plane displacement height, m
z_0 = roughness length, m
z_m = transition height, m.

Within the canopy ($z_1 < z < z_2$)

$$\frac{\partial \tau}{\partial z} = \rho \frac{Cd\overline{Ld}}{P_s} u^2 \tag{24}$$

where

u = wind speed, m s^{-1}
Cd = leaf drag coefficient
$\overline{Ld}$ = area-averaged stem and leaf area density, m^2 m^{-3}
P_s = leaf shelter factor.

Also

$$\tau = \rho K_m \frac{\partial u}{\partial z}, \tag{25}$$

where

K_m = momentum transfer coefficient, m^2 s^{-1}

and

$$d = \frac{\int_{z_1}^{z_2} u^2 z \, dz}{\int_{z_1}^{z_2} u^2 dz + \frac{\tau}{\rho}\Big|_{z_1} \frac{P_s}{\overline{Ld}Cd}}. \tag{26}$$

The derivation of (23), (24) and (25) can be found in Monteith (1973); they are commonly used to describe the absorption of momentum by a rough surface. Equation (26) was first suggested by Thom (1971), where d, the zero plane displacement height, is defined as the moment height for momentum absorption. We have added the second term, $\tau/\rho|z_1(P_s/\overline{Ld}Cd)$, to the original form, to take account of the momentum absorbed by the ground. In contrast to the approach used for determining the radiation fluxes, we now describe the effect of the vegetation morphology on the turbulent fluxes in terms of the area-averaged (denoted by an overbar) properties of the vegetation.

The momentum flux conditions at z_2 must be related to those above the transition height, z_m, so that continuous profiles of K_m and u extend from above z_m to the soil surface. Garratt (1978) and Raupach and Thom (1981) have noted that estimates of the momentum flux coefficient at $z = z_2$ were 1.5–2.0 times larger than a simple downward extrapolation of Eq. (23) would indicate. The value of K_m was observed by Garratt (1978) to approach the log linear estimate at a transition height, z_m, which was found to be roughly equivalent to the sum of the obstacle height and spacing length. Accordingly, we have set $z_m \approx z_2 + 2(z_2 - d)$, as $(z_2 - d)$ is roughly equivalent to obstacle height.

To obtain the characteristics of the canopy, z_0 and d, and to estimate the resistances, $\overline{r_b}$, r_d and r_a, we close the equation set (23)–(26), by adding an equation that describes the variation of K_m within the canopy air space. This leaves us with five equations in the five unknowns, u, K_m, z_0, d, and $\tau/\rho|z_1$. We have mentioned before that the use of "K-theory" within the canopy may be physically unrealistic, but because it yields reasonable results we shall use this method until suitable second-order closure models can be applied to the problem.

A number of assumptions can be made about the variation of K_m within the canopy. These are:

- $K_m \propto u$ — The data of Denmead (1976) and of Legg and Long (1975) qualitatively support this relationship;
- $K_m = K_m(z_2)$ — Jarvis et al. (1976) found this to be representative of the upper story of a coniferous canopy;
- $K_m = l_m^2 \dfrac{\partial u}{\partial z}$ — where l_m is a mixing length. (27)

Lastly, boundary conditions must be imposed on the above. The upper boundary condition is $u = u_r$, $z = z_r$, while the lower one may be specified in terms of the shear stress at ground level, defined by

$$\tau|_{z_1} = \rho C_{D_{gs}} u^2|_{z_1} \tag{28a}$$

and

$$\tau|_{z_1} = K_m \frac{\partial u}{\partial z}\Big|_{z_1}; \tag{28b}$$

$C_{D_{gs}}$, the drag coefficient, can be estimated from the size of the roughness elements on the ground.

Equations (23)–(28) may now be solved for any given set of values of z_2, z_1, $\overline{Ld}$, Cd, P_s and $C_{D_{gs}}$. Some experimentation indicates that the use of $K_m \propto u$ yields the best results. The parameters that are hardest to quantify are the shelter factor, P_s, reported to be between 1 and 4, depending on vegetation density, and the ratio of K_m at z_2 to the extrapolated log-linear value, taken here as equal to 2.0. The solution of the equation set (see Appendix A) yields estimates of z_0 and d and profiles of u and K_m. The calculated dependence of z_0 and d on the leaf area index, $\overline{Ld}(z_2 - z_1)$, is shown in Figs. 5 and 6. These results are in qualitative agreement with those of Shaw and Pereira (1982, Figs. 4 and 5) which were obtained with a second-order closure model. The calculated roughness length, z_0 (plotted as a function of the difference between the height of the canopy top, z_2, and the zero plane displacement, d), is shown in Fig. 7. It, too, follows the trend calculated by Shaw and Pereira (1982, Fig. 6).

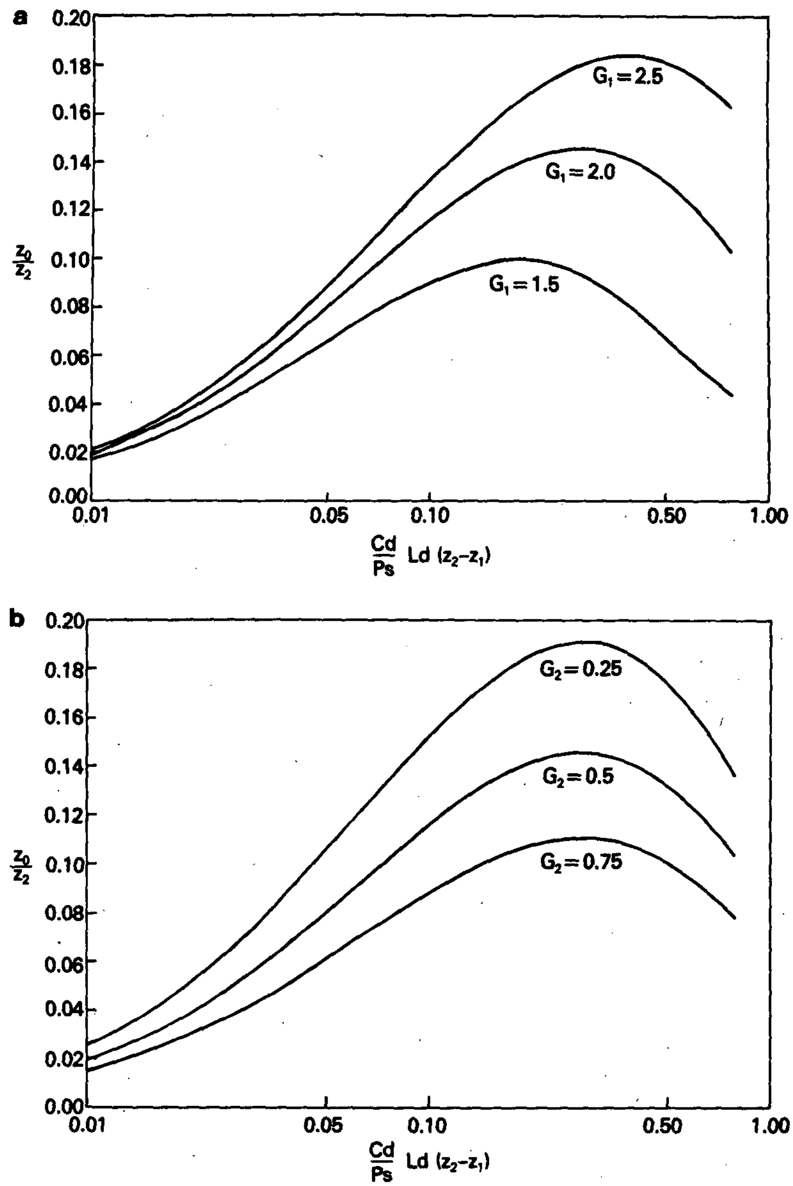

FIG. 5. Calculated SiB values of the roughness length, z_0, as a function of the leaf drag coefficient, leaf area index and shelter factor; z_0 is normalized by canopy height, z_2. Panel (a) shows the variation in z_0 with G_1, the ratio of the assumed value of K_m at z_2 to the values expected by extrapolation of the wind log profile (G_2 is held constant at 0.5). Panel (b) shows the variation in z_0 with G_2, the ratio of the assumed aerodynamic resistance between z_2 and the transition height, z_m, and the value expected from the wind log profile. (Here G_1 is held constant at 2.0.) Values of G_1 = 2.0 and G_2 = 0.5 are currently used in SiB. The horizontal axes of the above figures correspond directly with those in Figs. 4 and 5 of Shaw and Pereira (1982).

The profiles of u and K_m can now be used to obtain the aerodynamic resistances r_a, $\overline{r_b}$, and r_d. (The complete integrated solutions are given in appendix A.)

First, $\overline{r_b}$: the boundary layer resistance for a single leaf has been determined experimentally for many species (Goudriaan; 1977) and commonly yields an expression of the form

$$r_{b_i} = \frac{C_s}{L_i\sqrt{u_i}}, \qquad (29)$$

where

C_s = transfer coefficient
L_i = leaf area of ith leaf, m^2
u_i = local wind speed, m s^{-1}
r_{b_i} = boundary layer resistance of ith leaf, s m^{-1}.

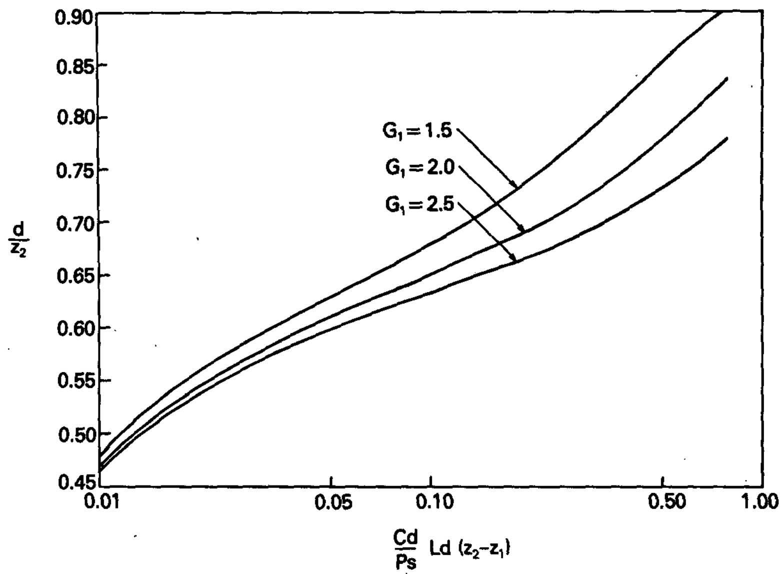

FIG. 6. Calculated (SiB) values of the zero plane displacement height, d, as a function of the leaf drag coefficient, leaf area index and shelter factor; d is normalized by the canopy height, z_2. Variation of d for three values of G_1 (see Fig. 5) is shown (d is independent of G_2). The horizontal axes correspond directly with those in Figs. 4 and 5 of Shaw and Pereira (1982).

A bulk boundary layer resistance may be assigned to a group of leaves if the individual resistances, r_{b_i}, are assumed to act in parallel:

$$\frac{1}{\bar{r}_b} = \sum_{i=1}^{n} \frac{L_i\sqrt{u_i}}{C_s P_s}, \tag{30}$$

where

$\bar{r}_b$ = area-averaged bulk boundary layer resistance, s m^{-1}
n = number of leaves
P_s = shelter factor.

Because we assume that the canopy leaf area density is constant with height (in SiB), and the variation of u with height in the canopy is known (from the previous set of calculations), we can write

$$\frac{1}{\bar{r}_b} = \int_0^{\bar{L}_t} \frac{\sqrt{u}}{C_s P_s} dL,$$

where

$\bar{L}_t$ = area-averaged canopy leaf and stem area index, m^2 m^{-2},

$$= \int_{z_1}^{z_2} \overline{Ld}\, dz,$$

so that

$$\frac{1}{\bar{r}_b} = \int_{z_1}^{z_2} \frac{\overline{Ld}\sqrt{u_{(z)}}}{P_s C_s} dz. \tag{31}$$

For neutral conditions (31) may be integrated to yield

$$\bar{r}_b = \frac{C_1}{\sqrt{u_2}}, \tag{32}$$

where

u_2 = wind speed at z_2, m s^{-1}
C_1 = surface dependent constant, obtained by integrating (31).

The water vapor and sensible heat source height h_a may be defined as the center of action of $\bar{r}_b$ in the canopy (see appendix A). The neutral value of the aerodynamic resistance to the transfer of heat and water vapor from the ground cover and the soil surface to h_a, may be written as

$$r_d = \int_0^{h_a} \frac{1}{K_s} dz = \frac{C_2}{u_2}, \tag{33}$$

where

C_2 = surface dependent constant.

The aerodynamic resistance between h_a and z_r, may be defined as

$$r_a = \int_{h_a}^{z_r} \frac{1}{K_s} dz = \int_{h_a}^{z_2} \frac{1}{K_s} dz + \int_{z_2}^{z_r} \frac{1}{K_s} dz, \tag{34}$$

where

K_s = heat/water vapor transfer coefficient, m^2 s^{-1}.

Under neutral conditions, the previous calculation of z_0, d and $u_{(z)}$ allows us to integrate the terms on the right-hand side of (34) and write

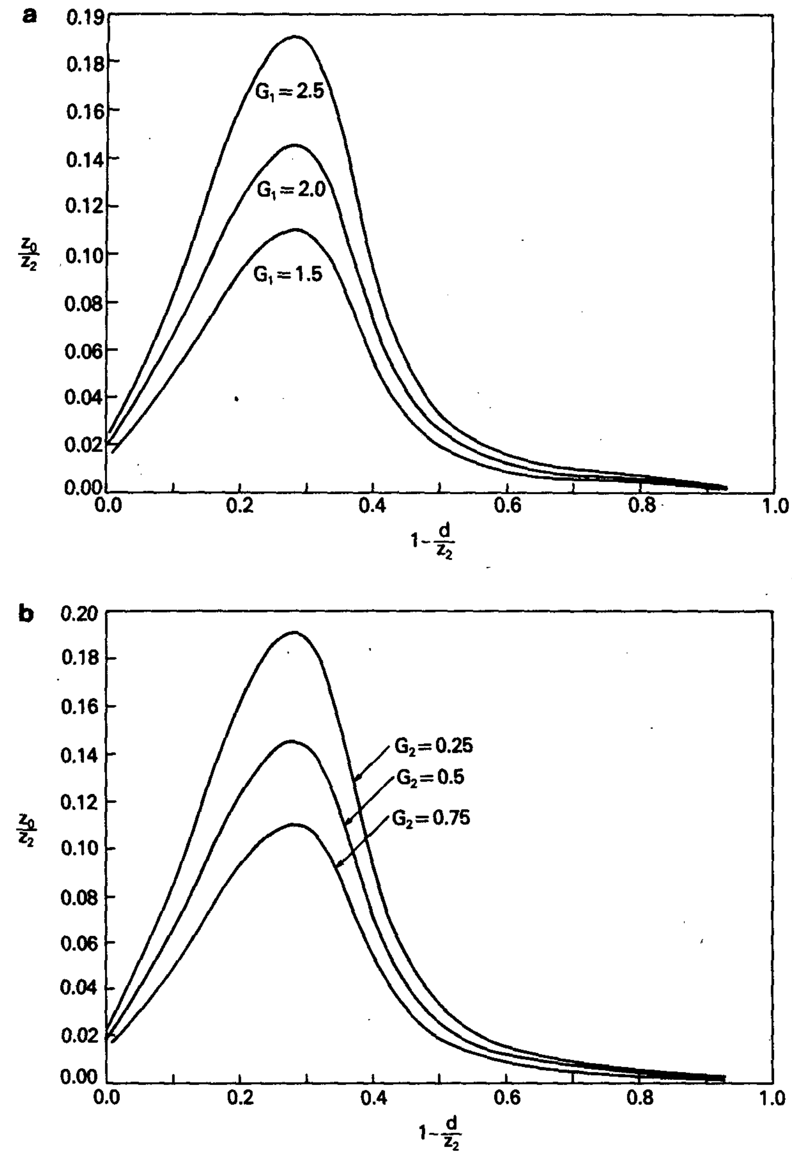

FIG. 7. Calculated SiB values of the roughness length, z_0, plotted against the calculated height between zero plane displacement, d, and the canopy top. Both quantities are normalized by canopy height, z_2. Panel (a) shows the variation in the relationship for three values of G_1, with G_2 constant at 0.5; while Panel (b) shows the same function for three values of G_2, with G_1 constant at 2.0. In all cases, z_0 is calculated to peak at intermediate leaf area indices. Axes correspond to those in Fig. 6 of Shaw and Pereira (1982).

$$r_a = \frac{C_3}{u_r}, \tag{35}$$

where

C_3 = surface dependent constant
u_r = wind speed at the reference height, m s^{-1}.

The above expressions for $\overline{r_b}$, r_d and r_a can be modified to take account of the effects of nonneutrality and, in the case of $\overline{r_b}$ and r_d, u_2 replaced by u_r. The relationship between u_2 and u_r and the full forms of the nonneutral resistances are given in appendix A. They yield:

$$\overline{r_b} = f_1\left(\frac{C_1}{\sqrt{u_r}}, T_c, T_a\right), \tag{36a}$$

$$r_d = f_2\left(\frac{C_2}{u_r}, T_{gs}, T_a\right), \tag{36b}$$

$$r_a = f_3\left(\frac{C_3}{u_r}, T_a, T_r\right), \tag{36c}$$

$$u_2 = f_4(u_r, T_a, T_r), \tag{36d}$$

where f_1–f_4 represent nonneutral modifications to the neutral estimates of the resistances and u_2.

The coefficients C_1, C_2 and C_3 need to be derived only once. A preprocessor program performs the integrations for a given surface configuration and the resultant coefficients are used for as long as the vegetation morphology remains the same. Crude modifications to the aerodynamic parameters are made if the depth of accumulated snow becomes significant compared to the height of the vegetation (see appendix B).

7. Surface resistances and water vapor fluxes

a. Transpiration and soil surface evaporation

The resistances to the transport of water vapor from within the canopy, ground cover and upper soil layer to the adjacent exterior air are defined as the canopy resistance, $\bar{r}_c$, ground cover resistance, r_g, and soil surface resistance, $r_{\rm surf}$, respectively. When the surface of the canopy is dry, $\bar{r}_c$ is assumed to be equal to the effect of all the canopy leaves' stomata acting in parallel. Below the canopy, the resistance to water vapor transfer is assumed to be a combination of the bulk stomatal resistance of the ground cover, r_g, and the diffusion resistance of the soil, $r_{\rm surf}$, acting in parallel.

Jarvis (1976) summarized his own and other researchers' work on the stomatal function of coniferous trees. The stomatal resistance of an individual green leaf, r_s, was taken to be a function of the normal incident visible or photosynthetically active radiation (PAR) flux density, the atmospheric water vapor pressure deficit, $f(T_a, e_a)$, the leaf temperature, T, and the leaf water potential, ψ_l.

Jarvis (1976) proposed the following expression for r_s:

$$r_s = [h_1 + h_2(Q_p - Q_m)]/h_1h_2(Q_p - Q_m) \times f(\psi_l)^{-1}f(T)^{-1}f(T_a, e_a)^{-1} \tag{37}$$

where

Q_p = photon flux density, μEin m^{-2} s^{-1}

$Q_m = \dfrac{1}{h_2}\dfrac{1}{r_{s_{\max}}}$, μEin m^{-2} s^{-1}

$r_{s_{\max}}$ = maximum value of stomatal resistance (i.e., when $Q_p = 0$), s m^{-1}

h_1 = maximal conductance as Q_p approaches infinity, m s^{-1} = $1/r_{s_{\min}}$

$r_{s_{\min}}$ = minimum stomatal resistance, s m^{-1}

$h_2 = \dfrac{\partial(1/r_s)}{\partial Q_p}$ when $Q_p = 0$

$f(\psi_l)$, $f(T)$, $f(T_a, e_a)$ = adjustment factors for the influence of leaf water potential, ψ_l, leaf temperature, T, and atmospheric water vapor pressure deficit, $[e^*(T_a) - e_a]$. The factors are limited to the range 0 to 1.

Rearrangement of the above yields

$$r_s = \left[\frac{a}{b + F_{s,l}} + c\right] f(\psi_l)^{-1}f(T)^{-1}f(T_a, e_a)^{-1} \tag{38}$$

where

a, b, c = constants determined from h_1, h_2 and Q_m in (37), J m^{-3}, W m^{-2}, s m^{-1}

$F_{s,l}$ = PAR flux incident on leaf surface, W m^{-2} (0.4–0.72 μ wavelength interval).

Following a similar methodology to that used for the derivation of $\bar{r}_b$, in (31), we define the bulk, area-averaged, stomatal resistance, $\bar{r}_c$, for the canopy by

$$\frac{1}{\bar{r}_c} = V_cN_c \int_0^{L_{tc}} \int_0^{\pi/2} \int_0^{2\pi} \frac{O(\xi, \theta)}{r_s(F_{s,\mu}, \xi, \theta)} \times \frac{\sin\theta}{f(\Sigma)^{-1}}\, d\xi d\theta dL \tag{39}$$

where

$O(\xi, \theta)$ = leaf angle distribution function

ξ, θ = leaf azimuth, inclination

N_c = fraction of L_{tc} that consists of live photosynthesizing leaves

$f(\Sigma)$ = product of all stress terms $= f(\psi_l)f(T)f(T_a, e_a)$.

The extinction of visible, photosynthetically active radiation (PAR) down through the canopy has been described by (9). Due to the small amount of scattering by phyto-elements in this wavelength interval ($\omega \leq 0.2$), we may also define the extinction of PAR by means of Goudriaan's (1977) semi-empirical expression to yield a more manipulable function

$$F_{s,\mu(L)} \approx F_{s,\mu(0)} \exp(-\kappa L) \tag{40}$$

where

κ = extinction coefficient $= \dfrac{G(\mu)}{\mu}(1 - \omega_s)^{1/2}$

$F_{s,\mu(0)}$ = downward PAR flux above canopy, W m^{-2}

$F_{s,\mu(L)}$ = downward PAR flux within canopy below a leaf area index of L, W m^{-2}

ω_s = scattering coefficient for PAR.

Using (38) and (40), we can expand equation (39) to give

$$\frac{1}{\bar{r}_c} = V_cN_c \int_0^{L_{tc}} \int_0^{\pi/2} \int_0^{2\pi} \frac{O(\xi, \theta)f(\Sigma)}{r_s(F_{s,\mu(0)}, \kappa, \xi, \theta)} \sin\theta\, d\xi d\theta dL. \tag{41}$$

Because the leaf water potential, leaf temperature and air vapor pressure within the canopy can be assumed to vary only slightly when compared to the extinction of radiation, $\bar{r}_c$ may be simplified to

$$\frac{1}{\overline{r_c}} \approx V_c N_c f(\Sigma) \int_0^{L_{tc}} \int_0^{\pi/2} \int_0^{2\pi} \frac{O(\xi, \theta)}{r_s(F_{s,\mu(0)}, \kappa, \xi, \theta)} \times \sin\theta \, d\xi d\theta dL. \quad (42)$$

Some analytical solutions to the integral part of the above equation have been obtained for a number of leaf angle distribution functions. These may be found in Sellers (1985) together with comparisons of the calculated values of $\overline{r_c}$ with field measurements.

The calculation of $\overline{r_c}$ in (42) requires the specification of a single downward PAR flux, $F_{s,\mu(0)}$, above the canopy. In SiB, the magnitude of this flux is obtained by summing the direct and diffuse PAR components and the mean value of μ is estimated from a calculation involving the incident angles of the diffuse and direct radiation weighted by their respective intensities. The above procedure may be used to obtain the bulk stomatal resistances for the canopy and the ground cover. The forms of the factors used to account for the effects of leaf water potential, vapor pressure deficit and leaf temperature are essentially the same as in Jarvis (1976):

$$\left.\begin{aligned} f(T) &= h_3(T - T_l)(T_h - T)^{h_4} \\ h_3 &= 1/(T_o - T_l)(T_h - T_o)^{h_4} \\ h_4 &= (T_h - T_o)/(T_o - T_l) \end{aligned}\right\}, \quad (43)$$

where

T = leaf temperature, (T_c or T_{gs}), K
T_o = optimum temperature, [$f(T) = 1$], K
T_h = upper temperature limit, [$f(T) = 0$], K
T_l = lower temperature limit, [$f(T) = 0$], K

$$f(T_a, e_a) = 1 - h_5[e_*(T_a) - e_a], \quad (44)$$

where

h_5 = species dependent constant, mb^{-1}
$e_*(T_a)$ = saturation water vapor pressure at canopy air space temperature, T_a, mb
e_a = vapor pressure in canopy air space, mb

$$\left.\begin{aligned} f(\psi_l) &= 1 - e^{-h_6\delta\psi} \\ \delta\psi &= \psi_l - \psi_0 \end{aligned}\right\}, \quad (45)$$

where

ψ_0 = leaf water potential at which stomata close completely, m
h_6 = species dependent constant, m^{-1}.

Of particular importance is the factor $f(\psi_l)$ which accounts for the effects of soil moisture stress and excessive evaporation demand. The leaf water potential, ψ_l, is calculated by using a catenary model of the water transfer pathway from root zone to leaf, following van der Honert (1948),

$$\psi_l = \psi_r - z_T - \frac{E_d}{\rho_w}(\overline{r_{\mathrm{plant}}} + \overline{r_{\mathrm{soil}}}), \quad (46)$$

where

ψ_l = leaf water potential, m
z_T = height of transpiration source, m
= h_a for the canopy vegetation
= 0 for the ground cover
ψ_r = soil moisture potential in root zone, m
$\overline{r_{\mathrm{plant}}}$ = area-averaged resistance imposed by the plant vascular system, s
$\overline{r_{\mathrm{soil}}}$ = area-averaged resistance of the soil and root system, s
ρ_w = density of water, kg m^{-3}
E_d = transpiration rate, kg $\mathrm{m}^{-2}\,\mathrm{s}^{-1}$.

The soil moisture potential in the root zone, ψ_r, is an average term obtained by summing the weighted soil moisture potentials of the soil layers from the surface to the rooting depth, z_d,

$$\psi_r = \frac{1}{z_d}\sum_0^{z_d} \psi_i D_i \quad (47)$$

where

ψ_i = soil moisture potential of the ith soil layer, m
D_i = depth of the ith soil layer, m
z_d = rooting depth, m.

The soil moisture potential of a layer, ψ_i, is taken from the empirical relationship of Clapp and Hornberger (1978),

$$\psi_i = \psi_s W_i^{-B} \quad (48)$$

where

W_i = soil moisture wetness in the ith soil layer = (θ_i/θ_s)
ψ_s = soil moisture potential at saturation, m
B = empirical constant.

The area-averaged resistance to the flow of water from the pores of the soil to the root cortex, $\overline{r_{\mathrm{soil}}}$, is described by a depth-averaged form of the relationship proposed by Federer (1979),

$$\overline{r_{\mathrm{soil}}} = (R/D_d + \alpha_f/K_r)/z_d, \quad (49)$$

where

$$\alpha_f = \frac{1}{8\pi D_d}\left[V_r - 3 - 2\ln\left(\frac{V_r}{1 - V_r}\right)\right],$$

and

R = resistance per unit root length, s m^{-1}
D_d = root density, m m^{-3}
V_r = volume of root per unit volume of soil, $\mathrm{m}^3\,\mathrm{m}^{-3}$
K_r = mean soil hydraulic conductivity in the root zone, m s^{-1}.

The soil hydraulic conductivity in the root zone, K_r, is obtained by a manipulation of the expressions of Clapp and Hornberger (1978) and of Milly and Eagleson (1982), which yields K_r as a function of ψ_r,

$$K_r = K_s\left(\frac{\psi_s}{\psi_r}\right)^{(2B+3)/B}, \tag{50}$$

where

K_s = saturated hydraulic conductivity in m s^{-1}.

Values of B, ψ_s and K_s for different soils are listed in Clapp and Hornberger (1978). The transpiration rates from the dry fractions of the vegetation are then given by

$$\lambda E_{d_c} = \frac{[e_*(T_c) - e_a]}{\bar{r}_c + \bar{r}_b}\frac{\rho c_p}{\gamma}(1 - W_c), \tag{51}$$

$$\lambda E_{d_g} = \frac{(e_*(T_{gs}) - e_a)}{r_g + r_d}\frac{\rho c_p}{\gamma}(1 - W_g)V_g, \tag{52}$$

where

E_{d_c}, E_{d_g} = transpiration rates of canopy vegetation and ground cover, m s^{-1}
$e_*(T_c)$, $e_*(T_{gs})$ = saturation vapor pressure at temperatures T_c, T_{gs}, mb
ρ, c_p = density, specific heat of air, kg m^{-3}, J kg^{-1} K^{-1}
γ = psychrometric constant, mb K^{-1}
λ = latent heat of vaporization, J kg^{-1}
W_c, W_g = wetness fraction of canopy and ground cover
= 1 when $e_*(T_c)$, $e_*(T_{gs}) < e_a$.

When $e_*(T_{gs})$ or $e_*(T_c)$ is less than e_a, dew formation occurs on the respective surface and the condensed moisture is added to the surface interception store, M_c or M_g. The transpiration rate is zero under these conditions.

Following the calculation of the transpiration rates, E_{dc} and E_{dg}, (46) is used to obtain ψ_{lc} and ψ_{lg}, after which the abstraction of water from the different soil layers is given by

$$E_{d_i} = \frac{D_i}{z_d}\left(\frac{\psi_i - \psi_l - z_T}{\bar{r}_{\rm plant} + \bar{r}_{\rm soil}}\right)\rho_w, \tag{53}$$

where

E_{d_i} = rate of extraction of transpired water from ith soil layer, m s^{-1}
= 0, if $E_d < 0$.

Equations (46), (47) and (53) must be applied with values of the parameters appropriate to the canopy and the ground cover. In the case of the ground cover, the rooting depth may not extend down to the bottom of the second soil layer; ψ_r and E_{d_i} are then calculated using (47) and (53) with the D_i terms multiplied by the fractional depths of those soil layers that are occupied by the roots. All of the calculations that relate the stomatal resistances to the abstraction of soil water are performed for both canopy and ground cover vegetation.

The direct evaporation from the soil surface, E_s, depends on the difference between the vapor pressure at the soil surface and the vapor pressure of the canopy air space, e_a. Because the gradient of vapor pressure in the soil changes sharply near the surface, a surface resistance, $r_{\rm surf}$, is used to relate the surface vapor pressure to the depth-averaged wetness of the top soil layer, W_1. For this, we currently use the empirical expression given by Shu Fen Sun (1982)

$$r_{\rm surf} = d_1 + d_2(W_1)^{-d_3}, \tag{54}$$

where

$r_{\rm surf}$ = bare soil surface resistance, s m^{-1}
= 0, when $e_*(T_{gs}) < e_a$
d_1, d_2, d_3 = empirical constants.

Equation (54) is used only when $e_*(T_{gs}) \geq e_a$. When $e_*(T_{gs}) < e_a$, which is the condition of the dew formation (negative evaporation), $r_{\rm surf}$ is set to zero.

The direct evaporation from the soil surface is given by

$$\lambda E_s = \frac{[f_h e_*(T_{gs}) - e_a]}{r_{\rm surf} + r_d}\frac{\rho c_p}{\gamma}(1 - V_g), \tag{55}$$

where

E_s = soil evaporation rate, m s^{-1}
f_h = relative humidity of the air at the soil surface
= $e^{(\psi_1 g/RT_{gs})}$, when $e_*(T_{gs}) \geq e_a$
= 1, when $e_*(T_{gs}) < e_a$
ψ_1 = soil moisture potential in the top soil layer, from (48), m
g = acceleration of gravity, m s^{-2}
R = gas constant for water vapor, J kg K^{-1}.

Setting $f_h = 1$ and $r_{\rm surf} = 0$ when $e_*(T_{gs}) < e_a$ makes the rate of dew formation on the soil independent of the amount of moisture in the soil.

b. *Interception and interception loss*

The interception and evaporation of precipitation held on the leaf surfaces is modeled simply in SiB. First, the interception of the rainfall is determined by an adaptation of the expression describing the exponential attenuation of radiation when the flux is vertical and the leaves are black. The rates of inflow (interception) and of outflow (drainage of water stored on the vegetation) for the canopy are then given by

$$P_c = P(1 - e^{-K_c L_{tc}})V_c,$$

$$D_c\begin{cases} = 0, & \text{when } M_c < S_c \\ = P_c & \text{when } M_c = S_c, \end{cases} \tag{56}$$

where

P_c = rate of rainfall interception by the canopy, m s^{-1}
D_c = rate of canopy drainage, m s^{-1}
M_c = water held on the canopy, m

S_c = maximum value of M_c, m
$\approx (2 \times 10^{-4}$ to $5 \times 10^{-4})\overline{L_{tc}}$, m
K_c = extinction coefficient for a canopy with black leaves and vertical flux
$= G(\mu)/\mu$, where $\mu = 1$

and for the ground cover

$$P_g = (P - P_c + D_c)(1 - e^{-K_g L_{tg}})V_g$$

$$D_g \begin{cases} = 0, & \text{when} \quad M_g < S_g \\ = P_g, & \text{when} \quad M_g = S_g, \end{cases} \tag{57}$$

where

P_g = rate of rainfall interception by the ground cover, m s^{-1}
D_g = rate of drainage from the ground cover, m s^{-1}
M_g = water held on the ground cover, m
S_g = maximum value of M_g, m
$\approx (2 \times 10^{-4}$ to $5 \times 10^{-4})\overline{L_{tg}}$, m
K_g = as for K_c, but for the ground cover.

The effective precipitation rate on the soil surface, P_0, is then given by

$$P_0 = P - (P_c + P_g) + (D_c + D_g); \tag{58}$$

$(P_0 - P_1)$ is the surface runoff where P_1 is the infiltration rate into the soil.

The calculation of the transfers described by Eqs. (56)–(58) is performed at the beginning of the time step, so that the vegetation interception stores, M_c and M_g, are determined prior to the energy balance calculations for the time step.

The quantities M_c, M_g are used to determine the fractional wetted areas of the canopy and ground cover, W_c, W_g. These wetted areas are assigned zero surface resistance so that the rates of evaporation from the wetted portions of the vegetation are

$$\lambda E_{wc} = \frac{[e_*(T_c) - e_a]}{\overline{r_b}} \frac{\rho c_p}{\gamma} W_c, \tag{59}$$

$$\lambda E_{wg} = \frac{[e_*(T_{gs}) - e_a]}{r_d} \frac{\rho c_p}{\gamma} W_g V_g, \tag{60}$$

where

E_{wc}, E_{wg} = rates of evaporation from wet portions of the canopy and ground cover leaves, m s^{-1}
W_c, W_g = fractions of the canopy and ground cover leaves that are wet
$W_c = M_c/S_c$, $0 \leqslant W_c \leqslant 1$, when $e_*(T_c) > e_a$
$= 1$, when $e(T_c) < e_a$
$W_g = M_g/S_c$, $0 \leqslant W_g \leqslant 1$, when $e_*(T_{gs}) > e_a$
$= 1$, when $e_*(T_{gs}) < e_a$.

When the calculated interception loss from the canopy or ground cover, E_{wc} or E_{wg}, is predicted to exceed the storage, M_c or M_g, during a time step, the calculation is repeated with the constraint that the respective interception loss is limited to the respective supply of stored water. The assumption of a constant value of W_c and W_g throughout the time step, Δt, introduces a measure of dependence between E_{wc}, E_{wg} and Δt; however, this effect is not significant for reasonable values of Δt ($\leqslant$1 hour).

It is assumed in (59) and (60) that the wet and dry parts of the leaves are at the same temperature. There are a number of arguments and some evidence (see Hancock et al., 1983) that this is better than an assumption of separate temperatures. The main theoretical argument in favor of this treatment is that energy is efficiently exchanged between the wet and dry portions of the foliage in the form of conduction and sensible heat transfers within the leaf laminar boundary layers and canopy air space.

The interception losses and condensation gains are included in the energy balance equations in section 9, where they are combined with the transpiration and the soil evaporation to yield the total latent heat fluxes from the canopy and ground.

The interception of snow by the canopy and ground is dealt with in the same way as rainfall. The evaporation rates from the snow-covered surfaces are adjusted for the additional energy required for sublimation (see appendix B).

The SiB canopy model has two notable simplifications:

(i) The assumption of constant potential differences, $(T_c - T_a)$ and $(e_*(T_c) - e_a)$, between all the canopy leaf elements and the canopy air space is not realistic. However, the effects of the differential absorption of radiation, the extinction of $\overline{r_b}$ and $\overline{r_c}$, and the mixing within the canopy air space should act to mitigate the error.

(ii) The simple addition of $\overline{r_c}$ and $\overline{r_b}$ is mathematically incorrect. The resistance to the transfer of water vapor from the canopy leaves to the local air space should be determined by integrating the inverse sum of r_s and r_b. Usually, however, $r_b \ll r_s$, and so the error is small.

8. Hydraulic diffusion of water in the soil

A three-layer isothermal model is used to determine the hydraulic diffusion and gravitational drainage of water in the soil. The transfer of water between adjacent layers is given by

$$Q_{i,i+1} = \bar{K}\left[2\frac{\psi_i - \psi_{i+1}}{D_i + D_{i+1}} + 1\right], \quad \text{for} \quad i = 1, 2$$

$$\bar{K} = \frac{D_i K_i + D_{i+1} K_{i+1}}{D_i + D_{i+1}} \tag{61}$$

where

$Q_{i,i+1}$ = downward flow from soil layer i to soil layer $i+1$, m s^{-1}
K_i = hydraulic conductivity of ith layer $= K_s W_i^{(2B+3)}$, m s^{-1}
$\bar{K}$ = estimated effective hydraulic conductivity between layers, m s^{-1}
W_i = soil moisture fraction of layer i $= (\theta_i/\theta_s)$
$\psi_i = \psi_s W_i^{-B}$, from (48), m
K_s = hydraulic conductivity at saturation, m s^{-1}.

In SiB, the flow out of the bottom layer is determined by gravitational drainage only, so that

$$Q_3 = \sin x K_s W_3^{(2B+3)}, \tag{62}$$

where x is the mean slope angle.

The equation set (5), (6), (7), (61) and (62) is combined and integrated using an implicit backward method operating on the final values of ψ_i. These calculations are carried out at the end of the time step following the heat flux calculations (see section 9). The infiltration of liquid water is assumed to be zero when the ground temperature, T_{gs}, is below the freezing point of water (see appendix B).

9. Solution of the governing prognostic equations

If we assume no storage of heat or moisture at any of the junctions of the resistance network shown in Fig. 2, we can write the area-averaged sensible and latent heat fluxes from the canopy and ground as follows:

a. From canopy vegetation to canopy air space

Sensible heat flux

$$H_c = \frac{2(T_c - T_a)}{\bar{r}_b} \rho c_p \tag{63}$$

Latent heat flux

$$\lambda E_c = (e_*(T_c) - e_a) \frac{\rho c_p}{\gamma} \left[\frac{W_c}{\bar{r}_b} + \frac{1 - W_c}{\bar{r}_b + \bar{r}_c} \right], \tag{64}$$

where

T_c = canopy temperature, K
T_a = temperature of the air at the canopy source height, K
$e_*(T_c)$ = saturation water vapor pressure at temperature T_c, mb
e_a = water vapor pressure of the air at the canopy source height, mb
W_c = wetted fraction of canopy
$\bar{r}_b$ = bulk boundary layer resistance for the canopy leaves, s m^{-1}
$\bar{r}_c$ = bulk stomatal resistance of the canopy leaves, s m^{-1}.

b. From ground (ground cover and soil) to canopy air space

Sensible heat flux

$$H_{gs} = \frac{(T_{gs} - T_a)}{r_d} \rho c_p \tag{65}$$

Latent heat flux

$$\lambda E_{gs} = [e_*(T_{gs}) - e_a] \frac{\rho c_p}{\gamma} \times \left(V_g \left[\frac{W_g}{r_d} + \frac{1 - W_g}{r_d + r_g} \right] + \frac{(1 - V_g) h_s}{r_{\text{surf}} + r_d} \right), \tag{66}$$

where

T_{gs} = temperature of the ground cover and soil surface, K
$e_*(T_{gs})$ = saturation water vapor pressure at temperature T_{gs}, K
V_g = fractional cover of ground vegetation
W_g = wetted fraction of ground cover
h_s = factor that adjusts for the relative humidity of the air at the soil surface
$$= \frac{f_h e_*(T_{gs}) - e_a}{e_*(T_{gs}) - e_a}$$
f_h = relative humidity of the air at the soil surface
$= e^{(\psi_1 g / R T_{gs})}$, when $e_*(T_{gs}) \geq e_a$
$= 1$, when $e_*(T_{gs}) < e_a$
ψ_1 = moisture potential in the top soil layer, m
r_d = aerodynamic resistance between the ground and the air at the canopy source height, s m^{-1}
r_g = bulk stomatal resistance of the ground cover, s m^{-1}
r_{surf} = bare soil surface resistance, s m^{-1}.

c. From canopy air space to reference height in the atmospheric boundary layer

Sensible heat flux

$$H_c + H_{gs} = \frac{(T_a - T_r)}{r_a} \rho c_p, \tag{67}$$

Latent heat flux

$$\lambda E_c + \lambda E_{gs} = \frac{(e_a - e_r)}{r_a} \frac{\rho c_p}{\gamma} \tag{68}$$

where

T_r = temperature of the air at the reference level within the atmospheric boundary layer, K,
e_r = water vapor pressure of the air at the reference level within the atmospheric boundary layer, mb,

r_a = aerodynamic resistance between the air at the canopy source height and the air at the reference level within the atmospheric boundary layer, s m^{-1}.

The preceding sections showed how the six resistances r_a, $\overline{r_b}$, r_d, $\overline{r_c}$, r_g and $r_{\rm surf}$ are calculated. Given these values and the calculation of Rn_c and Rn_{gs} from section 5, Eqs. (1) and (2) may be combined with (63)–(68) to eliminate T_a and e_a and yield two equations in T_c and T_{gs}. These may then be solved by an implicit backward method whereby:

Canopy

$$C_c \frac{\Delta T_c}{\Delta t} = Rn_c + \frac{\partial Rn_c}{\partial T_c}\Delta T_c + \frac{\partial Rn_c}{\partial T_{gs}}\Delta T_{gs} - H_c - \frac{\partial H_c}{\partial T_c}\Delta T_c - \frac{\partial H_c}{\partial T_{gs}}\Delta T_{gs} - \lambda E_c - \frac{\partial \lambda E_c}{\partial T_c}\Delta T_c - \frac{\partial \lambda E_c}{\partial T_{gs}}\Delta T_{gs}, \quad (69)$$

Ground

$$C_{gs} \frac{\Delta T_{gs}}{\Delta t} = Rn_{gs} + \frac{\partial Rn_{gs}}{\partial T_{gs}}\Delta T_{gs} + \frac{\partial Rn_{gs}}{\partial T_c}\Delta T_c - H_{gs} - \frac{\partial H_{gs}}{\partial T_{gs}}\Delta T_{gs} - \frac{\partial H_{gs}}{\partial T_c}\Delta T_c - \lambda E_{gs} - \frac{\partial \lambda E_{gs}}{\partial T_{gs}}\Delta T_{gs} - \frac{\partial \lambda E_{gs}}{\partial T_c}\Delta T_c \quad (70)$$

Adding ΔT_c and ΔT_{gs} to the initial values of T_c and T_{gs}, at time t_0, gives us T_c and T_{gs} for time $t_0 + \Delta t$.

The values of T_c, T_{gs}, at time $t_0 + \Delta t$, together with the calculated values of P_c, P_g and P_1, at time t_0, are then used in the finite difference equivalents of (3)–(8) to obtain the changes in M_c, M_g and W_1, W_2, W_3 over the time step Δt.

In summary, given (i) the initial state of the seven prognostic variables of SiB (T_c, T_{gs}, M_c, M_g, W_1, W_2 and W_3); (ii) the boundary conditions (T_r, e_r, u_r, $F_{\Lambda,\mu(0)}$ and P); and (iii) the morphological, physiological and physical properties of the canopy, ground cover and soil, we calculate: (i) the net fluxes of energy, mass and momentum across the lower boundary of the atmosphere in the form of:

Radiation $(Rn_c + Rn_{gs})$,

Sensible heat $(H_c + H_{gs})$,

Latent heat $(\lambda E_c + \lambda E_{gs})$ and

Shear stress $\vec{\tau}|_{z_r} = \tau|_{z_r}(\vec{u}_r/u_r)$

(ii) the time rate of change and the future state of the prognostic variables, and (iii) the initial and future state of the diagnostic variables T_a and e_a. Figure 8 shows the sequence of calculations performed by SiB over one time step.

10. Implementing SiB in a general circulation model

a. *Initialization of the prognostic variables of SiB*

In the current version of SiB, the prognostic variables are T_c, T_{gs}; M_c, M_g; W_1, W_2, W_3.

The temperatures of the canopy and of the ground cover and soil surface, T_c and T_{gs}, respectively, are fast-response or "quasi-diagnostic" variables; therefore, no large error will result if they are initialized with any values that are of the same order of magnitude as the true values.

The intercepted water stored on the leaves of the canopy and the ground cover, respectively M_c and M_g, may also be initialized in an arbitrary way as their combined upper limit, $S_c + S_g$, should not exceed 2–5 mm of water.

An accurate initialization of the soil moisture stores, W_1, W_2 and W_3, will be important as the total soil moisture available for transpiration can be as large as several hundred millimeters, which is equivalent to a few months worth of transpiration loss. Consequently, for climate predictions and for weather predictions beyond the range of a few days, the correct initialization of the soil moisture may be as important as the correct initialization of the surface pressure and atmospheric temperature fields. The best procedure for initializing the soil moisture (as well as T_c, T_{gs}; M_c, M_g) may be the one now used to initialize the prognostic variables of the atmosphere: the method of four-dimensional dynamical data analysis and assimilation.

b. *Specification of boundary conditions for SiB*

The forcing variables T_r, e_r and u_r are currently specified by all GCMs as grid area-averaged conditions.

At present, almost all general circulation models provide the downward thermal radiative flux, $F_{t,d}$, and the total of the downward shortwave fluxes, $F_{s,b}$, $F_{s,d}$, $F_{n,b}$ and $F_{n,d}$, at the lower boundary of the atmosphere. In whatever way the GCM calculates radiation flux through the atmosphere, the downward flux through the lower boundary must be partitioned into the five components required by SiB. In doing this, care must be taken to change neither the total downward flux at the lower boundary nor its vertical derivative at any height within the atmosphere. General circulation models that do not have a diurnal variation of insolation are not suitable for use with SiB.

A difficulty that may arise when SiB in its present form is coupled to existing GCMs concerns the relationship between the subgrid-scale variation of the pre-

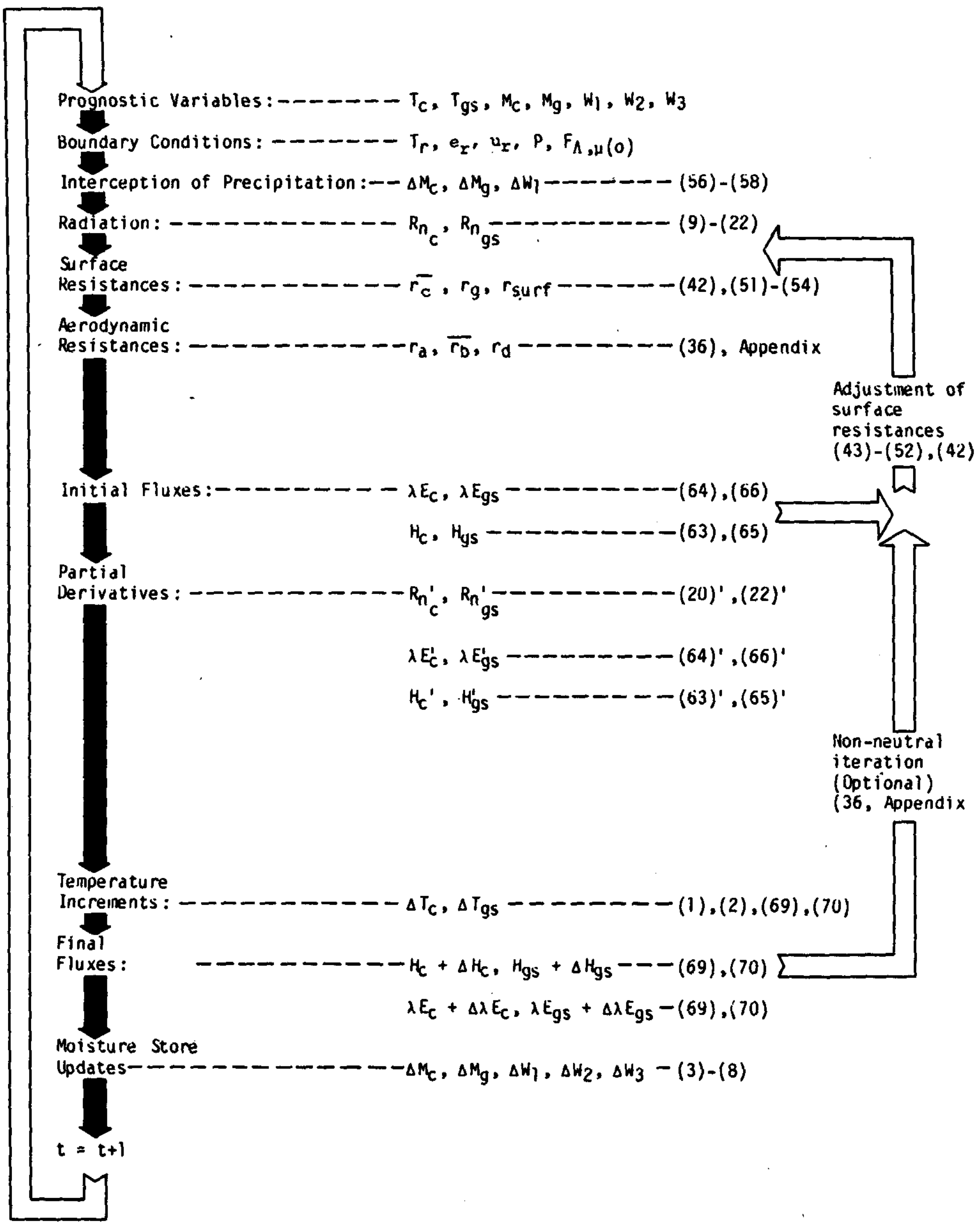

FIG. 8. Sequence of calculations used to advance the prognostic variables of SiB by one time step. Numbers in parentheses refer to equations in text. Dash symbol refers to partial differentiation with respect to T_c and T_{gs}.

cipitation, P', and the grid area-averaged rate of rainfall interception by the canopy and ground cover, P_c and P_g, given by (56) and (57).

In nature, there are two reasons for large subgrid-scale variations in rainfall: variability in topography and the relatively small scale of convective cells. Whatever the cause, subgrid-scale variability in precipitation rate will affect the time- and space-averaged values of rainfall interception and interception loss. By way of illustration, suppose that during a precipitation event and the drying out interval that follows, the area-averaged rainfall and potential interception loss are of the same magnitude. If the rainfall is uniformly spread over the area, nearly all of it will return to the atmosphere from the wet surface of the vegetation and none will infiltrate into the soil. In the case of convective rainfall regimes, the local precipitation yield is usually high compared to the storage capacities S_c and S_g; so that in its present form SiB will produce an underestimate of the rainfall infiltration into the soil. Generally, increased spatial variability in rainfall intensity will lead to lower interception losses.

Further work must be done to determine a means of quantifying the subgrid-scale distribution of convective rainfall if the area-averaged interception loss is to be calculated accurately by SiB.

11. Summary and discussion

The Simple Biosphere Model is an attempt to incorporate biophysical realism in a formulation of the land surface energy balance at a level of sophistication appropriate to existing atmospheric general circulation models. All the current elements of SiB are necessary for the realistic calculation of the radiation, heat and momentum fluxes, R_n, λE, H and $\vec{\tau}$.

In addition, equations must be added that will correctly partition the GCM produced grid area-averaged convective rainfall into grid area-averaged interception storage, interception loss, surface runoff and soil infiltration terms. To treat all of the GCM produced rainfall as if it were large-scale, and therefore uniformly distributed over the grid area, may result in significant error.

When these additions have been made, SiB can be used within GCMs for certain sensitivity studies, such as the influence of vegetation on the atmospheric circulation and rainfall, and for climate simulations when both the distribution of vegetation type and soil moisture are prescribed.

The SiB model, with prescribed vegetation parameters and interactive soil moisture, can be used for predictions of the atmospheric circulation and precipitation fields for short periods of up to a few weeks. Here, the initial state of the soil moisture and snowcover must be known.

As discussed earlier, the initial state of the soil moisture can best be obtained by four-dimensional dynamical data assimilation and initialization but this process cannot be used to initialize the parameters of the vegetation properties (see Table 1) until we have added appropriate governing equations. When this has been done, SiB could be used for predictions on longer timescales from a few weeks to a few years.

12. Independent tests of SiB

Prior to its use within a GCM, SiB can be tested and evaluated using micrometeorological measurements and biophysical data at a point. This will test the different subcomponents of the model and the functioning of the whole. Because high quality field data are not available for the world's natural vegetation communities, tests were made using measurements recorded over some agricultural crops and a coniferous forest. The results of these tests will be presented in a forthcoming paper.

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APPENDIX A

Derivation of Turbulent Transfer Coefficients, C_1, C_2, and C_3 and the Canopy Source Height, h_a

1. Bulk boundary layer resistance coefficient, C_1

Solution of Eqs. (24) and (25) in text gives the following expressions for K_m and u within the canopy ($z_1 < z < z_2$).

$$K_m = \sigma u, \tag{A1}$$

$$u = u_2(\beta_1 e^{-\beta_3 z} + \beta_2 e^{\beta_3 z})^{1/2},$$

$$\beta_3 = \left(\frac{2Cd\overline{Ld}}{\sigma P_s}\right)^{1/2}. \tag{A2}$$

The coefficients β_1 to β_3 are obtained by iteration on σ over Eqns. (23)–(27).

Calculation of C_1, the constant that determines the relation between the wind speed at the canopy top, u_2, and the bulk boundary layer resistance, $\bar{r}_b$, under neutral conditions, is given by integrating (31) with (A2) replacing u. This yields:

$$C_1 = \left[\frac{\overline{Ld}}{\beta_3 C_s}\frac{(\beta_2)^{1/4}}{P_s}4e^{\beta_3 z/4}(1 - S_{n1})\right]_{z_1}^{z_2},$$

$$S_{n1} = \left\{\sum_{n=1}^{\infty}\left(\frac{\beta_1}{\beta_2}\right)^n \frac{e^{-2n\beta_3 z}}{(8n-1)}\frac{(-1)^n}{n!4^n}\prod_{i=1}^{n}[4(n-i)-1]\right\}, \tag{A3}$$

if

$$\frac{\beta_2}{\beta_1} > e^{-2\beta_3 z} \quad \text{and} \quad \beta_1, \beta_2 > 0,$$

or if

$$\beta_1 < 0,$$

and

$$C_1 = \left[\frac{\overline{Ld}}{\beta_3 C_s}\frac{(\beta_1)^{1/4}}{P_s}4e^{-\beta_3 z/4}[1 + S_{n2}]\right]_{z_1}^{z_2},$$

$$S_{n2} = \left\{\sum_{n=1}^{\infty}\left(\frac{\beta_2}{\beta_1}\right)^n \frac{e^{2n\beta_3 z}}{(8n-1)}\frac{(-1)^n}{n!4^n}\prod_{i=1}^{n}[4(n-i)-1]\right\} \tag{A4}$$

if

$$\frac{\beta_2}{\beta_1} < e^{-2\beta_3 z} \quad and \quad \beta_1, \beta_2 > 0,$$

or if

$$\beta_2 < 0.$$

A simple modification to (31) is made for the effects of free convection when the canopy temperature exceeds the local air temperature. If sensible heat is being lost from a surface by free convection, the resistance to heat transfer may be given by

$$r_f = \frac{l}{D_H \cdot \mathrm{Nu}} \tag{A5}$$

where

r_f = resistance assuming free convection, s m^{-1}
l = horizontal length scale, m
Nu = Nusselt number
D_H = thermal diffusivity of air, m^2 s^{-1}.

The Nusselt number is often described empirically for planar surfaces (see, for example, Monteith; 1973) by:

$$\mathrm{Nu} = m_1 \mathrm{Gr}^{m_2}$$

where

$m_1 \approx 0.5$

$m_2 \approx 0.25$

where Gr is the Grashof number given by

$$\mathrm{Gr} = \frac{1}{273} g l^3 (T_{\mathrm{surf}} - T_{\mathrm{air}}) v, \tag{A6}$$

where

v = kinematic viscosity of air, m^2 s^{-1}
g = acceleration due to gravity
= 9.8 m s^{-2}

and m_1 and m_2 are coefficients determined from observation. Combination of the above equations with suitable values for the various constants yields

$$r_{fc} = 890\left(\frac{l}{T_{\mathrm{surf}} - T_{\mathrm{air}}}\right)^{1/4}. \tag{A7}$$

When dealing with a particular micrometeorological problem, it is normal practice to calculate the resistance assuming both forced and free convection and then to use the lower estimate in further calculations of heat transfer (Monteith, 1973). In SiB, we shall assume that both forms of transfer are contributing to convection over the whole unstable region and that, as conditions conducive to one form of transfer necessarily exclude significant interference from the other, we may write the *total* resistance as:

$$\frac{1}{r} = \frac{1}{r_{(\text{forced convection})}} + \frac{1}{r_{(\text{free convection})}}. \tag{A8}$$

The expression for $\bar{r}_b$ then becomes:

$$\frac{1}{\bar{r}'_b} = \frac{\sqrt{u_2}}{C_1} + \frac{\bar{L}_t}{890}\left(\frac{T_c - T_a}{l}\right)^{\frac{1}{4}}, \tag{A9}$$

where

$\bar{r}'_b$ = nonneutral value of $\bar{r}_b$, s m^{-1}.

The value chosen for the length scale, l, should be of the order of the leaf dimensions, i.e., a few centimeters. In any case, the estimates of $\bar{r}'_b$ given by (A9) are insensitive to the value of l due to the $\frac{1}{4}$ power term.

The combination of free and forced convection terms in Eq. (A9) is unrealistic from a physical viewpoint, but does result in a smoothly varying function that prevents canopy–air temperature differences from rising to unrealistic levels when windspeeds are low and the radiation input is high.

2. Soil surface to canopy source height coefficient, C_2

Here r_d is the aerial resistance to heat and vapor flow from soil surface to the canopy source height, h_a, as defined in (33). To begin with, we shall consider the resistance to momentum transfer, which is defined as:

$$r_{d_m} = \int_{z_{gs}}^{h_a} \frac{1}{K_m} \partial z,$$

where

r_{d_m} = resistance to momentum transfer, s m^{-1}
K_m = momentum transfer coefficient, m^2 s^{-1}
z_{gs} = effective ground roughness length, m.

Within this height range, there are two distinct transfer zones. These are:

(i) From soil surface to z_1: Here the resistance to momentum transfer is parameterized according to the conventional roughness length approach.

(ii) Within the canopy, z_1 to h_a: $K_m = \sigma u$. Thus:

$$r_{d_m} = \int_{z_s}^{z_1} \frac{1}{K_m} \partial z + \int_{z_1}^{h_a} \frac{1}{K_m} \partial z. \tag{A10}$$

The first term on the right-hand side of (A10) may be given in terms of the ground drag coefficient $C_{D_{gs}}$ and wind speed at z_1. Here it is estimated by a simple log-linear expression:

$$\int_{z_{gs}}^{z_1} \frac{1}{K_m} \partial z = \frac{1}{u_1}\left[\frac{1}{k}\ln\left(\frac{z_1}{z_{gs}}\right)\right]^2$$

$$= \frac{1}{u_2}(\beta_1 e^{-\beta_3 z_1} + \beta_2 e^{\beta_3 z_1})^{-1/2}\left[\frac{1}{k}\ln\left(\frac{z_1}{z_{gs}}\right)\right]^2. \tag{A11}$$

(iii) The resistance over the distance between z_1 and the canopy source height, h_a, is given by:

$$\int_{z_1}^{h_a} \frac{1}{K_m}\,\partial z = \frac{1}{\sigma}\int_{z_1}^{h_a} \frac{1}{u}\,\partial z$$
$$= \frac{1}{\sigma u_2}\int_{z_1}^{h_a} (\beta_1 e^{-\beta_3 z} + \beta_2 e^{\beta_3 z})^{-1/2}\partial z. \quad \text{(A12)}$$

Expansion, integration term by term, and collection of the terms in (A12) into a series form yields:

$$\int_{z_1}^{h_a} \frac{1}{K_m}\,\partial z = \left[\frac{-2e^{-\beta_3 z/2}}{u_2\beta_3\sigma\beta_2^{1/2}}\left(1 - \sum_{n=1}^{\infty}\left(\frac{\beta_1}{\beta_2}\right)^n \times \frac{e^{-2n\beta_3 z}}{n!(4n+1)}\prod_{i=1}^{n}\left(\frac{1-2i}{2}\right)\right)\right]_{z_1}^{h_a}$$
$$\text{if } \frac{\beta_2}{\beta_1} > e^{-2\beta_3 z} \text{ or } \beta_1 < 0 \quad \text{(A13)}$$
$$= \left[\frac{2e^{\beta_3 z/2}}{u_2\beta_3\sigma\beta_1^{1/2}}\left(1 + \sum_{n=1}^{\infty}\left(\frac{\beta_2}{\beta_1}\right)^n \times \frac{e^{2n\beta_3 z}}{n!(4n+1)}\prod_{i=1}^{n}\left(\frac{1-2i}{2}\right)\right)\right]_{z_1}^{h_a}$$
$$\text{if } \frac{\beta_2}{\beta_1} < e^{-2\beta_3 z} \text{ or } \beta_2 < 0. \quad \text{(A14)}$$

Equations (A11), (A13) and (A14) all have an inverse u_2 term. Addition of the constant portions of (A11) and either (A13) or (A14) yields an estimate of C_2, which when divided by u_2, gives the neutral value of r_d for momentum. An adjustment is made for the different diffusivities of water and heat in order to give r_d for latent and sensible heat transfer:

$$\left.\begin{aligned} r_d &= \frac{C_2}{u_2\phi_H} \\ \phi_H &= \left[1 + 9\frac{(T_{gs} - T_a)}{T_{gs}u_2^2}z_2\right]^{1/2}, \quad \phi_H \geq 1 \end{aligned}\right\}. \quad \text{(A15)}$$

Equation (A15) is based on the unstable transfer correction factor used by Goudriaan (1977) adapted from Businger et al. (1971), where the mean gradients of temperature and wind speed are estimated over the height of the canopy. No correction is made for stable conditions.

3. Canopy source height to reference height coefficient, C_3

The resistance to transfer of heat and moisture from the canopy source height, h_a, to the reference height, z_r, is described as the sum of three terms:

$$r_a = \int_{h_a}^{z_2} \frac{1}{K_s}\,dz + \int_{z_2}^{z_m} \frac{1}{K_s}\,\partial z + \int_{z_m}^{z_r} \frac{1}{K_s}\,\partial z, \quad \text{(A16)}$$

where

K_s = heat or moisture transfer coefficient, m^2 s^{-1}.

The first term on the right-hand side of (A16) is obtained in a similar manner as for the estimation of r_d within the canopy. Substitution of the limits h_a and z_2 in place of z_1 and h_a in (A13) and (A14) yields a constant divided by u_2, the value of the constant depending on β_1, β_2, β_3 and σ. The second term on the right-hand side of (A16) is obtained from the analytical expressions of Paulson (1970) except for a modification for the divergence between the expected (log-linear extrapolation) values of the transfer coefficients and their observed values in this region. The simplest adjustments that can be made are

$$K_m|_{z_2} = G_1 K_m^*|_{z_2}, \quad \text{(A17)}$$

where

K_m = "actual" value of K_m, m^2 s^{-1}
K_m^* = value of K_m as given by extrapolation of log profile
$G_1 \approx 2.0$ (from Garratt, 1978)

$$\int_{z_2}^{z_m} \frac{1}{K_m}\,\partial z = G_2\int_{z_2}^{z_m} \frac{1}{K_m^*}\,\partial z, \quad \text{(A18)}$$

where

$G_2 \approx 0.5$ (from Garratt, 1978)

$$\int_{z_2}^{z_m} \frac{1}{K_s}\,\partial z = G_3\int_{z_2}^{z_m} \frac{1}{K_s^*}\,\partial z \quad \text{(A19)}$$

where

$G_3 \approx 0.5$ (from Garratt, 1978).

Then,

$$\int_{z_2}^{z_m} \frac{1}{K_s}\,\partial z = G_3\frac{1}{ku_*}\left[\ln\left(\frac{z-d}{z_0}\right) - \phi_2\right]_{z_m}^{z_r}, \quad \text{(A20)}$$

$$\int_{z_m}^{z_r} \frac{1}{K_s}\,\partial z = \frac{1}{ku_*}\left[\ln\left(\frac{z-d}{z_0}\right) - \phi_2\right]_{z_2}^{z_m}, \quad \text{(A21)}$$

where

ϕ_2 = Paulson (1970) nonneutral correction factor for heat and vapor
u_* = friction velocity, m s^{-1}.

Equations (A17) and (A18) must be used in conjunction with (23)–(28) in the text to obtain the values of β_1, β_2, β_3, σ, d and z_0. Equation (A18) can then be used to obtain u_2 by

$$u_2 = u_m - \frac{u_*}{k}G_2\left[\ln\left(\frac{z-d}{z_0}\right) - \phi_1\right]_{z_2}^{z_m} \quad \text{(A22)}$$

where

u_m = wind speed at top of transition layer, m s^{-1}
ϕ_1 = Paulson (1970) nonneutral correction factor for momentum transfer.

Under neutral conditions, (A16) can be manipulated to yield r_a for heat and vapor in terms of a constant, C_3, divided by the wind speed at reference height. Under nonneutral conditions, the Paulson (1970) values ϕ_1 and ϕ_2 are nonzero and iteration must be used to obtain compatible values of u_* and u.

The above method of estimating C_3 is useful when considering boundary conditions a short distance above the canopy. When SiB is to be linked to a General Circulation Model, a parameterization on the lines of that proposed by Deardorff (1972) will be used to span the interval z_2–z_r.

4. The canopy source height, h_a

The position of the canopy source height can be estimated by obtaining the center of gravity of the $1/r_b$ integral. Thus:

$$\int_{z_1}^{h_a} \frac{\overline{Ld}}{r_b}\,\partial z = \int_{h_a}^{z_2} \frac{\overline{Ld}}{r_b}\,\partial z = \frac{1}{2}\int_{z_1}^{z_2} \frac{\overline{Ld}}{r_b}\,\partial z = \frac{1}{2\bar{r}_b}. \quad \text{(A23)}$$

The rightmost expression in (A23) is obtained simply by multiplying C_1 by 2. We may obtain h_a by replacing one of the limits in (A3) or (A4) by successive estimates of h_a until equality with $2C_1$ is reached. (Note that h_a is independent of wind speed.)

APPENDIX B

Incorporating the Effects of Snow into the Simple Biosphere (SiB) Model

A preliminary attempt has been made to account for the effects of snow and ice within SiB. This currently consists of simple modifications to some of the parameters and calculations described in the rest of the paper. The various effects are itemized by process below.

(i) *Definition of snowfall.* All precipitation is currently assumed to be at the same temperature as the air at the reference height, T_m. If this is less than the freezing point of water, T_f, the precipitation is taken to be snow.

(ii) *Interception and accumulation of snowfall.* The canopy is assumed to intercept snow in the same way as rain, up to the same mass limit, S_c. Snow may accumulate on the ground beyond the ground cover interception capacity, S_g. So:

$$0 \leqslant M_c \leqslant S_c; \quad \text{for all } T_c, \quad \text{(B1a)}$$

$$0 \leqslant M_g \leqslant S_g; \quad T_{gs} > T_f,$$

$$0 \leqslant M_g; \quad T_{gs} \leqslant T_f. \quad \text{(B1b)}$$

A simple calculation is performed to carry out adjustments to T_c and T_{gs} following the interception of precipitation. The calculation involves the surface temperatures, T_c and T_{gs}, the heat capacities, C_c and C_{gs}, and the temperature and mass of the precipitation, T_m and $P_c\Delta t$ and $P_g\Delta t$. The adjustment gives rise to adjusted weighted values of T_c and T_{gs} and may involve melting or freezing of the intercepted precipitation.

The depth of snow intercepted by the ground, z_f, is calculated as

$$z_f = 5M_g, \quad \text{(B2)}$$

it being assumed that the snow depth is five times the equivalent water depth.

The fractional area of ground covered by snow, F_f, is taken as

$$F_f = M_g/0.01, \quad T_{gs} \leqslant T_f \quad 0 \leqslant F_f \leqslant 1, \quad \text{(B3)}$$

which assumes full coverage of the ground when the snow is 0.05 m deep.

(iii) *Radiation.* The scattering coefficient of the leaf elements in the canopy is adjusted for the amount of intercepted snow.

$$\tilde{\omega} = \omega\left(1 - \frac{M_c}{S_c}\right) + \omega_f\frac{M_c}{S_c}, \quad T_c < T_f \quad \text{(B4)}$$

where

$\tilde{\omega}$ = mean scattering coefficient of canopy elements and intercepted snow
ω = scattering coefficient of phyto-elements
ω_f = scattering coefficient of snow.

The reflectance of the ground is similarly adjusted:

$$\tilde{A}_{gs} = A_{gs}(1 - F_f) + A_fF_f, \quad T_{gs} < T_f \quad \text{(B5)}$$

where

$\tilde{A}_{gs}$ = albedo of snow-covered ground
A_{gs} = albedo of ground in absence of snow (see previous section)
A_f = snow albedo.

Currently, the snow is assumed to reflect radiation only (no transmission) so that $\omega_f = A_f$. The values of ω_f, A_f are assumed to be 0.8 in the visible and 0.4 in the near-infrared wavelength intervals, being reduced to 60% of these values when the surface temperature T is close to or at melting point, i.e., when $(T_f - 0.1) \leqslant T \leqslant T_f$.

(iv) *Turbulent transfer.* Simple adjustments to the turbulent transfer coefficients are made to account for the burying of vegetation by accumulated snow. A scale height, h_f, is introduced.

$$h_f = \frac{z_2 - z_f}{z_2},$$

$$T_{gs} \leqslant T_f$$

$$0.05 \leqslant h_f \leqslant 1. \quad \text{(B6)}$$

Then

$$\tilde{d} = z_2 - (z_2 - d)h_f, \quad \text{(B7a)}$$

$$\tilde{z}_0 = \frac{z_0}{z_2 - d}[z_2 - (z_2 - d)h_f], \quad \text{(B7b)}$$

$$\tilde{C}_1 = C_1 h_f, \quad \text{(B7c)}$$

$$\tilde{C}_2 = \frac{C_2}{h_f}, \quad \text{(B7d)}$$

where the '$\sim$' sign refers to the adjusted value of the quantity.

Also an adjustment to the interception capacity of the canopy is made,

$$\tilde{S}_c = S_c h_f. \quad \text{(B7e)}$$

(v) *Energy partition.* The heat capacity of the canopy or ground is calculated as a function of the biomass and/or soil material and intercepted water mass. (An upper limit is currently placed on the ground heat capacity for high values of z_f.)

Energy and mass exchange are treated as before except that an adjustment is made following the flux calculation.

$$E = \left(\frac{1}{\lambda + \lambda_f}\right)\lambda E \quad \text{(B8)}$$

where

λE = latent heat flux, calculated as in section 8, W m^{-2}
λ = latent heat of vaporization, J kg^{-1}
λ_f = latent heat of crystallization, J kg^{-1}
E = adjusted water mass flux, kg m^{-2} s^{-1}.

Additionally, a check is made to determine whether either of the surface temperatures have been predicted to pass through T_f. If so, adjustments are made to either melt or freeze the intercepted precipitation and the surface temperature is modified accordingly.

(vi) *Runoff.* When $T_{gs} < (T_f - 0.1)$, no infiltration or runoff occurs. When $(T_f - 0.1) < T_{gs} < T_f$, snowmelt may run off.

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