APPENDIX A

Prescription of Land Surface Parameters and the Conceptual Models

Three broad strategies are used to model land-surface processes in GCMs: first, the prescription of land surface parameters; second, the use of conceptual models; and finally, the use of biophysically based models. According to Sellers (1987), in earlier GCMs, particularly before the work of Dickinson (1984), all surface parameterizations had been confined only to the prescription of land surface parameters and the use of conceptual models. The two approaches are documented below following descriptions given by Sellers (1987, 1991 and 1992).

A.1: Prescription of Land Surface Parameters

Three principal equations are involved in this first approach, each describing the exchanges of radiation, momentum, and heat.

<u>Radiation</u>:

Radiant energy flux at the land surface can be described in term of the resultant shortwave radiation, downward longwave radiation and longwave radiation emitted by the surface. Here, the net radiation flux absorbed by the land surface can be expressed, with emphasis on surface condition, as

$$R_n = \int_0^{\infty} F_l(1 - \alpha_l).dl + \varepsilon_s F_t - \varepsilon_s \sigma_s T_s^4 \qquad (A.1)$$

where R_n = net radiation flux absorbed by the land surface, W m^{-2}; F_l = solar radiative flux density at wavelength l incident on the surface, W m^{-2}; α_l = spectral albedo; ε_s = emissivity of the surface; F_t = long-wave flux downward from the atmosphere, W m^{-2}; σ_s = Stefan-Boltzmann constant, W m^{-2} K^{-4}; and T_s = surface radiative temperature, K.

Momentum:

The shear stress exerted on the atmosphere by the surface depends on the wind velocity at a reference height and may be given by

$$\tau = \rho C_D u_r \mathbf{U}_r \tag{A.2a}$$

where τ = shear stress, kg m^{-1} s^{-2}; ρ = density of air, kg m^{-3}; C_D = surface drag coefficient; $\mathbf{U_r}$ = wind velocity at reference height z_r , m s^{-1}; and u_r = absolute value of $\mathbf{U_r}$. Here C_D may either be prescribed or obtained from:

$$C_D = \left(\frac{k}{\ln(z_r / z_o) - \phi_1} \right)^2 \tag{A.2b}$$

where k = von Karman's constant, given as 0.41; z_r = reference height, m**;** z_o = roughness length, m; and ϕ_1 **=** nonneutral Paulson (1970) coefficient.

Heat fluxes:

The energy available to the surface (R_n), is then partitioned into three flux terms; (i) soil heat flux, (ii) latent heat flux, and (iii) sensible heat flux and may be given in a balance equation as:

$$R_n - G = \lambda E + H \tag{A.3a}$$

where G = soil heat flux, W m^{-2}**;** E = evaporation rate, kg m^{-2} s^{-1}; λ = latent heat of vaporization, J kg^{-1}; and H = sensible heat flux, W m^{-2}**.**

The soil heat flux (G) is related to the ground heat capacity and the rate of change of ground temperature and may be given by:

$$G = C_g \frac{dT_g}{dt} \tag{A.3b}$$

where C_g = ground heat capacity, J m^{-2} K^{-1}; and T_g = ground temperature, K.

The ratio of sensible and latent heat fluxes is called as the Bowen ratio (B), and is given by

$$B = \frac{H}{\lambda E} \tag{A.3c}$$

Early GCMs used this equation set with prescribed fields of parameters α_l, C_D or z_o, and also B and C_g. Often, the prescribed fields (or the surface parameterizations) were held constant over the annual cycle and soil heat flux G was omitted by setting $C_g = 0$. Sometimes, the prescribed fields could be seasonally varied and they could also be tuned to give acceptable results, frequently in sensitivity studies. For example, Sud and Smith (1985a, b), in using the Goddard Laboratory for Atmospheric Sciences (GLAS) GCM, had prescribed roughness length z_o as 45 cm to represent vegetation, and taken a value of 0.02 cm to represent deforestation and desert. Sud and Smith also prescribed the surface drag coefficient C_D as C_U^2, representing a decrease in the drag coefficient for all deserts (C_U is defined as friction coefficient). Simple prescription of land surface parameters such as this, however, is extremely limited as a research tool because of the following two main reasons:-

i. No feedback effects can be simulated with prescribed surface properties, and
ii. These quantities vary drastically from year to year in many regions, causing difficulties in obtaining realistic climatological fields of these surface parameters.

A.2: The Conceptual Models

The shortcomings of the prescribed field approach were overcome by the implementation of a conceptual model taking care of some feedback effects operating between the land surface and atmosphere. A general outline of their common features is presented here. Comprehensive descriptions on these models have been presented by Carson (1982).

The conceptual model still retains the basic of equations (A.1), (A.2) and (A.3) in the prescribed field approach. Under this conceptual approach, however, <u>the parameters in those equations are made to depend on the land surface condition as predicted by the model</u>. In particular, soil moisture storage is represented as a "bucket" filled by precipitation and emptied by evaporation and runoff, and the governing equation for the soil moisture wetness fraction W is:

$$\frac{dW}{dt} = \frac{1}{\theta_s D}\left(P - \frac{E}{\rho_w} - R_o\right) \tag{A.4}$$

where W = soil moisture wetness fraction (= θ/θ_s); θ = volumetric soil water content, m^3m^{-3}; θ_s = value of θ at saturation, m^3m^{-3}; D = thickness of the hydrologically active soil layer, m; P = precipitation rate, m s^{-1}; ρ_w = density of water, kg m^{-3}; and R_o = runoff rate, m s^{-1}.

Equation (A.4) allows a modeller to explicitly specify some of feedback effects in the surface formulation as permitted in the addition part of the equation. Also, maximum moisture storage of the soil is given by the product $\theta_s D$.

In term of feedback effects in radiation exchange, surface albedo α_l could be made a function of soil moisture, for example:

$$\alpha_l = 0.30 - 0.15W \tag{A.5}$$

Carson (1982) gives a summary of surface albedo used in atmospheric GCM by modellers, and notes that some of them also include snow- and ice-covered surfaces in their parameterization schemes by specifying surface albedo as a function of snow or ice depth.

For latent heat fluxes, the conceptual models also improve the prescribed field approach. The following set of equations (A.6a-d) are modifications of the Penman (1948) expression for evaporation from an open water surface. The most common methodology is to calculate potential evaporation using this formulation:

$$\lambda E_p = \frac{(e_*(T_s) - e_r)}{r_a}\frac{\rho C_p}{\gamma} \tag{A.6a}$$

or

$$\lambda E_p = \frac{\Delta(R_n - G) + [e_*(T_r) - e_r]\rho C_p / r_a}{\Delta + \gamma} \tag{A.6b}$$

where E_p = potential evaporation rate, kg m^{-2} s^{-1}; $e_*(T)$ = saturated vapour pressure at temperature T, mbars; e_r = vapour pressure at reference height, mbars; T_r = air temperature at reference height, K; C_p = specific heat of air, J kg^{-1} K^{-1}; Δ = slope of saturation vapour pressure versus temperature curve, mbar K^{-1}; γ = psychrometric constant, mbar K^{-1}; and r_a = aerodynamic resistance to the turbulent transfer of heat and vapour, s m^{-1} .

Here,

$$r_a \approx \frac{1}{k^2 u_r}\left(\ln\frac{z_r}{z_o} - \phi_1\right)\left(\ln\frac{z_r}{z_o} - \phi_2\right) \tag{A.6c}$$

or

$$r_a = \frac{1}{C_D C_V u_r} \tag{A.6d}$$

where ϕ_2 = nonneutral Paulson (1970) correction factor for vapour transfer; and C_V = vapour or heat transfer coefficient.

The potential evaporation rate given by (A.6) is usually adjusted for the limiting effects of soil moisture by:

$$E = \beta E_p \tag{A.7}$$

where β is a prescribed function of W and it decreases with decreasing W, or a function of the amount of water in a 'bucket'. This function is known as 'β-function'. The value of β varies from 1 (freely available soil moisture) to 0 (dry soil condition).

The combination of equations (A.6) and (A.7) seems to give realistic bounds to the evaporation rate and also to describe its decline with soil moisture depletion. If aerodynamic resistance r_a takes a value for an open-water surface (e.g., a freely ventilated surface covered with well-watered short grass), then (A.6a) and (A.6b) will underestimate the evaporation rate from a saturated natural surface, which is almost always rougher. Alternatively, if an r_a value appropriate to vegetation is used (i.e., with roughness length z_o on the order of a few centimetres to metres), excessively large evaporation rates will be predicted, often exceeding net radiation. As noted by Sellers (1987), while such a large rate would be typical of the evaporation rate from a wetted vegetation canopy, it will differ considerably from the transpiration rate of most natural surfaces, even when W = 1. The potential evaporation rate E_p given by (A.6) may therefore be either that of an open-water surface or of a saturated canopy, but not of a freely transpiring vegetation canopy.

The combination of either (A.6a) or (A.6b) with (A.7) will yield different estimates for the evaporation rate. Using (A.6a), which has T_s (i.e., surface radiative temperature) which is a prognostic variable in most GCMs, the derived value of E_p will be too high or excessive evaporation in arid regions even when W is small. On the other hand, the use of (A.6b) that has T_r (i.e., air temperature at reference height, appropriate of an open-water lysimeter) is more representative in predicting E_p although the predicted surface temperature T_s is not consistent with it.

Monteith (1973) modified the Penman (1948) formula by defining evaporation rate from a vegetated surface in terms of a surface resistance r_c, that is approximately equal to the resistance imposed by all the leaf stomata acting in parallel. The equation, namely Penman-Monteith equation is as follows:

$$\lambda E = \frac{\Delta(R_n - G) + \rho C_P [e_*(T_r) - e_r]/r_a}{\Delta + \gamma \dfrac{r_a + r_c}{r_a}} \quad \text{(A.8)}$$

where r_c = surface or canopy resistance, s m^{-1}. Physically, equation (A.8) is more appropriate in partitioning of energy at the surface than the combination of (A.6) and (A.7). By combining (A.8) with (A.6b), the β function of (A.7) can be expressed as:

$$\beta = \frac{\Delta(R_n - G) + [e_*(T_r) - e_r]\rho C_P / r_{a1}}{\Delta(R_n - G) + [e_*(T_r) - e_r]\rho C_P / r_{a2}} \; \frac{\Delta + \gamma}{\Delta + \gamma(r_{a1} + r_c)/r_{a1}} \quad \text{(A.9)}$$

where r_{a1} = aerodynamic resistance for the vegetated surface, s m^{-1}; and r_{a2} = assumed aerodynamic resistance for E_P calculation in (A.7), s m^{-1}. Therefore, from (A.9), if $r_a = r_{a1} = r_{a2}$, which is an appropriate aerodynamic resistance for the actual surface rather than open water surface (which is much higher), then β is a function of r_a and r_c only. If, instead, r_{a2} is set to the original Penman open-water form, then β depends on all the variables (i.e., (R_n - G), T_r, e_r, r_{a1}, r_{a2}, and r_c) as shown in (A.9) making the conceptual models fail to give satisfactory formulations when using evapotranspiration data sets.

The conceptual models are, therefore, unreliable for prediction of soil moisture and their various parameters have questionable physical significance. In spite of the failings, however, the models can provide some accounting for the past accumulation of precipitation minus evaporation and runoff merely as qualitative indicators of the importance of various surface-related processes. Those equations from (A.5) to (A.9) do describe certain feedback effects from radiation and latent heat fluxes used by early GCMs for sensitivity studies. No feedback effects of surface changes on momentum transfer have ever been incorporated into a GCM simulation using conceptual models (Sellers, 1987).