LAND SURFACE PARAMETRIZATIONS - BASIC CONCEPTS AND REVIEW OF SCHEMES

P R Rowntree
Meteorological Office
Bracknell, England,

Summary: The land surface affects the atmosphere through its radiative characteristics which modify the total energy available, and the surface and subsurface thermal and hydrological characteristics which control the partitioning of energy and water fluxes. The basic concepts involved in these processes are described and parametrizations used in a number of models are reviewed.

1. INTRODUCTION

For more than a decade, evidence has been accumulating from climate and forecast model experiments that the atmosphere is sensitive to variations in processes at the land surface. The experiments of Charney (1975) and Charney et al (1977) demonstrated considerable sensitivity to albedo and the parametrization of evaporation respectively. Accurate initial conditions were shown to be important in the context of predictions on timescales from a few hours or days (Walker and Rowntree (1977), Rowntree and Bolton, 1983) up to several months (Carson and Sangster, 1981). It is evidently important to represent land surface processes realistically; this requires an adequate parametrization scheme, including specification of suitable data sets of vegetation and soil types, and initial soil moisture.

The sensitivity experiments mentioned above, and others reviewed by Mintz (1984), used relatively simple parametrization schemes and lacked geographical variations of most surface characteristics. More complex schemes have however been developed in recent years; firstly geographical variations of a few important parameters such as albedo, roughness length and root depth were introduced (e.g. Hansen et al, 1983); then, more elaborate and hopefully more realistic parametrizations have been developed (Dickinson, 1984; Sellers et al, 1986). As a necessary introduction to the discussion of parametrizations in section 3, I shall first discuss, in section 2, the basic concepts behind the parametrizations. Some uncertainties and outstanding problems are reviewed in section 4.

2. BASIC CONCEPTS FOR PARAMETRIZATIONS OF LAND SURFACE PROCESSES

2.1 Equations for surface and subsurface thermal and moist processes

For a layer of ground between depths z and $z+\delta z$, neglecting horizontal sub-surface transfers of heat and water, and defining z, the ground heat flux G and the water flux M as positive downward,

$$\rho_g\ C\ \partial T/\partial t = -\partial G/\partial z + Q_g \qquad (1)$$

$$\partial m/\partial t = -\partial M/\partial z + N \qquad (2)$$

where ρ_g and C are the soil density and heat capacity, m is the soil water content (mass per unit volume)and Q_g and N are source/sink terms. For the heat budget, the only significant sub-surface heat source is due to moisture phase changes. In the moisture budget, provided we consider m (and M) to refer to the sum of water vapour and liquid water contents (and flux) there are no sources and sinks except those associated with melting and freezing. With this definition, the 'surface' is strictly the interface between vegetation and air, and it is there that the surface boundary condition

$$M(0) = -E(0) + (P + M_s - Y(0)) \qquad (3)$$

applies. The terms in parentheses represent the net contribution of surface and atmospheric processes (rainfall P and snowmelt M_s less surface run-off Y(0)) to the downward water flux at the soil surface, whilst E(0) is the evapotranspiration which may be partly a sink at the soil surface, but in the presence of vegetation also takes water out of the soil throughout the root zone, and transfers it to the atmosphere throughout the canopy.

The surface boundary condition for heat fluxes is

$$G(0) = R_N(0) - H(0) - LE(0) \qquad (4).$$

(see next subsection for definitions).

The sub-surface heat flux G(z) can be represented by a diffusive term of form

$$G(z) = -\ \lambda_g\ \partial T/\partial z \qquad (5)$$

Both heat capacity C and conductivity λ_g depend mostly on moisture content so that changes in m can affect G(z), including G(0).

The sub-surface moisture flux M(z) may be represented (following Dickinson (1984)) in terms of the soil water suction $\Phi = z - \Phi_\omega$, where z is the negative of the gravitational potential or, equivalently the depth in the soil, and Φ_ω is the matric potential which is determined by surface tension and other forces binding water to the soil. Then

$$M(z) = -K(z)\ \partial\Phi/\partial z \qquad (6)$$

Here K is the hydraulic conductivity of the soil.

2.2 Radiative fluxes

(i) Solar radiation

The energy available for partitioning between net upward longwave radiation, the conductive flux into the soil and the turbulent fluxes into the atmosphere is derived from solar radiation. The absorption of solar radiation at the earth's surface can be written as:

$$R_{SN} = R_S\Downarrow - \int_0^\infty R_S\Downarrow_\lambda\ \alpha_\lambda d\lambda = (1-\alpha)\ R_S\Downarrow \qquad (7)$$

Here, α, thus defined, is the surface albedo and $R_S\Downarrow$ is the total downward solar flux, while the suffix λ denotes the corresponding spectral values for wavelength λ. An example of why the spectral variations are important is that, because of absorption by water vapour and droplets in the near infrared, the solar flux below clouds is more intense than with clear skies (relative to the total flux) at wavelengths of less than 0.7μm, for which vegetation is particularly strongly absorbing (α<0.1), but snow is especially highly reflective (α>0.9).

Spectrally averaged surface albedo varies from near 0.1 over forests (Oguntoyinbo (1970), Stewart (1971)) to as high as 0.4 over some sandy deserts (e.g. Rockwood and Cox (1978), Kondratyev et al (1982)) and around 0.9 over fresh snow (Kondratyev et al, 1982). With mean clear sky daily solar radiation of 300 W m^{-2} or more in summer in middle and high latitudes, and all year in the tropics , the sensitivity to the land surface characteristics in the absence of snow can approach 100 W m^{-2} – enough energy to warm the atmosphere below a tropopause at 200 mb by over 1 K/day. Even greater sensitivity is possible with snow, as the albedo of forests is little affected by snow once the branches are snowfree, so that albedo differences of 0.6 are possible between forests and grassland or tundra at middle and high latitudes or on subtropical mountains.

(ii) Longwave radiation

At the temperatures characteristic of the earth's surface, radiation is emitted at wavelengths longer than the approximate upper limit (3μm) of radiation received from the sun. The peak of the emission is at about 11μm, but there is strong emission from about 5 to 50μm. The emissivity of most natural surfaces is near to 1 over most of this range (i.e. the surface is almost a black body). The atmosphere emits over a similar range of wavelengths because it has a similar temperature; however, at some wavelengths, the atmosphere is far from being a black body; this is especially so in the so-called water vapour window near 9-12 μm in which there is only weak water vapour absorption and emission, and some strong but narrow trace gas absorption bands, notably due to ozone.

The net absorbed longwave radiation

$$R_{LN} = \varepsilon\!\Downarrow R_{L}\!\Downarrow - \varepsilon\!\Uparrow \sigma T^{4} \qquad (8)$$

Here, there are two emissivities, defined in a similar way to α in (7), one appropriate for the spectral composition of the downward radiation, one for the upward, near black-body, radiation from the surface, whose intensity depends on the fourth power of the surface temperature. Outside the water vapour window, $R_L\!\Downarrow$ originates from levels close to the ground and so usually differs little from the emission from the earth's surface. It is therefore the emissivity in the window which is important for the surface radiation budget. Data collected by Kondratyev et al (1982) show that this is nearer to unity than the integrated emissivity, being typically 0.96 to 0.98. Generally then, errors in the common model assumption that ε=1 are not likely to exceed about 0.04. This implies an error (with a relatively high net longwave radiation of 100 W m^{-2})

$$\delta R_N \approx 4 \text{ W m}^{-2}$$

Emissivities as low as 0.7 have been estimated for some Saharan surfaces with a large quartz content from satellite data at 9μm by Prabhakara and Dalu (1976). Their data suggest that average 9-12μm values probably exceed 0.8 even for these regions, so that R_N for such surfaces could be increased by as much as 20 W m^{-2}. Errors in the estimation of emissivity are thus usually much less serious than errors in albedo for the calculation of surface fluxes.

2.3 Partitioning of the net radiation

(i) Latent heat flux

The available (net radiative) energy is partitioned between three fluxes; one, the surface soil heat flux has already been discussed in section 2.1(i). We consider next the evaporative or latent heat flux. The evaporation for a homogeneous surface can be written in a number of ways:

$$E = \rho\ w'q' = \rho\ C_E\ V\ \delta q = \rho\ \delta q\ /r_E = \rho\ \delta q\ /\ (r_S + r_A) \qquad (9)$$

Here, w' and q' are small perturbations of vertical velocity and specific humidity respectively, C_E is the surface transfer coefficient for moisture, δq is the atmospheric specific humidity deficit below saturation at a reference height (z_1), commonly 10 m, and r_E is the resistance to moisture transfer. The formal definition of the flux of moisture is given first, then the formulation commonly used in the simpler GCM parametrizations, and then expressions in terms of resistances. Note that for multilayer vegetation, more complicated formulations may be necessary such as those discussed in section 3.4(d).

As shown in equ (9), the resistance r_E can be separated into the surface resistance r_S and the atmospheric resistance r_A. For vegetation, the former normally represents the stomatal resistance to transfer from moist surfaces within the leaves. It decreases to zero when there is water (dew or intercepted rainfall or snow) covering the stomata and also, in the absence of vegetation, with a moist soil surface. The resistance r_A represents that between the leaf or soil surface and the atmosphere and is generally taken to be the same as that appropriate for the sensible heat flux discussed below.

(ii) Sensible heat flux

Energy is also transported by turbulence in the sensible form:

$$H = \rho\ c_P\ w'T' = \rho\ c_P\ C_H\ V\ \delta T = \rho\ c_P\ \delta T\ /\ r_A \qquad (10)$$

Here, T' is a temperature perturbation and δT is the difference between the surface temperature (T_*) and the potential temperature at z_1.

2.4 Typical values of atmospheric resistance to heat and moisture transfer and roughness length

The atmospheric resistance (r_A) to moisture transfer is a function of boundary layer structure and surface roughness through both the transfer coefficient C_E and the wind speed u. To relate r_A to roughness length z_O

and gradient wind speed V_G, it is necessary to use Rossby similarity theory to find u. The dependence of r_A on z_O and V_G for a range of values of coriolis parameter f and for different values of the Rossby similarity theory parameters A and B is discussed in Rowntree (1989).

Estimates of roughness length (z_O) for different surfaces have been reviewed by Garratt (1977) on the basis of the extensive literature on the subject. Garratt notes that values fall in the range

$$0.02 < z_O/h < 0.2 \qquad (11)$$

where h is the height of roughness elements. Garratt reports observed values of z_O ranging from 3×10^{-4} m for desert to 1-5 m for forests. Values of r_A associated with such a range of z_O are presented in Table 1. Note the greater fractional changes in r_A for a factor of 10 change in z_O at large z_O.

It is common in numerical modelling to assume the roughness lengths for temperature (z_T) and water vapour (z_M) to be equal to z_O; indeed, in some models z_O is assumed to be smaller than z_M. However, Garratt suggests $z_O/z_M \simeq 7$ and some studies propose larger ratios. The sensitivity to this assumption is illustrated in Table 1 by comparison of the values of r_A for $z_O = z_M$ and for $\ln(z_O / z_M) = 2$, close to Garratt's proposal. This shows that r_A is higher with the second assumption, with the fractional change a maximum, nearly a doubling, for the largest z_O (1m) quoted.

Table 1: Variations in r_A (sm^{-1}) for selected z_O and geostrophic wind (V_G) (ms^{-1}) ($f=10^{-4}\ s^{-1}$)

(a) $z_O = z_M$

z_O (m)	10^{-4}	10^{-3}	10^{-2}	10^{-1}	1
V_G=20	62.6	44.2	28.8	16.4	6.9
V_G=10	120.7	84.8	55.0	31.1	13.0
V_G=5	232.5	162.5	104.8	59.0	24.4
V_G=2.5	447.2	311.2	199.7	111.7	46.0

(b) $\ln(z_O/z_M) = 2$

z_O (m)	10^{-4}	10^{-3}	10^{-2}	10^{-1}	1
V_G=20	73.5	53.7	37.1	23.5	12.8
V_G=10	141.7	103.2	70.9	44.6	24.2
V_G=5	272.9	197.8	135.2	84.6	45.6
V_G=2.5	524.9	378.8	257.5	160.2	86.0

The calculations in Table 1 are for near-neutral conditions. This is generally appropriate as evaporation mostly occurs with upward energy flux when departures from neutrality are small. It is relevant that Van Zyl and De Jager (1987) found the Penman-Monteith equation, in which these values of r_a may be used, to give good accuracy when compared with lysimeter data, and that the accuracy was not improved by allowing for stability variations.

2.5 Components of the surface moisture budget

A parallel role to that of radiation in the heat budget is played by precipitation in the moisture budget (equ. (3)). Rainfall is deposited on the surface and, after some evaporation directly from the surface, either of a vegetative canopy, if present, or of the soil, can infiltrate the surface. This infiltration is limited to a maximum rate, which is commonly assumed to depend on the vegetation type, and on the soil moisture content. If this maximum is exceeded, surface runoff occurs. An important point here is that since rainfall rate varies over a model gridbox, the maximum infiltration may be exceeded when the gridbox mean is well below the maximum value.

Snowfall accumulates on the surface, and may lose mass through sublimation before eventually melting, when it enters the ground much like rainwater. However, the infiltration characteristics of frozen soil differ substantially from those of unfrozen soil (e.g. Alexeev et al, 1973); the first water infiltrating freezes and unless the latent heat released is sufficient to melt the soil, subsequent infiltration is blocked, leading to surface runoff. The sensitivity of the Meteorological Office (MO) model to variations of this type has been discussed by Mitchell and Warrilow (1987). The magnitude of the effects indicates a need to take proper account of freezing processes in the soil, and also for explicit modelling of snow and the heat transfer through it (e.g. Neeman et al (1988)). Other hydrological aspects are discussed in section 3.

2.6 Typical values of surface resistance to evaporation

As indicated above, the surface resistance, r_s, is zero in the presence of sufficient surface moisture such as when a vegetation canopy is wet. Otherwise, r_s depends on the vegetation type and a number of atmospheric and hydrological variables affecting the supply of and demand for moisture. These include soil moisture distribution relative to root development, solar radiation, atmospheric vapour pressure deficit, wind speed and temperature (see Dickinson (1984) and Stewart (1988) for detailed discussions of the hydrological and atmospheric aspects, respectively and the discussion of Dickinson's treatment later in this paper). Even for the relatively simple case of freely transpiring vegetation, observed values of r_s from different sources sometimes conflict; there appears to be agreement on 40 to 60 s/m for growing crops and 80 to 130 s/m for forests in the absence of low temperatures or high vapour pressure deficits (see Perrier (1982), Thompson et al (1981), Buckley and Warrilow (1988)). Grassland values are more variable - from 60 to 200 s/m can be found.

3. PARAMETRIZATIONS OF LAND SURFACE PROCESSES USED IN GENERAL CIRCULATION MODELS

3.1 Radiation

(i) Solar albedo

Until the early 1980s, the albedo was in most models the only geographically varying land surface parameter, apart of course from orography. It was represented by a single spectrally averaged value, commonly taken from the estimates by Posey and Clapp (1964). These have been criticised for the low value given to tropical forests (e.g. Dickinson, 1980), but were perhaps superior to some of the data sets which replaced them, based on early satellite estimates. More recent satellite data appear more satisfactory, though it is likely that some errors remain. Difficulties which may still be important are the elimination of the effects of small clouds, including their shadow effects, and the problems presented by the shadows of vegetation when the sun is not in the zenith (Franklin, 1988).

The use by Hansen et al (1983) of vegetation data sets prepared by Matthews (1983) allowed the use of typical values of albedo and other land surface parameters for each vegetation type; this approach has also been followed by the MO model using Wilson and Henderson-Sellers (1985)'s data sets, and the Biosphere-Atmosphere Transfer Scheme (BATS) of Dickinson et al (1986), using a combination of the two data sets, together with the classification of ecotypes by Olson et al (1983).

As discussed in section 2.2, there are important spectral variations of albedo. A first step towards including these was taken by Hansen et al with separate specification of albedos for the visible and near infrared; BATS and the SiB (Simple Biosphere) model of Sellers et al (1986) have included a similar breakdown. However, SiB differs from the other models in that it calculates the albedos given solar zenith angle and leaf transmittances and reflectivities for each spectral interval, based on data on the geometry of the canopy. Sellers et al show that a realistic diurnal variation can be obtained.

The effects of snow on albedo are allowed for in all models. However, the sophistication varies considerably. Some early models used a fixed snow albedo for all snow covered land; others included a linear or square root dependence on snow depth, typically with a maximum value of 0.6. A major advance on this was made by Hansen et al (1983), who allowed for the masking effects of vegetation and for snow age:

$$\alpha = \alpha_G + (\alpha_S - \alpha_G)\,[1 - \exp(-d_S/d_S^*)] \qquad (12)$$

where α_G and α_S are the albedos of snow-free ground and snow of infinite depth, d_S and d_S^* are snow depth and the masking depth of vegetation in liquid water equivalent. and α_S is a function of snow age

$$\alpha_S = 0.5 + 0.35 \exp(-A_S/5) \qquad (13)$$

Here A_S is the age in days of the upper snow layer,

$$A_S(t+\Delta t) = \{A_S(t) + (1 - A_S(t)/A_{inf})\,\Delta t\} \exp(-\Delta d_S/d_C) \qquad (14)$$

which tends to an old age limit of 50 days, but is reduced exponentially by new snow, d_C (=0.2 cm) being sufficient to refresh the snow albedo. Typical masking depths are said to range from 0.2 m for tundra and grass to 5 and 10m for deciduous and coniferous forest respectively. These seem too large, and should perhaps be in snow depths not water equivalent.

Dickinson et al (1986) also include a dependence of snow albedo on age, with for example for a cosine zenith angle of 0.5

$$\alpha_V = 0.95(!-0.2A); \qquad \alpha_{IR} = 0.65(1-0.5A) \qquad (15)$$

where α_V and α_{IR} are the albedos for the visible and infrared parts of the spectrum. The age A is defined as $\tau/(1+\tau)$ with

$$\Delta\tau = (10^{-6}\,s^{-1})\,(r_1 + r_1^{10} + d) \qquad (16)$$

where

$$r_1 = \exp(5000\,(1/273.16 - 1/T_{g1}) \simeq \exp(-0.07(273.16 - T_{g1})) \qquad (17)$$

The parameter d is 0.01 over Antarctica and 0.3 elsewhere, presumably to allow for the effects of blowing snow over Antarctica. The term in r_1^{10} is intended to increase ageing when the temperature approaches freezing point. A snowfall of 1 cm water equivalent is assumed to restore the surface age to that of new snow, smaller snow increments being allowed to contribute to the rejuvenation in a linear manner. Masking is treated rather more explicitly than by Hansen et al, the fraction of the grid square covered by snow being calculated as a function of vegetation height - about $20z_o$ - and snow depth. For short grass (z_o = 2cm), 10 cm of snow would give 33% cover, for long grass (z_o = 10cm), 6% cover. The 'short' grass is much closer in character to typical winter pasture.

(ii) Longwave emissivity

Most models have assumed emissivity to be unity. The smallness of the error incurred by this assumption was discussed in section 2.2. Exceptions to the general rule include the GISS Model II of Hansen et al in which spectral dependence of emissivities was included for deserts, snow and ice. Both the BATS and SiB parametrizations assume $\varepsilon = 1$, though the facility to calculate it in a similar way to albedo is included in SiB.

3.2 Soil thermal processes

Most of the models reviewed by Carson (1982) either included no surface heat storage, calculating an equilibrium surface temperature at each timestep, or used a single soil layer to represent soil thermal processes. The heat capacity was calculated so as to provide a realistic diurnal variation, by specifying a heat capacity $(C\lambda_G/\omega)^{1/2}$, where ω is the diurnal period (2π/1 day).

Some models included the dependence of thermal capacity C and conductivity λ_G on soil moisture. It should be noted that inclusion of the dependence of C on soil moisture requires careful consideration if energy conservation is required. Most (?all) models do not consider the heat capacity of precipitation in the atmospheric heat balance, accounting only for the latent heat. It is thus difficult to modify the soil temperature to allow for the infiltration of rainfall into the soil, yet in allowing the heat capacity to vary with moisture content it is implicitly assumed that the precipitation has the same temperature as the soil. A possible justification for neglecting energy conservation in this context (P J

Sellers, p.c.) is that the net surface infiltration, in the long term average, tends to equal the deep runoff, the effects of which on heat content are also neglected . The temperature of this deep runoff will be close to the annual mean surface temperature so that the energy neglected during surface infiltration will approximately balance that neglected during the runoff process. Royer et al (1981) and Hansen et al (1983) included two-layer soil temperature schemes, with moisture dependent C and λ_G. Hansen et al specifically note that changes in thermal properties conserve energy but do not indicate how.

Deardorff (1978) proposed a 'force-restore' scheme to optimise the simulation of surface temperature in a two parameter model, showing that it represented diurnal variations much more realistically than the single layer models. The scheme predicts the surface temperature T_S and the mean (T_M) of T_S over a day or longer.

$$\partial T_S/\partial t = 2G / C\delta_O - \omega (T_S - T_M) \tag{18}$$

$$\partial T_M/\partial t = G / C\delta_M \tag{19}$$

Here δ_O and δ_M are e-folding depths for the daily cycle and another period longer than a day. An extension of the scheme to three layers to allow representation of longer periods has been proposed by Carson (unpublished), while Dickinson et al (1986) use a version which damps the surface temperature towards a prescribed annual mean value.

Warrilow et al (1986) have tested the original force-restore scheme and Carson's modification in a one-dimensional context, with sinusoidal forcing over a range of frequencies. The results were compared with the true solution following the method of Jacobsen and Heise (1982) (Figure 1). The force-restore schemes, though succeeding in simulating variations on the specified timescales, perform rather poorly at other periods, an undesirable feature for realistic data where, for example, there is considerable variability at periods of 10 days to a month (Wallace and Blackmon (1983)).

Fig 2 shows the behaviour of a 4-layer scheme designed by Warrilow to give a more even response. Amplitude and phase are well represented at periods from a day to 2 to 3 years. The other curves in the figure are for two models in which the layer depths increase geometrically with depth (e.g. 5, 20, 80, 320cm). As discussed by Warrilow, low frequency amplification

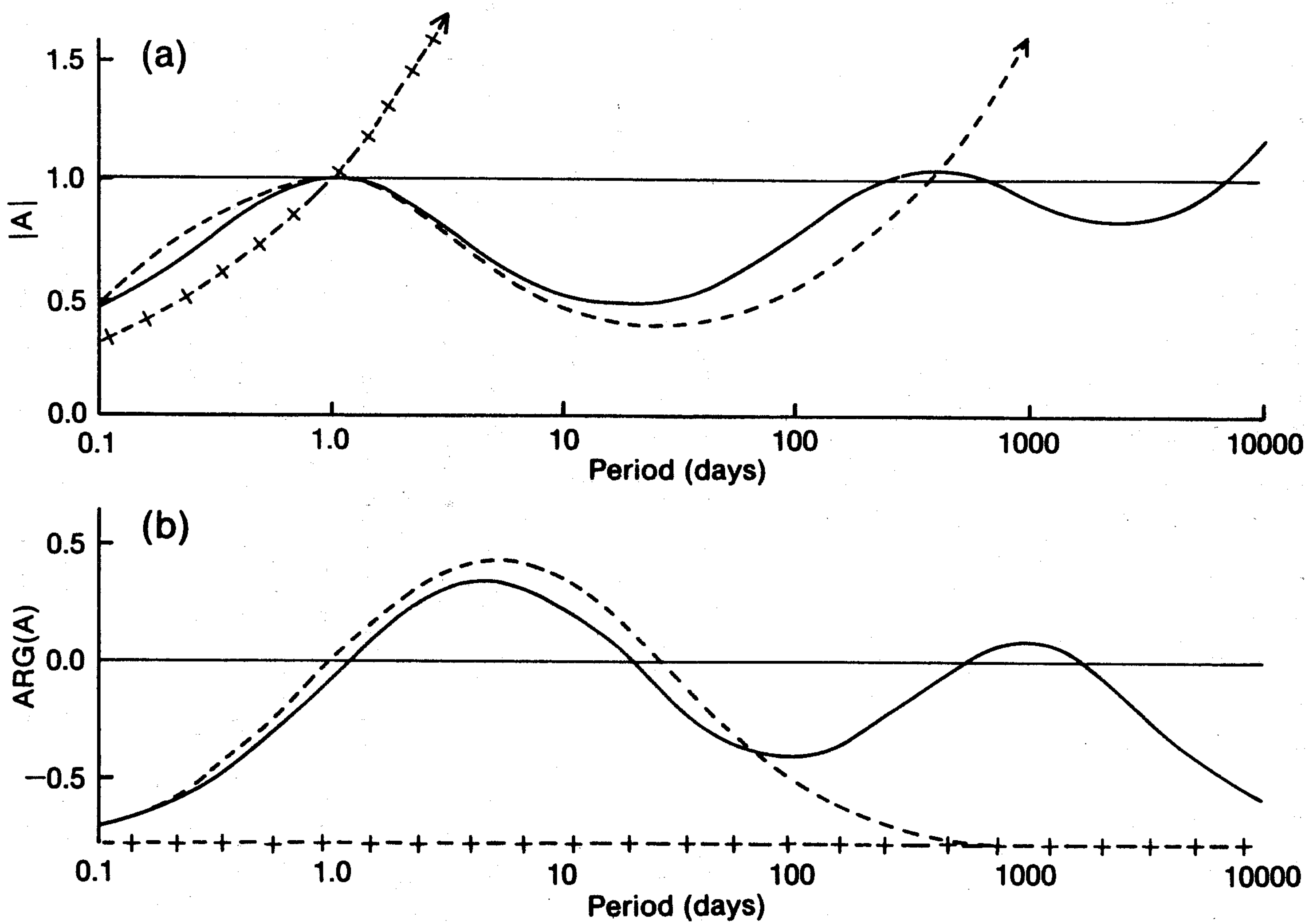

Figure 1 : Amplitude and phase responses of single layer (+ - + -), Deardorff force-restore (- - -) and Carson force-restore (———) schemes.

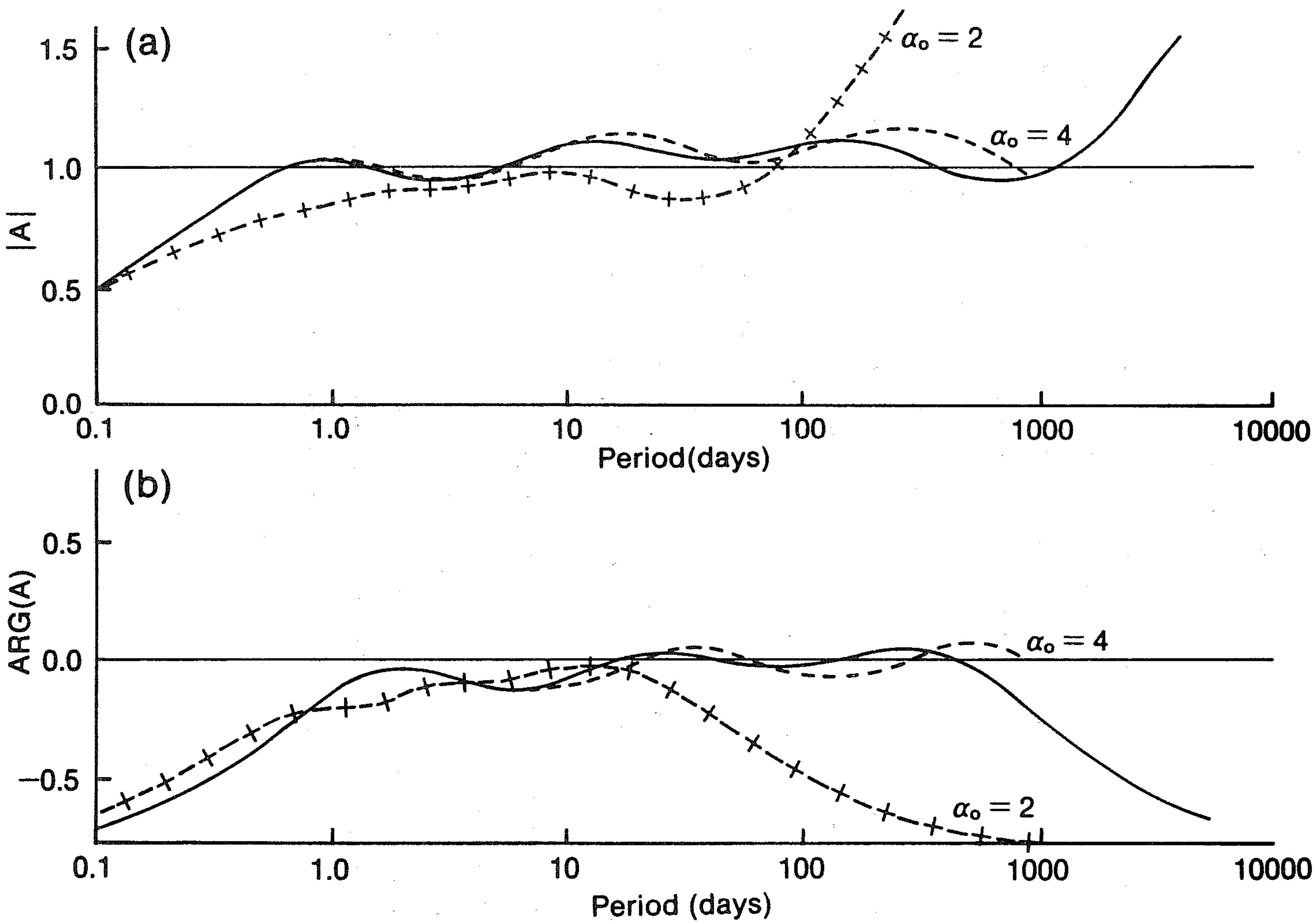

Figure 2 : As Figure 1, but for Warrilow's 4-layer schemes (see text). (α_o is the ratio of each layer's depth to the depth of the layer above).

is characteristic of models with zero flux at the lower boundary. The alternative assumption of a fixed lower boundary temperature is not acceptable for climate simulations. A version of this scheme has been incorporated in the MO fine mesh model, replacing the single slab model, leading to significant improvements in the prediction of night minimum temperatures (Davies and Warrilow, 1986).

In winter, the modelling of soil temperatures becomes more complex for two reasons, the insulating effect of snow and the occurrence of phase changes of water in the soil. Hansen et al (1983) allowed for the former by decreasing the conductivity in the presence of snow, giving the snow insulation properties characteristic of new snow. Warrilow (1989) has used a similar approach, but assumed a higher conductivity and density characteristic of old snow. Neither of these schemes nor BATS attempts to model the snow-soil interface. However, BATS does allow for ageing of the snow and its effect on density and conductivity (assumed proportional to each other), as indicated in section 4.1. BATS also allows for the effects on soil temperature of changes of phase of soil water, all but 15% of the soil water freezing between 0 and -4°C. This range of 4K appears rather large, at least during the melting process.

3.3 Parametrizations of soil moisture

(i) Input terms

The input terms to the surface water budget in eq. (3) are the rainfall and snowmelt less the evaporation. In most models up to the early 1980s, the interception of water by vegetation was not represented, so that the rainfall was assumed to be directly available at the soil surface. Evaporation is discussed in section 3.4.

Dickinson (1984) followed the proposals of Deardorff (1978) and introduced a canopy in BATS, with the part of the rainfall falling on the vegetated area being intercepted and available for evaporation with no surface resistance (see section 3.4). The maximum storage is $0.2L_{SAI}$ in mm, where L_{SAI} is the leaf and stem area index. Excess water is assumed to drip to the ground. Condensation on the foliage is also added to the interception store. The SiB scheme is similar in formulation, but there are two foliage layers, the canopy and ground cover, and the intercepted fraction is dependent on canopy parameters in a similar way to the interception of radiation. In the MO model, the interception capacity is an explicit function of the vegetation type. The interception capacity is given a nonzero value over bare soil to allow for the relatively rapid evaporation of near-surface water; in a model with only one soil layer it would be difficult to allow for this in any other way.

(ii) Infiltration and surface runoff

For many years the standard GCM parametrization of soil moisture was the simple 'bucket' model (Manabe 1969), based on equ. (3); the bucket was of finite depth m_{MAX} and the runoff Y was zero until the bucket was full, after which Y was sufficient to maintain $m = m_{MAX}$. More elaborate treatments designed to allow some runoff for $m < m_{MAX}$ were incorporated in some models. A few models included no hydrology at all; the NCAR CCM simply assumed that evaporation over land was a fixed fraction (0.25) of the potential value. Hansen et al (1983) relaxed some of the constraints, allowing different field capacities (m_{MAX}) for different vegetation types, and including two layers with upward diffusion between them which depended on vegetation and time of year, so effectively representing the transfer of water from the lower layer by roots during the growing season. Runoff from the upper layer was taken as proportional to its water content and to the rainfall, though with a sudden increase when m reached m_{MAX}.

The BATS formulation introduced a more physically-based approach to the parametrization of soil moisture. We saw in section 2 (Equ. (6)) that the moisture flux in the soil can be described in terms of the soil water suction Φ and the hydraulic conductivity K. K is expressed by Dickinson et al (1986), after Clapp and Hornberger (1978), in terms of the fractional soil water content s (ratio of soil water volume to volume of voids in the soil) as

$$K = K_O s^{2B+3} \qquad (20)$$

Similarly Dickinson assumes

$$\Phi_w = - \Phi_O s^{-B} \qquad (21)$$

In these expressions, B is a nondimensional parameter which varies with soil type between about 3 for sand and 11 for clay, and Φ_O and K_O are values at saturation. Φ_O does not vary with soil type in Dickinson's parametrization, but K_O is greater for sand by two orders of magnitude relative to clay. Water is represented in two layers, the surface layer and the total rooting layer, which includes the surface layer.

SiB uses a similar parametrization in a 3-layer model. The moisture in the upper layer is directly available for evaporation into the atmosphere, while the second layer's water can be tapped only by roots. Transfer of water in the third layer is by diffusion and gravitational drainage only.

Surface runoff is parametrized in BATS in terms of the fractional saturation b (soil water density / saturated soil water density) as b^4F where F is the net input of water ($P+M_s-E$) unless the second layer soil temperature is below freezing in which case the larger quantity bF is used to allow for blocking of infiltration by ice. The remainder of the water at the soil surface infiltrates and is added to the upper soil water reservoir. Note that this formulation takes no explicit account of the vegetation type, though the interception of water by the canopy makes some allowance for this. In SiB, the infiltration is zero if the top layer is saturated or the ground surface temperature is frozen, but otherwise is limited only by the saturated hydraulic conductivity K_o.

In reality, rainfall is not distributed evenly over a grid square. In the MO model, an exponential frequency distribution function is assumed for convective rainfall in determining the surface runoff due to the maximum infiltration rate being exceeded, so that

$$Y = P \exp(-\varepsilon F/P) \qquad (22)$$

where F is the maximum infiltration rate and ε is 1 for largescale rainfall and 0.3 for convective rainfall. The maximum infiltration rate is also a function of vegetation and soil types, with

$$F = F_s \left(v \beta_v + 0.5 (1-v)\right) \qquad (23)$$

Here, v is the vegetated fraction, F_s is the bare soil infiltration capacity, dependent on soil type and β_v is the vegetation infiltration enhancement factor, which varies from 2 for crops to 6 for forest. The factor 0.5 allows for crusting of bare soil.

(iii) <u>Percolation and deep runoff</u>

In the 'bucket' formulations, surface and deep runoff were not separated, whilst in Hansen et al's scheme, the lower soil layer could not reach saturation because it received water only when the upper layer was wetter than the lower layer in terms of fractional saturation. Both BATS and SiB parametrize percolation between the layers in terms of the gradient of the soil water potential φ, defined earlier. Deep runoff is the gravitational drainage term which in BATS is $K_0 s^{2B+3}$, giving a large sensitivity to s. For example, for a loam with B=5 and K_0 of 1.3×10^{-5} m/s, deep runoff with s=0.5 is about 0.2 mm/day, while with s=0.6 and the same soil, runoff is ten times greater. The SiB formulation is similar to BATS but allowance is made for a mean slope. The drainage is dependent on the water content of the third layer rather than that of the whole soil represented in the model as it is in BATS.

In the MO scheme, deep runoff is a function of the soil hydrological properties

$$Y_G = K_0 \; ((\theta-\theta_w)/(\theta_0-\theta_w))^c \qquad (24)$$

This formulation is based on Eagleson (1978), where c varies from 5 for sand to 11 for clay. θ_w is the soil moisture at wilting point for the vegetation and θ_0 the saturation value of θ, the soil moisture volume concentration.

3.4 Turbulent fluxes

(i) General

The turbulent fluxes have generally been formulated using transfer coefficients based on Monin - Obhoukov similarity theory

$$C_D = C_{DN} \; F(Ri) \qquad (25)$$

where

$$C_{DN} = [\; k \; / \; \ln \; (z/z_0)]^2 \qquad (26)$$

is the neutral drag coefficient and Ri is Richardson number (k is the Von Karman constant). The stability dependence takes various forms based on different boundary layer studies. Mahrt (1987) has pointed out that the dependence should be weaker for gridbox average fluxes than for point fluxes. The transfer coefficients for heat and for moisture in moist conditions (see section on evaporation for a more general discussion of moisture transfer) have been related to C_D either by another Ri-dependent relation or, in simpler models, by assuming equality or a constant proportionality, e.g. $C_{HN} = 1.35\; C_{DN}$ in the Hansen et al (1983) model. This latter assumption is equivalent to assuming that the roughness lengths for heat and moisture are larger than those for momentum. This is contrary to the relation proposed by some writers on the boundary layer as discussed in section 2.4.

In many early models, as discussed by Carson (1982), the stability and even in some cases the z_o dependences were ignored. Also, in the absence of vegetation datasets, the spatial variations of z_o were neglected, a constant z_o (or C_D where stability dependence was ignored) being assumed for all land (even in some models for all land and sea!). This did not allow the high surface temperatures which are observed over smooth land surfaces (W. T. Roach, personal communication. Hansen et al introduced vegetation dependent z_o and have been followed in this by the BATS, SiB and MO models. In BATS, the neutral drag coefficient is a linear combination of those for vegetation, bare soil and snow according to the fractional coverages of the grid square. Allowance is made for the snow cover according to its depth, based on the assumption that half the vegetation is covered by snow for snowdepth of $10z_o$, and half the bare ground with depth of 1cm. C_D for snow is the same as for ocean. In the MO model, the mean z_o for the gridbox is obtained from that for the component parts of the gridbox by averaging $(1 / \ln(l/z_o))^2$, where $l=550L$ in metres, L being the latitudinal gridlength in degrees. SiB uses the characteristics of the vegetation canopy - fractional area coverage, leaf angle distributions, canopy top and bottom heights and stem and leaf area indexes in diffusion equations to describe the absorption of momentum by the canopy and ground. Thus, z_o is a derived rather than a prescribed parameter. The parameters are modified if the depth of snow becomes significant compared to the height of vegetation.

(ii) Heat flux

This follows the formulation for momentum in most respects except where slightly different transfer coefficients are used as indicated above. In BATS and SiB the representation of a canopy as described below for evaporation also affects the formulation for heat flux to a more limited extent.

(iii) Evaporation

Parametrizations of evaporation used in general circulation models (GCMs) have been reviewed by Carson (1982) and, updated where appropriate, by Rowntree (1984). Most of the parametrizations of evaporation reviewed at that time were based on eq. (9) in the form:

$$E = \rho\ C_E\ V\ \delta q \qquad (27)$$

with C_E approximated by $\beta\ C_{EO}$; C_{EO} is the transfer coefficient when r_s =

0, generally taken as a constant, with the same value for all land. β is a soil moisture dependent parameter varying between 1 in moist ($r_s = 0$) and 0 in arid conditions. Comparison of equ. (27) and (9) shows that

$$\beta = r_A / (r_s + r_A) \qquad (28)$$

Typically, however, β was taken as a linear function of the soil moisture m. A more realistic parametrization similar to (28) was tested by Laval et al (1984), giving substantial decreases in evaporation. Note that (27) may also be written

$$E = \beta E_P \qquad (29)$$

Here E_P is at first sight the evaporation for a moist surface, commonly called the potential evaporation. However, this is a misleading interpretation in that the saturation mixing ratio q_s in the term δq ($\equiv q - q_s$) in (27) is calculated for the surface temperature which in dry conditions will be considerably enhanced due to the weakness of evaporative cooling; observationally, on the other hand, E_P is usually estimated for a moist surface as in the work of Priestley and Taylor (1972) which might otherwise justify the use of (29). Mintz and Serafini (1981) proposed an alternative approach in which a separate wet surface energy balance was used to compute the surface temperature needed for the calculation of E_P.

(iv) <u>More elaborate representations of turbulent energy fluxes</u>

By introducing the explicit control of evaporation by soil and plant processes, and including the role of the foliage and canopy following the work of Deardorff (1978), Dickinson (1984) has placed the parametrization of evaporation on a more scientific basis. More recently, Sellers et al (1986) have developed a scheme based on similar principles, though differing in many detailed features. In this section, I shall describe Dickinson's scheme, the Biosphere-Atmosphere Transfer Scheme (BATS), since it is the earlier of the two, and then mention some of the differences between the two.

Dickinson uses equ (9) to calculate evaporation with the resistances r_s and r_A. Separate calculations are made for evaporation from the wet parts and transpiration from the dry parts of the canopy, and for evaporation from the soil. The total is calculated by weighting these estimates by the fractions of the gridbox occupied by each, based on the prescribed bare soil fraction for the surface type, and an estimate of the wet fraction of the canopy, based on the water calculated to be held by the canopy following dew or interception of rainfall. The maximum canopy water content is proportional to the leaf and stem area L_{SAI} defined below.

The bulk stomatal resistance is expressed as

$$r_S = r_{SMIN} R_L S_L M_L / A_D < r_{SMAX} \qquad (30)$$

Here r_{SMIN} is the prescribed minimum stomatal resistance factor for the vegetation specified for the gridbox, (250 s/m, except 150 for mixed farming);

r_{SMAX}, the upper limit on r_S, is 10,000 s/m;

R_L represents the dependence of r_S on solar radiation, with minimum value of 1 for overhead sun;

S_L is a temperature dependence, equal to $(1-0.0016\ (298-T)^2)^{-1}$;

$A_D = (\sigma_F L_{SAI} L_D)$ represents the area of the transpiring surface, with σ_F the fraction of the gridbox covered by vegetation, L_{SAI} the area of leaf and stem per unit area of gridbox, and L_D the unwetted fraction of this free to transpire; typical values of these quantities used by Dickinson lead to maximum values of A of about 5 for forests and savanna, 3.6 for tundra, 1.6 for short grass, 0.6 for semi-desert and 0 for desert.

M_L is used to restrict transpiration to the maximum that the soil/root system can supply, having a minimum value of 1; the transpiration cannot exceed the plant's ability to supply water through the roots and stems. This may be seriously limited by soil moisture deficits. Dickinson et al (1986) assume that transpiration is restricted to the product of a maximum value (of order 1.5×10^{-7} m s^{-1}, or 0.54 mm/hour) and a term $(1 - W_{LT})$ dependent on soil water potential for each model layer where

$$W_{LT} = (s^{-B} - 1)/(s_w^{-B} - 1) \qquad (31)$$

Here s_w is the soil water content for which transpiration becomes zero - some water being unavailable. The value of s_w increases from near 0.1 for sand to over 0.5 for clay. The dependence of W_{LT} on s through (31) is illustrated by the following values for a typical value B = 5 with the value of s_w (0.125) suggested by Dickinson (1984).

Table 2: Variation of W_{LT} with soil moisture s for B = 5, s_W = 0.125

s	1	.8	.6	.4	.3	.2	.175	.15	.14	.13	.125
W_{LT}	0	.0001	.0004	.003	.013	.095	.19	.40	.57	.82	1.0

Note the small effect of variations in s for s > 0.3. The abrupt decrease in water availability over a small range of s is consistent with the observations summarized by Priestley and Taylor (1972). However, it may not be appropriate to apply such a sharp cutoff to a model gridbox of scale 100km or more with considerable inhomogeneity in several respects (rainfall, soil and vegetation type, slope, etc).

The atmospheric resistance is r_A where

$$r_A^{-1} = (A_W\ C_F\ (U_{AF}\ /\ D_F)^{1/2}) \qquad (32)$$

Apart from the A_W, which corresponds to A_D, but for the whole surface, so including the non-transpiring part of the vegetation, this is an expression for laminar boundary flow. For most surfaces, it is 0.05 $U_{AF}^{1/2}$, where U_{AF} is the wind speed within the canopy, modelled as u $C_D^{1/2}$, where u is the windspeed at anemometer level. For forest, with z_O = 1m, C_D (at 10m) is 0.03 from (26); the typical windspeed in Shuttleworth et al (1984)'s data is u = 2 m/s, which is consistent with calculations of u using Rossby similarity theory (Rowntree, 1989) with V_G = 10 m/s. With A_W typically about 8 for forests and savanna, $r_A \simeq 4$ s/m, which is below the estimates for forest of about 15 s/m with geostrophic winds of 10 m/s in Table 1, even with $z_O = z_M$. Shuttleworth (1988) estimates values of 34.2/u for r_A in Amazonian rainforest, though with some uncertainty. There appears to be scope for further investigaton in this area, though, as noted by Shuttleworth, estimates of forest transpiration show only limited sensitivity to r_A (hence the uncertainty of observed estimates). With wet vegetation, the sensitivity is greater because r_S=0. Here Shuttleworth suggests it is adequate to assume $z_O = z_M$, for which the results given in Table 1 apply.

The formulations of energy exchange in BATS depend on the differences between foliage characteristics (temperature T_F and saturation specific humidity $q^{SAT}{}_F$), and the corresponding values for the canopy air space (T_{AF}, q_{AF}); these latter have to be determined. The canopy air is assumed to have negligible energy capacity so continuity of fluxes can be assumed:

$$H_A = H_F + H_G \quad (33)$$

where suffixes A, F, G, refer to the flux to the atmosphere, from the foliage to the canopy air and from ground to canopy air respectively. The latter two depend on differences between the appropriate surface and the canopy air, the first on differences between canopy air and bottom model atmospheric layer. The continuity assumption allows solution of the equations for the canopy air space characteristics. The transfer coefficients from ground to canopy air and from canopy air to air above are assumed to be the same as for momentum, though the wind speeds within the canopy (described above) are used for the ground to canopy air fluxes over the vegetated fraction of the gridbox.

Finally, in this discussion of BATS, it is useful to note that there are 14 specified parameters for each of the 18 land cover/vegetation types. These are maximum vegetation cover and its variation with temperature, roughness length, depths of the total and upper soil layers, root distribution between the layers, albedos for $<$ and $>$ 0.7μm, minimum r_s, maximum and minimum leaf area indexes, a stem and dead matter area index and measures of leaf dimension and light sensitivity. For soil, for each of 12 texture classes, there are six parameters (porosity, maximum soil suction, saturated hydraulic and thermal conductivities, and B and s_w as defined in sections 3.3(ii) and 3.4(iv) respectively); additionally, for each of 8 colour classes, there are dry and wet soil albedos for $<$ and $>$ 0.7μm (see Dickinson et al, 1986, for the actual values).

The formulation of evaporation in the SiB or simple biosphere model of Sellers et al (1986) is similar in concept to the BATS scheme, while differing in detail. Some differences are the inclusion of a linear dependence of r_s on vapour pressure deficit, and the more complicated form of the limitation of evaporation by soil moisture, though the fundamental parametrization of its dependence on s is similar. The Meteorological Office (MO) model (Warrilow et al (1986), Warrilow (1989)) occupies an intermediate position on the scale of complexity. The evaporation is calculated assuming $E = \beta E_w$, where E_w is the evaporation allowing for minimum stomatal resistance, defined according to the vegetation type, but with the limitation by soil moisture represented by $\beta = m/(5cm) \leq 1$. Eight vegetation-dependent and eight soil-dependent parameters have geographical variations based on the Wilson and Henderson-Sellers data sets.

4. UNCERTAINTIES

There remain many uncertainties in the parametrization of land surface processes for climate models. The specification of data sets of land surface characteristics is a major problem and requires considerable work on the interpretation of satellite data. This task is in progress as part of the World Climate Research Programme under the aegis of the ISLSCP (International Satellite Land Surface Climatology Project), but there are many difficulties, not least the specification of soil characteristics.

Problems with the parametrizations themselves are also numerous. At present, all schemes effectively assume a single value for each parameter for each model gridbox, even though this value may be calculated as a mean based on the distribution of vegetation types in the box. This basically linear assumption makes little allowance for the nonlinear effects of inhomogeneities such as: (i) the spatial variations of moisture availability on the scale of kilometers or less between swampy valleys and well-drained slopes or between forest and ploughed fields in a drought; (ii) the early melting of snow on the slopes that receive the most solar radiation; (iii) the increased cloudiness and so reduced evaporation in areas which have most precipitation - whether through chance on a diurnal scale over uniform terrain or systematically on long timescales over uplands. Some models (SiB, MO) do allow for the spatial variability of convective precipitation and the consequent enhancement of surface runoff, but this is perhaps the only example of parametrizing to allow for the enormous contrast between the observational scale of a few metres and the modelling scale of 100 km upward.

Other uncertainties include the relation between z_O and z_M discussed in sections 2 and 3, and the associated problem of representing orographic roughness. The simulation of snowcover in wooded terrain may be adequately parametrizable using models which explicitly represent the canopy, but the effects of drifting in allowing an early return to snowfree evaporation conditions are nowhere represented. The representation of freezing processes in the soil and the infiltration of snowmelt are other potentially important topics which are at best crudely parametrized.

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