CHAPTER 7: THE LAND SURFACE SCHEME OF THE UNIFIED MODEL AND RELATED CONSIDERATIONS

7.1: Introduction

The atmosphere is sensitive to variations in processes at the land surface. This was shown in the earlier modelling experiments by Charney (1975) and Charney *et al.* (1977) which demonstrated considerable atmospheric sensitivity to albedo and the parameterization of evaporation respectively. As also indicated in the major deforestation experiments using GCMs reviewed in Chapter 4, deforestation is said to alter various land surface properties, producing considerable impacts on climate. In term of the atmospheric dynamics that control day-to-day weather, the surface exerts its influence on the free atmosphere through the atmospheric boundary layer. This can be from a few tens of metres to one to two kilometre deep depending on the stability (Smith, 1993) which determines the intensity and depth of the turbulent transport of momentum, heat and moisture.

In modelling, prediction of variables characterising the thermodynamic and hydrological state of the earth's surface is important since most human and other biological activity takes place at or very near the surface. A model, therefore, must represent surface processes, which interact with those in the atmosphere on all space and time scales. Subsurface thermodynamics and hydrology, as well, must be modelled to predict surface quantities.

In this chapter, a summary will be given of the land surface scheme of the UK Meteorological Office Unified Model (UM) employed in this study. In this discussion, however, only key mathematical formulas are quoted; the scheme's complete mathematical formulation, as coded in the UM, is given in detail in the UM documentation papers by Gregory *et al.* (1994) and Smith (1993). Emphasis is placed on those aspects that are important for the change of vegetation parameters that are involved in deforestation experiments undertaken in this study. Those parameters are surface roughness lengths, surface "stomatal" resistance for evaporation, root depth, surface "canopy" capacity, vegetation fraction, infiltration "enhancement" factor and snow-free albedo. Other related quantities relevant to land surface properties are also discussed.

In most models, the hydrology of the land surface is generally represented by predicting soil moisture and snow depth using one or more layers (Manabe, 1969). Surface temperature is either determined diagnostically assuming a local radiative balance, or computed using a prognostic equation for one or more layers of soil (i.e., a multilayer). In the Unified Model, a multilayer model (Smith, 1993) for both soil temperature and soil moisture is included. This is a four-layer soil thermodynamic model with layer depths chosen to optimised the response to forcing on time-scales from 1 day to 2-3 years. From this multilayer model, land surface temperature is computed from the balance between the net heat flux from the atmosphere and the surface soil heat flux. Snow depth is increased by snowfall and reduced by sublimation and snowmelt. Snow insulation is represented by reducing the thermal conductivity between the top two soil layers.

The surface-hydrology scheme in the Unified Model includes a model of the vegetation canopy (Gregory *et al.*, 1994). Moisture can be retained in the canopy or transferred to the soil or atmosphere. The canopy is capable of intercepting some of the incoming rainfall (Warrilow *et al.*, 1986). The spatial inhomogeneity of rainfall within a grid box is allowed for by specifying an exponential distribution of local rainfall rates (Dolman and Gregory, 1992). Convective rainfall is assumed over 30% of the canopy (cf. 100% for dynamical rain) and water stored in the canopy evaporates at the potential rate. Both throughfall (i.e., rainfall that is not intercepted by the canopy) and snowmelt would augment the soil moisture content. The surface runoff occurs when the throughfall or snowmelt rate exceeds the maximum infiltration rate of the soil and slow gravitational drainage from the root zone. The total runoff, which is the sum of surface runoff, increases with the soil moisture content. Different vegetation and soil types can be specified. The soil types are used to determine the surface albedo as a component to calculate the local radiative balance. For example, for most applications of the Unified Model that have been reported, the spatially varying specification of vegetation and soil properties for the land-surface scheme is based on a 1^{o} x 1^{o} climatology (Wilson and Henderson-Sellers, 1985).

The above-summarised scheme, which is in used by the Unified Model, is described in detail by Smith (1993) and Gregory and Smith (1994).

7.2: Subsurface, Surface and Boundary Layer Processes

Processes within the subsurface, surface and boundary layer in the UM comprise of a number of major components, which interact with each other by passing calculated variables when appropriate. The six major components involved in the UM are: (i) sea-ice thermodynamics; (ii) soil thermodynamics; (iii) turbulent surface exchange and boundary layer mixing coefficients; (iv) the surface and boundary layer equations and their implicit solution; (v) adjustment of the surface evaporation and sublimation and the surface temperature increment; and (vi) boundary layer cloud. A summary of each of the components, apart from the sea-ice and snow components which are not directly relevant to this study, will be given in turn in the following paragraphs.

7.2.2: Soil thermodynamics

The soil thermodynamics component calculates the heat flux through the soil at land points. Heat transport through the soil is modelled with a multilayer scheme in which this component updates the "deep" soil temperatures, but not the top soil layer temperature. The heat flux between the top two soil layers is output for use in the implicit calculation of the increments to the top soil temperature and boundary layer variables.

In the multilayer soil thermodynamics model, the soil heat flux, H_s, and the rate of change of soil temperature, T_s, are derived from the continuous equations. When both z and H_s are defined to be positive downwards, the continuous equations involved are

$$H_s = -\lambda_s \partial T_s / \partial z \tag{7.1}$$

and,

$$C_s \partial T_s / \partial t = -\partial H_s / \partial z \tag{7.2}$$

where Σ_s is the thermal conductivity of the soil (units: $Wm^{-1}K^{-1}$) and C_s is the volumetric specific heat (or heat capacity) of the soil (units: $Jm^{-3}K^{-1}$). These two parameters are climatologically prescribed, geographically varying quantities depending on the soil type. The soil ancillary file used in this study includes the two parameters, which are derived from the global archive of land cover and soils data. The UM Documentation Paper No. 70 (Jones, 1995) describes their derivation from the Wilson and Henderson-Sellers (1985) global archive of land cover and soils data.

At the top boundary of the soil model, the downward heat flux is the sum of the net radiative flux and the turbulent sensible and latent heat fluxes at the surface, (i.e., $R_{N\downarrow *}$ - H_* - LE_*). The turbulent fluxes are defined to be positive upwards. At the bottom boundary, the heat flux is assumed to be zero, as it need to be for climate simulations.

A predictive equation in the UM (equation for the rate of change of the deep soil temperatures from top layer downwards) is obtained by discretizing with respect to z and combining (7.1) and (7.2) [see Smith (1993) for details]. There are two coefficients in the formulation, which are defined in terms of the soil layer thickness $-z_r$ (r = layer number) and the soil diffusivity $P_s=\Sigma_s/C_s$. The heat flux between soil layers r and r+1 is then obtained in term of normalised layer thickness N_r (= $-z_r/-z_1$). The $-z_1$ is chosen to be the depth of a temperature wave characteristic frequency α_1 which satisfies the continuous equation, (7.1) and (7.2).

The ratio of the multilayer model's surface temperature response to that of the analytic solution to the continuous equations is a function of the normalised frequency α/α_1 and the normalised layer thicknesses N_r. Therefore, the main concern when choosing the number of soil layers and the value of the arbitrary parameters of the scheme (such as the layer thickness $-z_r$) is to ensure that the (complex) ratio of the finite difference scheme's solution for the surface temperature to the analytic solution has amplitude close to unity and phase close to zero for a range of surface forcing frequencies occurring in nature. For the soil model in the UM, it has been found that a minimum number of four soil layers is required to give a good amplitude and phase response to forcing periods in the range half a day to a year (Smith, 1993). In this study, four deep soil layers (or five soil layers, including the surface) was chosen in all the experiments undertaken. The values of the α_1 and N_r are set by the model for all gridpoints involved in any simulation.

A representation of snow insulation is also included in the soil thermodynamics component. The effect of snow lying on the ground insulating the soil below is also represented, simply by reducing the thermal conductivity between the top two subsurface layers.

7.2.3: Turbulent surface exchange and boundary layer mixing coefficients

The turbulent surface exchange and boundary layer mixing coefficients component treats the processes that determine the turbulent fluxes of momentum, heat and moisture at the surface and through the boundary layer. This is the centre of the UM's land surface scheme since the implementation in the code of this component involves all the subroutines from the other components.

It is a standard practice to represent the mean vertical surface turbulent flux, F_{x*}, of any conservative quantity X by

$$F_{x*}/\rho_* = \overline{(\omega' X')}_* = -c_x |v_1 - v_0| (X_1 - X_0) \tag{7.3}$$

where ω' and X' define F_{x*} in conventional turbulence notation (i.e., in terms of the surface eddy covariance of X and and the vertical velocity component ω; the Ψ_* is the atmospheric density at the surface; v_1 is the mean horizontal wind at the lowest model level; v_0 is the velocity of the surface (identically zero for land points but equal to the surface ocean current at sea points); $X_1$and X_0 are the values of X at model level 1 and at the surface respectively; c_x is the turbulent surface exchange (or bulk transfer) coefficient. The turbulent surface exchange coefficient, c_x, is a function of atmospheric stability, surface roughness and other parameters characterising the physical and physiological state of the surface and any vegetation.

For (7.3) to be a good approximation in the specification of the bulk transfer, the bottom model level is assumed to be within the "surface flux" layer, which is typically a few tens of metres in depth. The UM allows the choice of scheme with either only one model layer or more than one model layer for the boundary layer. In one model layer or the "local mixing" scheme, the turbulent flux, F_x, of a conservative quantity X is parameterized using a first-order turbulence closure

$$F_x / \rho = \left(\overline{\omega' X'}\right)_* = -K_x \, \partial X / \partial z \tag{7.4}$$

where K_X is the turbulent mixing coefficient for X which is in general a function of a mixing length, the local wind shear and atmospheric stability. The rate of change of the quantity X due to turbulent mixing (TM) is then

$$\left(\partial X / \partial t\right)_{TM} = g \, \partial F_x / \partial p \tag{7.5}$$

By choosing more than one model layer, the model allows non-local mixing of heat and moisture in unstable conditions. In unstable, rapidly mixing regions, the fluxes are not in fact closely related to local gradients since eddies or plumes which are doing the mixing have large vertical extent and correlation. Forming flux divergences over relatively thin model layers (particularly the lowest model layer) also causes numerical instability problems when the timestep is large and the turbulent mixing coefficients are large.

The boundary layer of more than one model layer uniformly distributes the heating and moistening resulting from the divergence on the fluxes between the surface and the top of the boundary layer. The surface fluxes, therefore, are given by (7.3) and the top-of-boundary-layer fluxes are given by (7.4).

The diffusive nature of the equations requires the numerical time-stepping scheme to be chosen carefully to ensure numerical stability of the solution for reasonably large timesteps. In this study, the five model layer scheme was chosen to cover the boundary layer with the 30-minute time stepping used. This setting has already been tested in the UK Meteorological

Office, and shown not to give stability problems with the implicit numerical scheme used in the UM.

7.2.3.1: The surface fluxes of momentum, heat and moisture

The interaction between the atmosphere and the underlying surface is driven by the realisation that the surface fluxes of momentum, heat and moisture determine to an important degree the steady state of the atmosphere. The atmospheric part of this interaction takes place in the boundary layer near the surface layer, where the Monin-Obukhov hypothesis applies. The Monin-Obukhov similarity hypothesis is a generally accepted framework for describing the surface layer and virtually all numerical models make use of it in one way or the other.

Equation 7.3 is used in various forms by replacing the conservative quantity X with the actual variables in the atmosphere, producing their associated surface fluxes. Based on (7.3) the surface fluxes of momentum, heat and moisture can be obtained as follows:

(i) Momentum: If X is set to the vector horizontal wind, v, the surface turbulent flux of momentum is obtained. Conventionally the surface stress, $\therefore_*$, is defined to be the downward momentum flux at the surface to replacing ($-F_*$) and a drag coefficient, c_D, to replace c_x as the new bulk transfer coefficient. For land points v_0 is identically zero but for ocean points (including those with sea-ice) v_0 is the ocean surface current. In configurations without coupling to an ocean model (as is used in this study, see Chapter 8), these quantities are climatological values, though they can also be from analysed values. They are predicted if the atmosphere and ocean models are coupled.

(ii) Heat: if X is set to the liquid/frozen water static energy, $s_L = C_pT_L + gz$, the surface fluxes of sensible heat, H_*, is obtained with a bulk transfer coefficient c_H involved.

(iii) Moisture: if X is set to total water content, q_w, the surface flux of moisture, E_*, is obtained. Similar to (ii), the bulk transfer coefficient, c_H, is also involved with a factor, $\bar{\ }$, included when assuming $q_* = q_{sat}(T_*, p_*)$.

New terms, then, can be defined in terms of the surface fluxes using the above three replacements of X, which are useful for describing the dependence of the bulk transfer coefficients, c_D and c_H, on the surface layer stability and surface parameters. They are (a) surface friction velocity, v_s ; (b) surface layer scaling parameters for temperature, T_s ; (c) surface layer scaling parameters for moisture, q_s ; (d) surface buoyancy flux, F_B ; (e) surface layer buoyancy scaling factor, B_s ; and (f) the important length scale called the Monin-Obukhov length, $L_s = v_s^2 / (kB_s.)$, where k is von Karman constant (k = 0.4).

7.2.3.2: The bulk transfer coefficients for momentum, heat and moisture

As mentioned above, the surface fluxes are given in terms of the bulk transfer coefficients, c_D and c_H. These coefficients, actually, depend on the surface layer stability and surface parameters. The UM uses the most widely accepted approach, the Monin-Obukhov hypothesis, for relating surface layer gradient wind, temperature and moisture to their corresponding surface turbulent fluxes. The approach is applicable strictly for a fully turbulent surface layer, under stationary and horizontally homogeneous conditions. Under the hypothesis, the surface layer gradients of wind, temperature and moisture can be written in term of the scaling parameters, v_s, T_s, q_s, and L_s, as follows:

$$\frac{\partial v}{\partial z} = \frac{v_s}{kz}\Phi_m(z/L_s) \qquad [v = v_0\ (=0 \text{ for land points}) \text{ at } z = z_{0m}] \tag{7.6}$$

$$\frac{\partial T_L}{\partial z} + \frac{g}{c_p} = \frac{T_s}{kz}\Phi_h(z/L_s) \qquad [T_L = T_* \quad at \quad z = z_{0h}] \tag{7.7}$$

$$\frac{\partial q_w}{\partial z} = \frac{q_s}{kz}\Phi_h(z/L_s) \qquad [q_w = q_* \quad at \quad z = z_{0h}] \tag{7.8}$$

where $>_x$ is a **universal similarity function** of z/L_s. The universal similarity function in principle may be different for each transferable quantity X and which has to be determined from analysis of surface layer data. However, it has been assumed that the similarity functions for sensible heat and moisture are the same. The equations are bounded by the lower boundary conditions as stated on their right. The z_{0m} and z_{0h} are the surface roughness lengths for momentum and sensible heat respectively. The roughness length for moisture has been assumed to be equal to that of heat, although in principle it may be different. It is a characteristic that the roughness length for momentum might be much larger than the other two.

It is convenient to define the model's height coordinate origin at the height where $v = v_0$ by making the transformation $z' = z - z_{0m}$, modifying the terms in (7.6), (7.7) and (7.8) so that the lower boundary conditions for the modified forms are now as follows:

$$\left.\begin{aligned} v &= v_0 && \text{at } z' = 0 \\ T_L &= T_* && \text{at } z' = z_{0h} - z_{0m} \\ q_w &= q_* && \text{at } z' = z_{0h} - z_{0m} \end{aligned}\right\} \quad (7.9)$$

Integrating modified forms of (7.6), (7.7) and (7.8) from the respective lower boundary conditions to the bottom model level at $z' = z_1$ (i.e., the height of the first model level), and substituting for v_s, T_s and q_s, will give the expressions for c_D and c_H in terms of the functions Φ_m and Φ_h:

$$c_D = \left(k / \Phi_m(\varsigma_1, \varsigma_{0m})\right)^2 \quad (7.10)$$

$$c_H = \left(k/\Phi_m(\varsigma_1, \varsigma_{0m})\right)\left(k/\Phi_h(\varsigma_1, \varsigma_{0h})\right) \quad (7.11)$$

where $\varsigma_{0m} = z_{0m} / L_s$, $\varsigma_{0h} = z_{0h} / L_s$ and $\varsigma_1 = (z_1 + z_{0m}) / L_s$. From (7.10) and (7.11), it can be seen that the exchange coefficients will depend, among other things, on their related roughness lengths.

As surface layer buoyancy scaling factor $B_s \to 0$, neutral conditions are approached and $\varsigma \to 0$ for all non-zero, finite z. It is known that the similarity function, Φ_x, approaches unity in this limit, and c_D and c_H can be defined for neutral conditions by

$$c_{\mathrm{D}} = c_{\mathrm{DN}} = \left(k \big/ \ln\left(z_1 + z_{0\mathrm{m}}\right) / z_{0\mathrm{m}}\right)^2 \tag{7.12}$$

$$c_{\mathrm{H}} = c_{\mathrm{HN}} = \left(k \big/ \ln\left(\left(z_1 + z_{0\mathrm{m}}\right) / z_{0\mathrm{m}}\right)\right)\left(k / \ln\left(\left(z_1 + z_{0\mathrm{m}}\right) / z_{0\mathrm{h}}\right)\right) \tag{7.13}$$

The quantity ς_1 is a non-dimensional measure of the stability of the surface layer. It is, however, not convenient for use in a surface layer parameterization in a numerical model since it is defined in terms of the surface fluxes which are the quantities which need to be calculated. ς_1 can be written in terms of the bulk Richardson number, Ri_{B},of the surface layer:

$$\varsigma_1 = \frac{k\, c_{\mathrm{H}}}{c_{\mathrm{D}}^{3/2}} Ri_{\mathrm{B}} \tag{7.14}$$

The bulk Richardson number of the surface layer, Ri_{B}, can be defined in terms of the difference in buoyancy between model level 1 and the surface (or surface layer buoyancy difference), ΔB, and wind shear:

$$Ri_{\mathrm{B}} = \left(z_1 + z_{0\mathrm{m}}\right) \quad \Delta\mathrm{B} / \left|v_1 - v_0\right|^2 \tag{7.15}$$

Ri_{B} is a suitable measure of the surface layer stability for an atmospheric model since it is can be calculated readily from the basic model variables.

Similarly, by dimensional analysis, the gradients of temperature and moisture variables in the free convective limit can also be defined, with the involvement of a free convective roughness length, $z_{0\mathrm{f}}$. c_{D} and c_{H}, therefore, are specified directly as functions of Ri_{B}, z_1, $z_{0\mathrm{h}}$, $z_{0\mathrm{m}}$.and $z_{0\mathrm{f}}$. The dependence of c_{D} and c_{H} on stability as Ri_{B} becomes large and positive, therefore, has been chosen to be a decreasing function, which only tends to zero for infinite Ri_{B} . As $|v_1 - v_0|$ in unstable conditions, "free convection" takes place. In terms of the Richardson number the free convective limit corresponds to $Ri_{\mathrm{B}} \rightarrow -\infty$.

As seen above, to evaluate the surface exchange coefficients c_D and c_H requires that the roughness lengths for momentum, z_{0m}, heat and moisture, z_{0h}, and free convective turbulence, z_{of} are specified. The UM has assumed z_{0f} equal to z_{0h} for land and sea-ice, but given as a constant value for ocean points without sea-ice.

For land points the z_{0m} and z_{0h} are set to z_0 where

$$z_0 = \begin{cases} \max(\min(z_{0V},\ 5\text{x}10^{-4}),\ z_{0V} - 4\text{x}10^{-4}\,S) & \text{if } S < 5\text{x}10^{3} \\ z_{0V} & \textit{otherwise} \end{cases} \qquad (7.16)$$

In (7.16), S represents the mass of snow per unit area in kg m^{-2}, and z_{0V} is the **roughness length** representing the effects of vegetation and very small-scale surface irregularities (not orography). The vegetation roughness length, z_{0V}, is a climatologically prescribed, geographically varying quantity depending on the vegetation and land use. For the UM, Jones (1995) describes their derivation from the Wilson and Henderson-Sellers (1985) global archive of land cover and soils data.

For ocean points with sea-ice the z_{0m} and z_{0h} are set to $z_{0(SICE).}$ The UM uses $z_{0(SICE)} = 3\text{x}10^{-3}$ m, which corresponds to c_{DN} (10m) = 2.4x10^{-3}, which is a compromise from various c_{DN} (10m) quoted earlier by Overland (1985). Overland quotes values of the neutral drag coefficient at 10 metres for various types of sea-ice, which can be translated into roughness lengths using (7.12) with z_1 = 10 m.

For ocean points without sea-ice, the z_{0m} and z_{0h} are set to $z_{0m(SEA)}$ and $z_{0h(SEA).}$respectively. The UM according to Smith (1993), however, fixes $z_{0h(SEA)}$ at 10^{-4} m, though $z_{0m(SEA)}$ is varied following the formula proposed by Charnock (1955).

7.2.3.3: The treatment of the surface flux of moisture

The standard equation, (7.3), implies that the surface flux of moisture is given by

$$E_*/\rho_* \;=\; -c_H\left|v_1 - v_0\right|\left(q_{w1} - q_*\right) \tag{7.17}$$

The implied value of the surface specific humidity, q_*, is inextricably linked to the parameterization of surface hydrology and it is not easy to predict explicitly. q_* is assumed to be the saturated specific humidity corresponding to the surface temperature and pressure, $q_{SAT}(T_*,p_*)$, when the surface is ocean, sea-ice, or snow-covered land, or when $q_{w1} > q_{SAT}(T_*,p_*)$ for land. In all these cases, the factor $\bar{\ }$ mentioned in *(7.2.3.1)* is, therefore, equal to unity.

For land surfaces with a positive moisture flux, the turbulent flux into the atmosphere from the soil moisture store, E_s, is calculated in the UM using the "resistance method" by Monteith (1965), which gives

$$\frac{E_s}{\rho_*} = \frac{\left(q_* - q_{w1}\right)}{r_A} = \frac{\left(q_{SAT}\left(T_*,p_*\right) - q_*\right)}{r_S} = \frac{\left(q_{SAT}(T_*,p_*) - q_{w1}\right)}{r} \tag{7.18}$$

where r_A and r_S represent the aerodynamic resistance and **<u>surface resistance to evaporation</u>**, respectively.

By comparison of (7.18) and (7.17), it is clear that r_A is given by

$$r_A \;=\; \left(c_H\left|v_1 - v_0\right|\right)^{-1} \tag{7.19}$$

The aerodynamic resistance (r_A), therefore, represents the efficiency of the atmospheric turbulence in the evaporation process. The surface resistance to evaporation (r_S) is referred to as the stomatal resistance in the model, which characterises the physiological control of water loss through a plant community. The r_S, in effect, represents all the stomata of all the leaves acting in parallel so that the plant community acts like a "giant leaf". The use of the Monteith's resistance method requires the specific humidity in the sub-stomatal cavity to be saturated, $q_{SAT}(T_*,p_*)$. The source of moisture in the transpiration process is the sub-stomatal cavity of the leaf where the air is saturated (or nearly saturated), unless the plant is under severe water stress or is dessicated.

The effective value of r_S for a grid-box is a complicated function of the type and condition of the vegetation, the soil moisture content, the near surface air temperature and humidity and the amount of solar radiation reaching the surface. This means that to have a real representation of r_S, the model should include an interactive parameterization of its values in the run. This study, however, only used a version of the model which takes a climatologically prescribed, geographically varying quantity of r_S depending on the vegetation type. For the UM, Jones (1995) describes the derivation from the Wilson and Henderson-Sellers (1985) global archive of land cover and soils data.

With this prescription of r_S, therefore, the dependence of the surface moisture flux is not on the surface resistance parameter, rather it depends on the moisture content. A soil moisture availability factor, $\bar{s}$, hence, is introduced into the equation for the flux of moisture from the soil to the atmosphere:

$$E_S = \rho_* \Psi_S \left(q_{w1} - q_{SAT}(T_*, p_*) \right) / r_A + r_S \tag{7.20}$$

Equation (7.20) parameterises the flux of moisture which comes from the subsurface water, i.e. the soil moisture store. The soil moisture availability factor, $\bar{s}$, depends on the dimensionless volumetric soil moisture concentration, χ, such that

$$\Psi_S = \begin{cases} 0 & 0 \leq \chi < \chi_w \\ (\chi - \chi_w)/(\chi_C - \chi_w) \quad \text{for} & \chi_w < \chi < \chi_C \\ 1 & \chi_C \leq \chi \end{cases} \tag{7.21}$$

where χ_w is the residual value of χ at the wilting point (i.e., the value of χ below which it becomes impossible for vegetation to remove moisture from the soil), and χ_c is a critical value of χ below which the flux of soil moisture to the surface or the plant roots is restrained (i.e., below which $\bar{s} < 1$).

In the UM, χ is defined in terms of the soil moisture available for evapotranspiration, m, as follows:

$$\chi = \begin{cases} m/(\rho_w D_R) + \chi_w & \text{if } D_R > 0 \\ \chi_w & \text{otherwise} \end{cases} \tag{7.22}$$

where D_R is **root depth** of vegetation and ρ_w is the density of liquid water. χ_W and χ_C, as used in the UM, are climatologically prescribed, geographically varying parameters depending on soil type and D_R is a similar parameter depending on vegetation type. For the UM, Jones (1995) again describes the derivation of these parameters from the Wilson and Henderson-Sellers (1985) global archive of land cover and soils data.

In addition to the subsurface water store given in (7.20) above, the model also represents the effect of a surface water store. The surface water store includes a vegetative canopy store and water lying on the soil surface directly exposed to the atmosphere. This surface water store is commonly called "canopy". Note that the canopy is a generalised term for the surface water store, not limited to a vegetated surface alone.

The water in the surface store evaporates with only aerodynamic resistance since this water does not go through the soil, root and leaf stomata system. The grid-box mean "potential evaporation", E_p, is given by

$$E_p/\rho_* = -c_H |v_1 - v_0| \left(q_{w1} - q_{SAT}(T_*, p_*)\right) = -\left(q_{w1} - q_{SAT}(T_*, p_*)\right)/r_A \tag{7.23}$$

and the grid-box mean "canopy evaporation" is defined to be

$$\hat{E}_A = \beta\ E_p \tag{7.24}$$

where

$$\beta = \begin{cases} c/c_M & \text{if } c_M > 0 \\ 0 & \text{otherwise} \end{cases} \tag{7.25}$$

Here c represents the canopy water content, c_M is the **canopy water capacity**. The canopy capacity is a climatologically prescribed, geographically varying parameter depending on the vegetation fraction and type. For the UM, Jones (1995) describes its derivation from the Wilson and Henderson-Sellers (1985) global archive of land cover and soils data.

In the Penman-Monteith energy balance scheme used in this study, however, the ϑ-function is treated according to the formulation described in Appendix A [c.f. Equation (A.8) and (A.9)] rather than directly from Equation (7.25). In this scheme, the evaporation rate from a vegetated surface is defined in terms of a surface (canopy) resistance to evaporation, r_s, that is approximately equal to the resistance imposed by all the leaf stomata acting in parallel.

The grid-box mean flux of water from the soil is given by

$$\hat{E}_s = (1 - \beta) E_s \tag{7.26}$$

The total flux of moisture from a land grid-box is then

$$E_* = \hat{E}_A + \hat{E}_s \tag{7.27}$$

Another related process regarding how rainfall is intercepted by the surface water store will be discussed separately in Section 7.3.

7.2.4: The surface and boundary layer equations and their implicit solution

The surface and boundary layer equations and their implicit solution component calculates the increments to the surface temperature and boundary layer temperature, moisture and wind using an implicit numerical scheme. The surface and boundary layer scheme works at the lowest boundary layer levels of the atmospheric layers. For land points of the model, the surface and boundary layer scheme updates the deep soil temperatures, and for both land points and ocean points with sea-ice, the scheme updates grid-box mean surface temperature. The updating of the liquid/frozen water temperatures, total water contents and horizontal wind components, however, takes place for all model points. Surface moisture fluxes are output for use in the surface and subsurface hydrology component (see Section 7.3), which updates the amount of lying snow, the surface (or "canopy") water store and the soil moisture.

In the following paragraphs a brief summary of the prognostic equations for surface temperature, turbulent mixing of heat and moisture will be given. The implicit solution for those equations, too, will be discussed only briefly; their complete finite difference forms can be referred to in Smith (1993).

7.2.4.1: The surface temperature equation

The surface temperature is either predicted or prescribed, depending on the type of model points and configurations, either coupled or atmosphere only. In either configuration, the sea surface temperature is input but not updated by the surface and boundary layer scheme. Overall, for ocean points with no sea-ice, the surface temperature is predicted by an ocean model in coupled configurations, but it is prescribed from a climatology or from an analysis of observations in the atmosphere-only configuration.

For land points, the rate of change of surface temperature is given by

$$\frac{dT_*}{dt} = A_{S1}\left(R_{N\downarrow *} - H_S - H_* - L\,E_*\right) \tag{7.28}$$

A_{S1} (= $1/C_s\Delta z_1$)is the reciprocal area heat capacity of the top soil layer when C_s is the volumetric specific heat capacity of the soil and Δz_1 is the top soil layer thickness. $R_{N\downarrow *}$ is the net downward radiative flux at the surface calculated in the radiation scheme. H_S is the heat flux from the top to next-to-top soil layer calculated in the soil thermodynamics [as discussed in (7.2.2)]. H_* and E_* are the sensible heat and moisture fluxes at the surface respectively (both positive upwards). L is the latent heat and is set appropriately either with or without lying snow.

The prognostic equations of (7.28), are solved implicitly by the model using a finite difference method and their formulation as employed in the UM is given in detail in Smith (1995).

7.2.4.2: The equations for turbulent mixing of momentum, heat and moisture

For local turbulent mixing, the prognostic equations for the vector horizontal wind, v, the liquid/frozen water temperature, T_L, and the total water content, q_w, are given below:

$$\frac{\partial v}{\partial t} = -g\frac{\partial \boldsymbol{\tau}}{\partial p} + \left.\frac{\partial v}{\partial t}\right|_{nt} \tag{7.29}$$

$$\frac{\partial T_L}{\partial t} = g\frac{\partial F_{TL}}{\partial p} + \left.\frac{\partial T_L}{\partial t}\right|_{nt} \tag{7.30}$$

$$\frac{\partial q_w}{\partial t} = g\frac{\partial F_{qw}}{\partial p} + \left.\frac{\partial q_w}{\partial t}\right|_{nt} \tag{7.31}$$

where $(\partial v/\partial t)|_{nt}$, $(\partial T_L/\partial t)|_{nt}$ and $(\partial q_w/\partial t)|_{nt}$ are the rate of change of v, T_L and q_w from all the other parts of the model, i.e., the non-turbulent (nt) contributions. The term $\boldsymbol{\tau}$ in (7.29) represents stress in the boundary layer and $(\partial\boldsymbol{\tau}/\partial p)$, $(\partial F_{TL}/\partial p)$ and $(\partial F_{qw}/\partial p)$ are the vertical gradient of $\boldsymbol{\tau}$, fluxes for T_L and q_w.respectively.

The prognostic equations of (7.29), (7.30) and (7.31) are solved implicitly using finite differencing method. The formulation used in the UM is given in detail in Smith *et al.* (1995). The equation for turbulent mixing of momentum in the boundary layer [i.e. Equation (7.29)] is treated only for local mixing. Although the turbulent mixing of heat and moisture are also treated as non-local mixing in unstable conditions, (7.30) and (7.31) are still applicable. In the case of non-local mixing, however, the total turbulent fluxes from layer (k-1) to layer k for $2 \le k \le N_{rm1}$ in the finite difference scheme are treated differently compared to the local mixing alone since both the non-local, "rapid mixing" (rm) and local mixing (lm) are involved. Note that, if $N_{rm1} < 2$, then all mixing is local. Smith (1995) gives detail regarding the finite difference formulations for all the range of k which are calculated by the UM.

7.2.5: Adjustment of the surface evaporation and sublimation and the surface temperature increment

The component covering the adjustment of the surface evaporation and sublimation and the surface temperature increment does an adjustment when one or more of the land surface water stores (lying snow amount, canopy water and soil water) become exhausted during a timestep. The surface moisture fluxes and hence the boundary layer temperature and moisture increments may need some adjustment if the land surface hydrological stores are too low to sustain the fluxes over a timestep. It also adjusts the sea-ice surface temperature back to the freezing point if melting is taking place.

At land points, this component detects the above occurrence and the moisture flux is adjusted so that no more than the available snow or water enters the atmosphere as vapour. To correspond, the latent heat flux, the surface temperature and the temperature and moisture in the boundary layer need to be adjusted.

An adjustment is also done at ocean points with sea-ice fraction greater than zero. If the ice surface temperature is greater than its melting point, then it will be adjusted to reduce to its melting point, T_M. Associated with this implied melting, the latent heat flux is also calculated.

7.2.6: Boundary layer cloud

The turbulent transport calculated by the boundary layer scheme takes into account the latent heating effects of boundary layer cloud; it does this by using "cloud conserved" variables. This sub-component calculates the temperature, humidity, cloud water contents and cloud amounts from the updated conserved variables.

The updated values of the total water content, q, and the liquid/frozen water temperature, T_L, for layers 1 to boundary layer levels are calculated in the surface and boundary layer equations discussed in (*7.2.4*). Immediately after the surface and boundary layer calculations, the updated temperature (T), specific humidity (q), cloud liquid water ($q_c^{(L)}$), cloud ice water ($q_c^{(F)}$), and cloud fraction (C) are calculated by calling a subroutine of another component that calculates the saturated specific humidity and large-scale cloud.

7.3: Canopy and Land Surface Hydrology Processes

This section presents a summary of the parameterization of land surface hydrology processes used in the UM (excluding snow-related processes which are not directly relevant to this study). The summary is mainly based on the report by Gregory *et al.* (1994).

7.3.1: Canopy and surface hydrology

The effect of vegetation on the soil moisture budget of the land surface cannot be ignored in the model. Falling water, before reaching the soil, is intercepted by the canopy resulting in drier soils and, therefore, affecting the soil hydrology cycle. The complex interaction of falling water with the surface canopy can only be modelled with simple schemes in GCMs. Warrilow *et al.* (1986) describes various schemes that have been employed in a variety of models in their discussions on the interacting mechanisms of the falling water and the canopy.

The UM has an additional canopy mechanism which can retain water falling through it and so reduce the water supply to the soil moisture store. Water can also evaporate from the canopy into the atmosphere. As discussed in Section (7.2) water stored in the canopy evaporates with only aerodynamic resistance, so more easily than evapotranspiration from the underlying soil. We should not be misled into assuming the "canopy" refers only to a vegetated surface in the model. The "canopy" is generally referred to as surface water store, with or without the presence of vegetation. The properties of the canopy (or surface) water store, therefore, are spatially varying depending upon the vegetation type and fractional cover within a grid-box.

7.3.1.1: Canopy interception and throughfall

Gregory *et al.* (1994) describe the canopy interception and throughfall scheme of the UM according to derivation by Warrilow *et al.* (1986), with modifications as suggested by Shuttleworth (1988). The derivation of Warrilow *et al.*'s has been based on the earlier work of Rutter *et al.* (1971).

Two major differences between the scheme of Warrilow *et al.* and that of Shuttleworth are:-

(i) *In terms of the assumption made for the throughfall:* In the treatment of the interception of falling water within the canopy, Warrilow *et al.*'s scheme allows the throughfall to occur even before the canopy is full. Shuttleworth's scheme, however, assumes no throughfall occurs unless the water fall rate is greater than that required to fill the local canopy within a model timestep. This means if the water fall rate is less than model timestep, then all water is intercepted by the canopy.

(ii) *In terms of the assumption made on water falling onto the top of the canopy:* Warrilow *et al.*'s scheme assumes that water falling onto the top of the canopy did so evenly over the grid-box. Shuttleworth's scheme, however, assumes that water falling on the canopy does so only over a limited area of the grid-box and within that area the local water rates are exponentially distributed.

For (i), the UM follows the formulation of Warrilow *et al.* since their assumption is more realistic than Shuttleworth's assumption. According to the argument presented by Gregory *et al.*, the capture of water by plants and the dripping of water through it (which throughfall represents) both occur in the vertical over a similar length scale. Similar time-scales, therefore, are expected for both processes. As the result, throughfall would occur even though the water (rain) fall rate is smaller than that required to fill the local canopy within a model timestep. Furthermore, it is physically true that some parts of the canopy water may fall through directly.

For (ii), on the other hand, the assumption made by Shuttleworth is chosen by the UM. This is to make it consistent with the treatment of surface runoff in the model. The local rate of throughfall of water from the canopy to the surface is assumed to be exponentially distributed and to be occurring only over a fractional area, *M*, of the grid-box. This fractional area (*M*) is referred to as **vegetation fraction** in the vegetation ancillary file of the UM, which is a climatologically prescribed, geographically varying parameter following the Wilson and Henderson-Sellers (1985) global archive of land cover and soils data.

Falling water is either due to condensation onto the canopy or large-scale rain. Following the assumption by Shuttleworth, water is assumed to fall onto the top of the canopy over a fractional area ε with local rate, R_L (in kg $m^{-2}s^{-1}$). If R is the grid-box average of the fall rate (in kg $m^{-2}s^{-1}$), hence, the rate is assumed to be exponentially distributed with the function of R_L as follows:

$$f\left(R_L\right) = \frac{\varepsilon}{R}\exp\left[-\frac{\varepsilon\ R_L}{R}\right] \tag{7.32}$$

In the interception and throughfall mechanism, as water falls through the canopy a portion is captured and the remainder falls to the surface. The local throughfall rate of water from the base of the canopy to the surface is given as:

$$T_{FL} = R_L\frac{c}{c_M} \tag{7.33}$$

where c is the canopy water content (in kg m^{-2}, often given as equivalent "m" or "mm"), and c_M is the **canopy water capacity** (in kg m^{-2}). This c_M is referred to the maximum amount of water the canopy can hold. The canopy water content is assumed to be distributed evenly over the entire grid-box.

From (7.33), two cases are possible:

(i) if the local canopy is not filled by the amount of water intercepted, that is when

$$\hat{c}_L = c + \delta t \left(R_L - T_{FL} \right) \quad \leq \quad c_M \tag{7.34a}$$

where $|_L$ is the local canopy water content after interception (in kg m^{-2}), and Λt is the model timestep (in seconds). Replacing T_{FL} in (7.34a) using (7.33) yields

$$R_L \leq \quad \frac{c_M}{\delta t} \tag{7.34b}$$

(ii) if more water is intercepted during the timestep than it is possible for the local canopy to hold, i.e. if

$$\hat{c}_L = c + \delta t \left(R_L - T_{FL} \right) \quad > \quad c_M \tag{7.35a}$$

Similarly, replacing T_{FL} in (7.35a) yields

$$R_L > \quad \frac{c_M}{\delta t} \tag{7.35b}$$

For case (i), i.e. when $R_L\ \alpha\ C_M/\Lambda t$, the local throughfall rate of water from the base of the canopy to the surface T_{FL} follows (7.33). For case (ii), i.e. $R_L\ \beta\ C_M/\Lambda t$, the excess water is added to the local throughfall giving (after derivations):

$$T_{FL} \quad = \quad R_L - \left[\left(c_M - c \right) / \delta t \right] \tag{7.36}$$

The final formulation of the local throughfall rate (in kg m^{-2}) can be obtained where the throughfall averaged over the fractional area of the grid-box where rain occurs is given by

$$T_{FL} \quad = \quad \int_0^\infty T_{FL}\, f\left(R_L \right) d\, R_L \tag{7.37}$$

On expanding (7.37), evaluating by parts using (7.33) and multiplying by the fractional area M over which water falls gives the throughfall averaged over the grid-box

$$T_F^A \quad = \quad R\left[1 - \frac{c}{c_M}\right]\exp\left[-\varepsilon\frac{c_M}{R\delta t}\right] \quad + \quad R\frac{c}{c_M} \tag{7.38}$$

The rate of change of the canopy water content (in kg m^{-2}) as water falls through the canopy is given by

$$(\partial C/\partial t) \quad = \quad (R - T_F^A) \tag{7.39a}$$

And the final updated canopy water content, | in kg m^{-2} is

$$\hat{c} \quad = \quad c + (\partial C/\partial t)\,\delta t \tag{7.39b}$$

<u>*7.3.1.2: Surface hydrology*</u>

The throughfall water from the canopy on reaching the surface infiltrates the soil at a rate, K_{SV} (in kg m^{-2} s^{-1}), equal to the saturated hydrological soil conductivity, K_S, modified due to the presence of vegetation. If the throughfall rate exceeds the infiltration rate, there is surplus water on the surface and this runs off into rivers, lakes etc. The local surface runoff, Y_{SL} (in kg m^{-2}s^{-1}), therefore, is given by

$$Y_{SL} \quad = \quad \begin{cases} T_{FL} - K_{SV} & \text{if } T_{FL} > K_{SV} \\ 0 & \text{if } T_{FL} \leq K_{SV} \end{cases} \tag{7.40}$$

Similar to the treatment of local throughfall rate in (7.37), the surface runoff averaged over the area in the grid-box over which water falls is

$$Y_{SL} \quad = \quad \int_0^{\infty} Y_{SL}\, f(R_L)\, dR_L \tag{7.41}$$

The equations for the grid-box average surface runoff, Y_S^A, can be derived starting from the integral (7.41). The integral is firstly separated into two integrals (one with its limit $0 \rightarrow c_M/\delta t$, and another with limit $c_M/\delta t \rightarrow \infty$) added together. Hence, by considering water fall over a fractional area M, the grid-box average surface runoff is given by

$$Y_S^A = \begin{cases} R\dfrac{c}{c_M}\exp\left[\dfrac{-\varepsilon K_{SV}\, c_M}{Rc}\right] + R\left[1 - \dfrac{c}{c_M}\right]\exp\left[-\varepsilon\dfrac{c_M}{R\delta t}\right] & \text{if } K_{SV}\,\delta t \le c \\ R\exp\left[\dfrac{-\varepsilon\left(K_{SV} + P_M\right)}{R}\right] & \text{if } K_{SV}\,\delta t > c \end{cases} \tag{7.42a}$$

where P_M is equal to $(c_M - c)/\Lambda t$, a measure of canopy water content in relative to canopy capacity at each timestep.

Snowmelt is assumed to cover the whole grid-box, (i.e. M is set to 1.0). It does not interact with the canopy water store. Recall that the snow melting rate, S_M, is defined in (7.30). The surface runoff of any snowmelt, basically, is calculated from (7.42a) but with c_M replacing c and S_M replacing R. A simplified version of (7.42a) for snowmelt is given by

$$Y_S^{A(SM)} = S_M \exp(-\varepsilon K_{SV}/S_M) \tag{7.42b}$$

The rate at which the soil moisture content is increased by throughfall, therefore, is defined as

$$\frac{\partial m}{\partial t} = \left(T_F^A - Y_S^A\right) \tag{7.43}$$

The soil infiltration rate (K_{SV}) is pre-calculated and passed into the model as an ancillary field. It is obtained from the saturated hydrological soil conductivity, K_S (in kg m^{-2} s^{-1}) and the soil **infiltration enhancement factor**, ϑ_V (dimensionless), which accounts for the effects of root systems on the infiltration of surface water into the soil. K_{SV} is given by

$$K_{SV} = \beta_V K_S \tag{7.44}$$

The soil infiltration enhancement factor, ϑ_V, is also a climatologically prescribed, geographically varying parameter depending on the vegetation type and K_S is a similar parameter depending on soil type. The c_M is non-zero everywhere over land except where land-ice exists where it is zero. Bare soil is given a small canopy capacity (of 0.001m) to account for surface retention (such as in puddles etc).

Prior to any calculations of canopy and surface hydrology, water is removed from the canopy water store by evaporation. The updated water content, | , is given by

$$\hat{c} = c - \delta t E_c \tag{7.45}$$

where E_c is the evaporation from the canopy calculated by the component discussed in Section (7.2), boundary layer processes. Note that (7.45) applies only when E_c χ 0.

7.3.13: Implementation in the UM

There are three sources of the water which falls through the canopy before reaching the surface:-

i. Canopy condensation: defined as negative of evaporation (E_c) from the canopy. This is assumed to occur over all the grid-box by setting M equal to 1.0. Canopy condensation is calculated in boundary layer component as discussed in Section 7.2.

ii. Large-scale rain: defined as $R^{(LS)}$ and calculated in the large-scale precipitation scheme (Smith *et al.*, 1995). This is assumed to occur over all of the grid-box (i.e. M is set to 1.0)

iii. Convective rain: defined as $R^{(c)}$ and calculated in the convective scheme (Gregory and Inness, 1996). This is assumed to fall over 30% of the grid-box (i.e. M is set to 0.3)

In the UM, (7.39a,b) are applied to each of these water types in turn, starting with canopy condensation, then large-scale rain and finally convective rain. The canopy water content is updated between the calculations for each type. Equation 7.43 is used to calculate the rate of change of soil moisture content due to throughfall reaching the surface.

The calculation for the canopy interception is undertaken for all land types except land ice where canopy capacity, c, and infiltration rate, K_{SV}, are both defined to be zero. For land ice, any rainfall at the surface is runoff.

The state of precipitation at the surface is determined by the temperature of the lowest model layer, which may be above freezing when snow is lying on the surface. It is, therefore, possible for water to fall onto the canopy at a grid point which is covered with snow, and the canopy interception calculation is still carried out before the surface runoff calculation. This is reasonable as some vegetation may penetrate above the snow layer or bare patches with no snow may occur in the grid-box.

Contributions to throughfall and surface runoff (including that from snowmelt) are added and passed out as diagnostics. The rate of change of soil moisture content [in (7.43)] is also summed for each water type and passed across to the sub-surface hydrology component to update the soil moisture content.

7.4: Soil Hydrology Scheme

Both single layer and multilayer soil hydrology schemes are available in the UM. For the purpose of this study, the multilayer scheme was employed due to its advantages over the single layer scheme. Before describing the multilayer scheme, it is useful to look at the basic concept of the scheme in the single layer formulation.

7.4.1: Single layer scheme

The soil hydrology scheme of the UM updates soil moisture content, m, by taking account of:

(i) evaporation which removes moisture from the soil, i.e. when E_S χ 0;

(ii) throughfall of water from the canopy and snowmelt less surface runoff ($T_F + S_M - Y_S = F_W$, which is to be positive downwards), calculated in the canopy and surface hydrology scheme; and

(iii) subsurface runoff due to gravitational drainage, Y_G.

The soil moisture content (m) is first updated as follows:

$$m = m + \delta t\left(F_W - E_S\right) \tag{7.46}$$

Subsurface runoff, $Y_{G,}$ depends only on the soil moisture content and the process is a relatively slow compared to the surface runoff, Y_S, which depends on the canopy throughfall or snowmelt rate. For a single layer scheme, the parameterization for mass flux of Y_G follows the empirical formula by Eagleson (1978).

$$Y_G = \begin{cases} 0 & \chi < \chi_w \\ K_S\left(\left(\chi - \chi_w\right)/\left(\chi_s - \chi_w\right)\right)^C & \chi_w \le \chi \le \chi_s \\ K_S & \chi_s \le \chi \end{cases} \tag{7.47}$$

where _ and $\_{w}$ are volumetric soil moisture concentration [as defined earlier in (7.22)] and its residual value at the wilting point respectively; $\_{s}$ is the saturation value of _ attained when all the empty spaces between soil particles are filled with water; and K_S is the saturated hydrological conductivity of the soil as already mentioned during the discussion on surface hydrology.

Gregory *et al.* (1994) show details of the computation of Y_G. When it is computed, the soil moisture content in single layer scheme is updated according to:

$$m \quad = \quad m - \delta t\, Y_G \qquad (7.48)$$

7.4.2: *Multilayer scheme*

A multilayer soil hydrology scheme in the UM uses the same vertical discretisation as the soil thermodynamics of the model. Differing from single layer hydrology, the multilayer scheme needs to update the moisture content as its prognostic variable at each soil layer. The gravitational drainage is redefined as the drainage from the base of the total soil profile in multilayer scheme, rather than the drainage from the bottom of the root-zone as in the single layer scheme. The extension below the root zone gives better simulation of the changes in drainage. The gradient in the soil water tension at the base of the root zone is more realistic, rather than assuming it to be zero (the boundary condition for a single layer model). The multilayer model, therefore, is capable of simulating the partial recharge of the root zone during dry periods by water from below the zone. This is expected to give an extra advantage to improved simulation of the diurnal and seasonal variation in the surface evaporation and runoff flux. Single layer models produce too little drainage from the root-zone in wet periods and too much drainage in dry periods, tending to overestimate the variation in the surface evaporation and runoff flux (Gregory *et al.*, 1994).

With z representing the vertical coordinate (positive downwards), the continuity equation for soil moisture concentration, *Π*, is as follows:

$$\frac{d\theta}{dt} \quad = \quad -\frac{\partial W}{\partial z} - R \qquad (7.49)$$

where R is a sink term which represents the extraction of water by plant roots, and *W* is a water flux given by Darcy's Law:

$$W \quad = \quad K\left\{\frac{\partial \psi}{\partial z} + 1\right\} \qquad (7.50)$$

In Darcy's Law, *K* represents the hydraulic conductivity and ψ is the soil water tension which is chosen to be positive. To close the model it is necessary to assume that both *K* and ψ are a function of the soil moisture concentration. According to Gregory *et al.* (1994), there have been many forms of curves to satisfy the closure (e.g. Clapp and Hornberger, 1978; Eagleson, 1978). The closure used in the UM's scheme is the one derived by van Genuchten *et al.* (1991):

$$\psi = \psi_1 S^{-b} \left\{1 - S^{b+1}\right\}^{\frac{b}{b+1}} \tag{7.51}$$

$$K = K_s S^L \left[1 - \left\{1 - S^{b+1}\right\}^{\frac{1}{b+1}}\right]^2 \tag{7.52}$$

where K_s, ψ_1, b and L are all empirical soil dependent constants, and ***S*** is soil moisture dependent variable defined by:

$$S = \frac{\theta - \theta_r}{\theta_s - \theta_r} \tag{7.53}$$

where Π_s is the saturation soil moisture concentration and Π_r is the "residual" soil moisture concentration, below which drainage ceases.

The sink term, R, in the continuity equation (7.49) is related to the evaporation from the soil, E, by

$$E = \int_{z_d}^{0} R \, dz \tag{7.54}$$

where z_d is the root depth. The evaporation (E) is treated as an input to the soil hydrology module in the UM as it is calculated in the boundary layer scheme. Then R is determined by distributing the total evaporative demand vertically within the root zone and it is assumed that the local contribution to the flux is proportional to the root density, C_r, and the soil moisture concentration:

$$R = \frac{C_r\left\{\theta - \theta_w\right\}}{\int_{z_d}^{0} C_r\left\{\theta - \theta_w\right\}dz} E \tag{7.55}$$

where Π_w is the wilting soil moisture concentration below which evaporation ceases. The root density is assumed to increase quadratically from zero at the root depth, z_d, to a finite value at the surface following:

$$C_r = \mu\left(z_d - z\right)^2 \tag{7.56}$$

where T is a constant.

The multilayer model as mathematically described by equations (7.49) to (7.56) above is in finite difference form. Gregory *et al.* describe in detail the discretisation in their finite difference form, the numerical method used and the organisation in the UM's code.

7.5: Surface Albedo Specification

At the surface, a net radiative flux is produced by the radiation scheme for use by the surface calculations described in (7.28) earlier in sub-section (*7.2.4*). Ingram et. al. (1996) describe in detail the scientific aspects of the representation of radiative processes in the UM's calculations which are contained in two schemes: (i) the longwave scheme based on that described by Slingo and Wilderspin (1986), and (ii) the shortwave scheme based on Slingo (1985), including the extended scheme to include Slingo's (1989) spectral division and interactive cloud optical properties. Again, sea-ice considerations are omitted.

7.5.2: Albedo of land points

The albedo of snow-free land is specified as a function of vegetation, land use and soil type modified where necessary to remove unreasonable values. This **snow-free albedo** as used in this study, therefore, is a climatologically prescribed, geographically varying parameter depending on the vegetation. For the UM, Jones (1995) describes the derivation from the Wilson and Henderson-Sellers (1985) global archive of land cover and soils data.

Snow lying on the ground strongly modifies the surface albedo as well as having the non-radiative effects described in the earlier section in this chapter. This effect depends on vegetation, and on the depth, density and age of the snow. As reported by Robinson and Kukla (1984, 1985), albedo generally increases with snow depth, though bare patches tend to form during melting, giving smaller albedos than for the same average depth during accumulation, when the cover tends to be more even. The UM, however, uses only a simple form of relationship suggested by Hansen *et al.* (1983), which relates the snow amount S, snow-free albedo, I_0, and deep-snow albedo, $I_{D,}$ as follows:

$$\alpha_T = \alpha_0 + (\alpha_D - \alpha_0)(1 - e^{-as}) \tag{7.57}$$

where a is set to 0.2 m^2kg^{-1}, consistent with tuning experiments by the UKMO and the observations of Robinson and Kukla (1985).

Jones (1995) documented the derivation of maximum deep-snow albedos for each vegetation type, $I_{D.}$ The UM uses only a simple linear temperature dependence to determine I_D as follows:

$$\alpha_D = \begin{cases} \alpha_s & T_* \leq T_M - \Delta T \\ \alpha_s + 0.3\,(\alpha_0 + \alpha_s)(T_* - T_M + \Delta T)/\Delta T & \text{for } T_M - \Delta T \leq T_* \leq T_M \end{cases} \tag{7.58}$$

For deforestation experiments in this study, which involve the tropical region of Southeast Asia, only snow-free albedo is changed within the deforestation area. As snow occurrence is not expected in the area of study, deep-snow albedo change is not involved.

7.6: Summary

In describing the land surface scheme of the Unified Model in this chapter, emphasis has been placed on those aspects that are important for the change of vegetation parameters that are involved in deforestation experiments undertaken in this study. Those parameters are surface roughness lengths, surface "stomatal" resistance for evaporation, root depth, surface "canopy" capacity, vegetation fraction, infiltration "enhancement" factor and snow-free albedo. Understanding of their formulation as given in this chapter is useful since those parameters are modified when undertaking perturbed experiments in this study. Following the introduction of those parameters, extended information regarding the set-up of the above parameters which are involved in the model experiment is given in the next chapter.