UNIFIED MODEL DOCUMENTATION PAPER 25

CANOPY, SURFACE AND SOIL HYDROLOGY

by

D Gregory, R N B Smith (sections 1-4), and
P M Cox (section 5)

Version 3

November 1994

Model Version 3.4 **

**Librarian's Note: Users of versions 3.2 and 3.3 of
the Unified Model can use this document
if they ignore section 5.

Climate Research
Meteorological Office
London Road
BRACKNELL
Berkshire
RG12 2SY
United Kingdom

Modification Record (from version 3)		
Document version	**Author**	**Description....................**
3	P M Cox	Section 5 added, covering multilayer soil hydrology scheme.

CONTENTS

(1) INTRODUCTION

This paper describes the parametrization of land surface hydrology processes used in the UK Meteorological Office Unified Model. The following processes are described:-

Snow processes :

The accumulation of snowfall, calculation of snowmelt and the snowmelt heat flux.

Canopy and surface hydrology :

The decrease of canopy water content by evaporation, surface the increase in canopy water content through the interception of falling rain and hydrology condensation occurring within the canopy. The surface runoff of throughfall from the canopy is calculated together with the infiltration of water into the soil.

Soil hydrology:

The effect of evaporation and infiltration of surface water into the soil on soil water content. The sub-surface drainage of water is also calculated.

(2) SNOW PROCESSES AT THE LAND SURFACE (P251)

There are three processes represented in the model which change the amount of snow lying on the land surface:

(i) sublimation/deposition due to turbulent transport in the near-surface atmosphere;

(ii) accumulation of snowfall;

(iii) melting of snow when the surface temperature is above the melting point.

Other processes such as snow drifting and sliding down hill or mountain sides are not represented, nor is inhomogeneous melting due to the differing orientations of sloping surfaces.

Surface albedo, surface roughness and the thermal conductivity of the subsurface layer are all functions of the amount of surface snow; the dependencies are discussed in the parts of the model documentation dealing with the processes involving these quantities.

The three processes listed above are also represented for sea-ice surfaces in coupled atmosphere/ocean/sea-ice versions of the model. They are dealt with and documented elsewhere.

(i) The sublimation/deposition of snow

The sublimation/deposition rate is calculated as a positive/negative turbulent mass flux (positive upwards), E_I , in component P24 and passed to this component. The increment to the snow amount (held in the model as an areal density in kg m^{-2}) is minus the flux multiplied by the timestep. Snow amount,S, is then updated with this increment:

$$S = S - \delta t E_I \qquad \text{(P251.1)}$$

(ii) The accumulation of snowfall

Large-scale snowfall, $S_F^{(LS)}$, and convective snowfall, $S_F^{(C)}$, calculated by components P26 and P27 respectively are passed as mass fluxes (positive downwards) to this component. The increment to the snowdepth is found by multiplying the total snowfall rate (expressed by a mass flux) by the model timestep. The snow amount is then updated with this increment:

$$S = S + \delta t(S_F^{(LS)} + S_F^{(C)}) \qquad \text{(P251.2)}$$

(iii) The melting of snow

For land surfaces with the snow depth greater than zero and surface temperature (T_*) greater than the melting point of water (T_M) the snow melting rate (in kg m^{-2} s^{-1}) is given by

$$S_M = \min(S , (T_* - T_M)/L_F A_{S1})/\delta t \qquad \text{(P251.3)}$$

A_{S1} is the reciprocal areal heat capacity of the top soil layer (in J^{-1} m^2 K), as defined in component P242. In (P251.3) the snowmelt is constrained by both the amount of snow, S, (after updating as in (i) and (ii) above) and the amount of heat in the top soil layer available to melt the snow, i.e. $S_M \delta t \leq S$ and $L_F S_M \delta t \leq (T_* - T_M)/A_{s1}$.
The snow depth and surface temperature are then updated using

$$S = S - \delta t S_M \qquad \text{(P251.4)}$$

and

$$T_* = T_* - \delta t A_{S1} H_{SM} \qquad \text{(P251.5)}$$

where the latent heat flux required to melt the snow (inW m^{-2}) is

$H_{SM} = L_F S_M$ (P251.6)

S_M is output for use in P252 where the surface runoff of snowmelt is calculated. H_{SM} and S_M are diagnostic quantities which may be output. Because of the limitations on the snow melting rate imposed by (P251.3) the updated snow depth is ≤ 0 and the updated surface temperature is $\leq T_M$.

(3) CANOPY AND SURFACE HYDROLOGY (P252)

(i) Introduction

Vegetation can have a considerable effect on the soil moisture budget of the land surface. Falling water is intercepted by the canopy before reaching the soil, resulting in drier soils and a reduced soil hydrology cycle (Reifsnyder (1982)). Water stored in the canopy evaporates with only aerodynamic resistance and so more easily than evapotranspiration from the underlying soil (see UMDP No.24, sub-component P245 - surface evaporation and sublimation).

The interaction of falling water with the surface canopy is complex and it is only possible to use simple schemes in large-scale models. Warrilow et al. (1986) (hereafter referred to as WSS(86)) discusses the mechanisms by which falling water and the canopy interact and briefly describes schemes employed in a variety of models. The unified model has an additional surface water store 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. The properties of the canopy water store are spatially varying, depending upon the vegetation type and fractional cover with a grid box (see UMDP No.70)

The scheme described here is based on the formulation of WSS(86) but includes modifications along the lines suggested by Shuttleworth (1988) (hereafter referred to as S(88)).

(ii) Theory

WSS(86) assumed that water falling onto the top of the canopy did so evenly over the grid box. However this is inconsistent with their treatment of surface runoff in which the local rate of throughfall of water from the canopy to the surface was assumed to be exponentially distributed and to be occurring only over a fractional area e of the grid box.
S(88) describes an extension to WSS(86) in which the canopy interception is also calculated assuming that water falling on the canopy does so only over a limited area of the grid box and within that area the local water fall rates are exponentially distributed. Although S(88) uses the same surface hydrology scheme as WSS(86) his treatment of canopy interception is somewhat different. We think the formulation of canopy interception in WSS(86) is the more realistic of the two (see the Appendix for a brief comparison of the schemes) and the scheme outlined below uses the canopy interception formulation of WSS(86) modified according to S(88).

(a) Canopy interception

Water is assumed to fall onto the top of the canopy over a fractional area ε. Within that area the local water fall rate (either due to condensation onto the canopy, large-scale rain - see section (iii)), R_L in kg m^{-2} s^{-1}, is assumed to be exponentially distributed;

$$f(R_L) = \frac{\epsilon}{R} \exp\left[-\frac{\epsilon R_L}{R}\right] \qquad \text{(P252.1)}$$

where R is the grid box average water fall rate (in kg m^{-2} s^{-1}).

As water falls through the canopy a portion is captured, the remainder falling to the surface. The local throughfall rate of water from the base of the canopy to the surface is assumed to be given by

$$T_{FL}=R_L\frac{c}{c_M} \qquad (P252.2)$$

where c is the canopy water content (in kg m^{-2}), and c_M is the canopy water capacity (in kg m^{-2}), ie. the maximum amount of water the canopy can hold.

Note that although local rainfall rates are considered, the canopy water content is assumed to be distributed evenly over the entire grid box, i.e. the canopy capacity has no history of where water fell in a previous time step.

Equation (P252.2) applies if the local canopy is not filled by the amount of water intercepted, that is when,

$$\hat{c}_L=c+\delta t\,(R_L-T_{FL})\le c_M$$

$$\text{i.e.}\quad c+\left[R_L-R_L\frac{c}{c_M}\right]\delta t\le c_M \qquad (P252.3b)$$

$$R_L\left[1-\frac{c}{c_M}\right]\le P_M=\frac{c_M{}^{-c}}{\delta t} \qquad (P252.3c)$$

or
or

$$R_L\le c_M/\delta t \qquad (P252.3d)$$

where $\hat{c}_L$ is the local canopy water content after interception (in kg m$^{=2}$), and δt is the model timestep (in seconds).

If more water is intercepted during the timestep than it is possible for the local canopy to hold, that is if

$$\hat{c}_L = c + \delta t(R_L - T_{FL}) > c_M \qquad (P252.4a)$$

$$\text{i.e.}\quad c + \left[R_L - R_L\frac{c}{c_M}\right]\delta t > c_M \qquad (P252.4b)$$

or

$$R_L\left[1 - \frac{c}{c_M}\right] > P_M = \frac{c_M - c}{\delta t} \qquad (P252.4c)$$

or

$$R_L > c_M/\delta t \qquad \text{(P252.4d)}$$

then the excess water is added to the local throughfall

$$T_{FL} = R_L \frac{c}{c_M} + \left[c + R_L\left(1 - \frac{c}{c_M}\right)\delta t - c_M\right] / \delta t \qquad \text{(P252.5a)}$$

or

$$= R_L - P_M \qquad \text{(P252.5b)}$$

$$T_{FL} = \begin{cases} R_L - P_M \ \ if\ R_L > c_M/\delta t \\ R_L(c/c_M) \ \ if\ R_L \le c_M/\delta t \end{cases} \qquad \text{(P252.6a)}$$

Combining equations (P252.3-5) gives the final formulation of the local throughfall rate (in kg $m^{-2} s^{-1}$) The throughfall averaged over the fractional area of the grid box where rain occurs is given by

$$T_{FL} = \int_0^{\infty} T_{FL} f(R_L)\, dR_L \qquad \text{(P252.7a)}$$

$$= \int_0^{c_M/\delta t} R_L(c/c_M) f(R_L)\, dR_L + \int_{c_M/\delta t}^{\infty} (R_L - P_M) f(R_L)\, dR_L \qquad \text{(P252.7b)}$$

Evaluating equation (P252.7) by parts using equation (P252.1) gives,

$$T_F = \frac{R}{\epsilon}\left[1 - \frac{c}{c_M}\right]\exp\left[\frac{-\epsilon c_M}{R\delta t}\right] + \frac{Rc}{c_M} \qquad \text{(P252.8)}$$

$$T_F^A = R\left[1 - \frac{c}{c_M}\right]\exp\left[\frac{-\epsilon c_M}{R\delta t}\right] + R\frac{c}{c_M} \qquad \text{(P252.9)}$$

Multiplying by the fractional area, ϵ over which water falls gives the throughfall averaged over the grid box

The rate of change of the canopy water content (in kg $m^{-2} s^{-1}$) as water falls through the canopy s given by

$$(\partial C/\partial t) = (R - T_F^A) \qquad \text{(P252.10a)}$$

and the final updated canopy water content, $\hat{c}$ in kg m^{-2} is

$$\hat{c} = c + (\partial C/\partial t)\,\delta t \qquad \text{(P252.10b)}$$

(b) Surface hydrology

Water from the canopy (the throughfall) on reaching the surface infiltrates the soil at a rate K_{SV} (units kg m^{-2} s^{-1}), equal to the saturated hydrological soil conductivity, modified due to the presence of vegetation (see section 3(iii)). 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}) is given by

$$Y_{SL} = \begin{cases} T_{FL} - K_{SV} & \text{if } T_{FL} > K_{SV} \\ 0 & \text{if } T_{FL} \le K_{SV} \end{cases} \qquad \begin{matrix}(P252.11a)\\(P252.11b)\end{matrix}$$

Combining equation (P252.6a,b) with (P252.11a,b) gives

$$Y_{SL} = \begin{cases} R_L - P_M - K_{SV} \text{ if } R_L > (K_{SV}+P_M) \text{ and } R_L > c_M/\delta t & (P252.11c) \\ 0 \text{ if } R_L \le (K_{SV}+P_M) \text{ and } R_L > c_M/\delta t & (P252.11d) \\ R_L(c/c_M) - K_{SV} \text{ if } R_L > (C_M/c)K_{SV} \text{ and } R_L \le c_M/\delta t & (P252.11e) \\ 0 \text{ if } R_L \le (c_M/c)K_{SV} \text{ and } R_L \le c_M/\delta t & (P252.11f) \end{cases}$$

The surface runoff averaged over the area in the grid box where rainfalls is

$$Y_S = \int_0^{\infty} Y_{SL} f(R_L)\, dR_L \qquad (252.12a)$$

$$= \int_0^{c_M/\delta t} Y_{SL}\, f(R_L)\, dR_L + \int_{c_M/\delta t}^{\infty} Y_{SL} f(R_L)\, dR_L \qquad (P252.12b)$$

Consider the first integral on the r.h.s. of equation (P252.12b). This extends over local rainfall rates between 0 and $c_M/\delta t$. It is clear from equation (P252.11e,f) that if

$$(c_M/c) K_{SV} > c_M/\delta t$$

i.e.

$$K_{SV} \delta t > c$$

then no contribution to surface runoff arises from this term since $\mathbf{R_L > (c_M/c)K_{SV}}$ is required for surface runoff in this region of R_L -space.

Runoff can only occur if

$(C_M/C)K_{SV} < R_L \le c_M/\delta t$

$$\int_0^{c_M/\delta t} Y_{SL}\ f(R_L)\,dR_L = \begin{cases} \int_{(c_M/c)K_{SV}}^{c_M/\delta t} \left(R_L(c/c_M) - K_{SV}\right) f(R_L)\,dR_L & \text{if } K_{SV}\delta\ t \le c \\ 0 & \text{if} K_{SV}\delta t > c \end{cases}$$

$(P252.12c)$, $(P252.12d)$

Hence the first integral in equation (P252.12b) can be expanded as

Consider now the second integral on the rhs of equation (P252.12b) for local rainfall rates greater than $c_M/\delta t$, if

$(K_{sv} + P_M) \le c_M/\delta t$

i.e.

$K_{sv}\ \delta t \le c$

then all rainfall rates above $c_M/\delta t$ can contribute to surface runoff.
However if

$(K_{sv} + P_M) > c_M/\delta t$

i.e.

$K_{SV}\delta t > c$

then only rainfall rates greater than $(K_{SV} + P_M)$ can contribute. Thus the second integral for the parameter space $R_L > c_M/\delta t$ can be expanded as

$$\int_{CM/\delta t}^{\infty} Y_{SL}\ f(R_L)\,dR_L = \begin{cases} \int_{c_M/\delta t}^{\infty} (R_L - P_M - K_{sv})\ f(R_L)\,dR_L & \text{if } K_{SV}\delta t \le c \\ \int_{K_{SV}+P_M}^{\infty} (R_L - P_M - K_{SV})\,f(R_L)\,dR_L & \text{if } K_{SV}\delta t > c \end{cases}$$

$(P252.12e)$, $(P252.12f)$

Combining equations (P252.12c-f) and integrating by parts gives

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

$(P252.13a)$, $(P252.13b)$

Multiplying equations (P252.13a,b) by ϵ gives equations for the grid box average surface runoff

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

$(P252.14a)$ $(P252.14b)$

The rate at which the soil moisture content is increased by throughfall is

$\partial m/\partial t = (T_F^A - Y_S^{A)}$ (P252.15)

(iii) Implementation in the model.

This section describes how the formulation described above of canopy interception and the land surface hydrology is implemented in the unified model.

The canopy capacity (c_M) and soil infiltration rate (K_{sv}) are spatially varying climatological data sets. These are derived using the model land/sea mask and the 1° x1° vegetation and soil data sets of Wilson and Henderson-Sellars (1985) together with soil and vegetation parameters described by Buckley and Warrilow (1988) (see Unified Model Documentation Paper No.70).

The soil infiltration rate is obtained from the saturated hydrological soil conductivity, K_s (in $kgm^{-2}\, s^{-1}$) and the soil infiltration enhancement factor, β_v (dimensionless) accounting for the effects of root systems on the infiltration of surface water into the soil. K_{sv} is given by

$K_{SV} = \beta_V K_S$ (P252.16)

K_{SV} is precalculated and passed into the model as an ancillary field.

Canopy capacity is non-zero everywhere over land except where land ice exists where it is zero. Bare soil is given a small capacity (of 0.001m) to account for surface retention (in puddles etc).

Prior to any calculations of canopy and surface hydrology, water is removed from the canopy water store by evaporation

$\hat{c} = c - \delta t\, E_C$ (only when $E_C \ge 0$) (P252.17)

where $\hat{c}$ is the updated canopy water content and E_C is the evaporation from the canopy calculated by component P245.

The surface runoff of any snowmelt is then calculated from

$$Y_S^{A(SM)} = S_M \exp(-\epsilon K_{SV}/S_M) \qquad \text{(P252.14SM)}$$

The reason for using (P252.14SM) in place of (P252.14) is that snowmelt does not interact with the canopy water store. Note that (P252.14) with c_M replacing c and S_M replacing R gives the same result as (P252.14SM). This fact is used in the code; the same subroutine is called for calculating the surface runoff of snowmelt as for rainfall and condensation. Snowmelt is assumed to cover the whole gridbox, i.e. ϵ is set to1.0. The rate of change of soil moisture content is calculated using (P252.15).

Water from three sources falls through the the canopy before reaching the surface:

Canopy condensation : $E_C < 0$ calculated in component P245.
This is assumed to occur over all the grid box (i.e. ϵ is set to 1.0)

Large-scale rain : $R^{(LS)}$ calculated in component P26.
This is assumed to occur over all of the grid box (i.e. ϵ is set to1.0.)

Convective rain : $R^{(C)}$ calculated in component P27.
This is assumed to fall over 30% of the grid box (i.e. ϵ is set to 0.3.)

Equations (P252.10) and (P252.14) 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 (P252.15) is used to calculate the rate of change of soil moisture content due to throughfall reaching the surface.

The canopy interception calculation is carried out for all land types except land ice where the canopy capacity is defined to be zero. (The infiltration rate is also zero for land ice and any rainfall at the surface is runoff.)

It is possible for water to fall onto the canopy at a grid point which is covered with snow, as 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. In this case the canopy interception calculation is still carried out before the surface runoff calculation as some vegetation may penetrate above the snow layer or bare patches with no snow may occur in the gridbox.

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 (equation (P252.15)) is also summed for each water type and passed across to component P253 (sub-surface hydrology) to update the soil moisture content.

(4) SOIL HYDROLOGY (P253)

Soil moisture content is updated by this component to take account of:

(i) evapotranspiration which removes moisture from the soil, $E_S \geq 0$;

(ii) the throughfall of water from the canopy and snowmelt less surface runoff, $F_W = T_F + S_M - Y_S$, (positive downwards), calculated in P252;

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

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

$$m = m + \delta t(F_W - E_S) \qquad \text{(P253.1)}$$

The rate of removal of soil moisture due to subsurface runoff is then calculated. Unlike the surface runoff, Y_S , which depends on the canopy throughfall or snowmelt rate (see P252), the subsurface runoff, Y_G , is assumed to depend only on the soil moisture content and is a relatively slow process. Y_G , expressed as a mass flux, is parametrized by Eagleson's (1978) empirical formula:

$$Y_G = \begin{cases} 0 & \chi < \chi_W \\ K_S \left((\chi - \chi_W) / (\chi_S - \chi_W) \right)^C & \chi_W \le \chi \le \chi_S \\ K_S & \chi_S < \chi \end{cases} \qquad \text{(P253.2)}$$

where the volumetric soil moisture concentration, χ, is given by

$$\chi = m/\rho_W D_R) + \chi_W \qquad \text{(P253.3)}$$

where K_S is the saturated hydrological conductivity of the soil;
χ_W is the residual value of χ at the wilting point, i.e. that value of χ below which it becomes impossible for vegetation to remove moisture from the soil;
χ_S is the saturation value of χ attained when all the voids between soil particles are filled with water; and D_R is the root depth of vegetation.

K_S , χ , χ and c are climatologically prescribed, geographically varying parameters depending on the soil type and D is a similar parameter depending on vegetation type. Documentation Paper No.70 describes their derivation from the Wilson and Henderson-Sellers (1985) global archive of land cover and soils data.

To save computation time in the model the following parameter combination is precomputed:

$$B_S = K_S / (\chi_S - \chi_W)^C \qquad \text{(P253.4)}$$

Using (P253.2), (P253.3) and (P253.4) Y_G can be expressed as

$$Y_G = B_S \left((m/(\rho_W D_R) \right)^C \qquad \text{(P253.5)}$$

Y_G must be limited so that no more than the total soil moisture content is runoff in a timestep and the l.h.s. of (P253.5) is set to zero if either m or D_R is zero. The algorithm for calculating Y_G is therefore:

$$\text{if } m > 0 \text{ and } D_R > 0 \qquad Y_G = \min \left(m/\delta t,\ B_S \left(m/(\rho_W D_R) \right)^C \right)$$

(P253.6)

otherwise $Y_G = 0$

The soil moisture content is then updated according to

$$m = m - \delta t \, Y_G \qquad \text{(P253.7)}$$

APPENDIX

A comparison of the throughfall formulations of Warrilow et al. (1986) and Shuttleworth (1988).

WSS(86) derived an expression for throughfall based on the work of Rutter etÿal. (1971). It was assumed that water fell evenly over a homogeneous surface canopy. However the arguments can also be developed for a "local" canopy, which is more applicable to the theory presented in section (3(i)) and the work of S(88).

The rate of increase of local canopy water content, c_L , is

$$\rho_W \frac{\partial c_L}{\partial t} = I_{CL} - D_{RL} + E_{CL} \qquad \text{(P25.A.1)}$$

where I_{CL} is the local rate of interception by the canopy (kg m^{-2} s^{-1}),
D_{RL} is the local rate of drainage from the canopy (kg m^{-2} s^{-1}),
and E_{CL} is the local rate of canopy evaporation (kg m^{-2} s^{-1}).

It is assumed that

$$I_{CL} = \Lambda R_L \qquad \text{(P25.A.2)}$$

where Λ is the probability of rain being intercepted by the canopy.

The local throughfall rate is then defined as

$$T_{FL} = (1 - \Lambda)R_L + D_{RL} \qquad \text{(P25.A.3)}$$

Several simplifying assumptions are made:
(i) $\Lambda \rightarrow 1$;
(ii) When the canopy is full ($c_L = c_{ML}$, the local canopy capacity) $D_{RL} = R_L$;
(iii) When the canopy is empty ($c_L = 0$), $D_{RL} = 0$.

The form of D_{RL} chosen to meet criteria (ii) and (iii) is

$$D_{RL} = (c_L / c_{ML})R_L \qquad \text{(P25.A.4)}$$

which with assumption (i) above gives the local throughfall as

$$T_{FL} = (c_L / c_{ML})R_L \qquad \text{(P25.A.5)}$$

If it is assumed that the canopy is homogeneous over the grid box, implying no previous time history of where water previously fell, then

$$T_{FL} = (c/c_M)R_L \qquad \text{(P252.2)}$$

Using this formulation the derivation of section (3(i)) showed that the local throughfall can be written as;

$$T_{FL} = \begin{cases} R_L - P_M \quad if \quad R_L > c_M/\delta t & \text{(P25.A.6a)} \\ R_L(c/c_M) \quad if \quad R_L \le c_M/\delta t & \text{(P25.A.6b)} \end{cases}$$

Equation (P252.6a) accounts for the case of the rate of increase in canopy water content due to interception exceeding that necessary to fill the canopy within a timestep.

S(88) provides an alternative treatment of the interception of falling water within the canopy. The local throughfall rate is given by

$$T_{FL} = \begin{cases} R_L - P_M \quad if \quad R_L > P_M = (C_M - c)/\delta t & \text{(P25.A.6a)} \\ 0 \qquad if \quad R_L \le P_M = (c_M - c)/\delta t & \text{(P25.A.6b)} \end{cases}$$

Thus no throughfall occurs unless the water fall rate is greater than that required to fill the local canopy within a model timestep (P_M). If the water fall rate is less than P_M all water is intercepted by the canopy. This is analogous to the treatment of the infiltration of surface water into the soil (P252.11). However it is not clear whether this is a good analogy.

Consider the infiltration of surface water into the soil. If the infiltration time-scale is less than that for surface runoff it is reasonable to assume that if the water fall rate is less than the infiltration rate no surface runoff occurs. Infiltration occurs in the vertical over a depth of a few meters while runoff occurs in the horizontal with a length scale of tens of meters. Water which is draining horizontally can also infiltrate as it does so.

These scaling arguments cannot be applied to water falling through a vegetative canopy. 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. Hence similar time-scales might be expected for both processes. Also some parts of the canopy may capture water while other parts loose water previously retained. In other parts of the canopy water may fall directly through. The processes is perhaps best thought of as a "random walk" process in the downward direction.

On the basis of the above arguments it seems better to allow throughfall before the canopy is full and so the formulation of Warrilowet al.(1986) is to be preferred.

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(5) Multilayer Soil Hydrology Scheme (subroutine SOIL_HYD) (P254)

Introduction

Section 5 describes the multilayer soil hydrology model which has been designed to replace the single layer scheme in the Unified Model (UM). The new scheme uses the same vertical discretisation as the existing soil thermodynamics module (P242). Its prognostic variables are the moisture content of each soil layer. In addition it calculates the gravitational drainage (or "slow" runoff) from the base of the soil profile and diagnoses the updated value of the mean available soil moisture in the rootzone. The latter is equivalent to the prognostic soil moisture variable of the single layer scheme. However, "slow" runoff is redefined as the drainage from the base of the total soil profile, rather than the drainage from the bottom of the rootzone as in the single layer scheme.

The multilayer scheme is expected to contribute to an improved simulation of the diurnal and seasonal variation in the surface evaporation and runoff fluxes. Single layer models tend to overestimate this variation since they produce too little drainage from the root zone in wet periods (when the soil moisture concentration in the root zone exceeds that below it) and too much drainage in dry periods (when the opposite is true). A multilayer model which extends below the root zone is better able to simulate these changes in drainage since it effectively models the gradient in the soil water tension at the base of the root zone rather than assuming it to be zero (the boundary condition for a single layer model). In particular this means that the multilayer scheme can simulate the partial recharge of the root zone during dry periods by water from below.

Mathematical Model

The soil moisture concentration, θ, obeys the following continuity equation:

$$\frac{d\vartheta}{dt} = -\frac{\partial W}{\partial z} - R \qquad \text{P254.1}$$

where z is the vertical coordinate (positive downwards), W is a flux and R is a sink term which represents the extraction of water by plant roots. The water flux, W, is given by Darcy's Law:

$$W = K\left\{\frac{\partial \psi}{\partial z} + 1\right\} \qquad \text{P254.2}$$

where K is 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 forms for the conductivities, and the soil water tension as a function of the soil moisture concentration. There have been many forms suggested for these curves (e.g. Clapp and Hornberger, 1978, Eagleson, 1978) but the closure used here is that derived by van Genuchten et al. (1991):

$$\psi = \psi_1\, S^{-b}\left\{1 - S^{b+1}\right\}^{\frac{b}{b+1}} \qquad \text{P254.3}$$

$$K = K_s\, S^{L}\left[1 - \left\{1 - S^{b+1}\right\}^{\frac{1}{b+1}}\right]^2 \qquad \text{P254.4}$$

where K_s, Ψ_1, b and L are empirical soil dependent constants. The soil moisture dependent variable, S, is defined by:

$$S = \frac{\vartheta - \vartheta_r}{\vartheta_s - \vartheta_r}$$

where θ_s is the saturation soil moisture concentration and θ_r is the ‘‘residual’’ soil moisture concentration, below which drainage ceases. The sink term, R, is related to the evaporation from the soil, E, by:

$$E = \int_{z_d}^{0} R \, dz \qquad \text{P254.6}$$

where z_d is the root depth. The evaporation is calculated elsewhere in the boundary layer scheme of the UM, so E is treated as an input to the soil hydrology module and determining R amounts to distributing the total evaporative demand vertically within the root zone. It is assumed that the local contribution to the flux is proportional to the root density, C_r, and the soil moisture concentration:

$$R = \frac{C_r \{\vartheta - \vartheta_w\}}{\int_{z_d}^{0} C_r \{\vartheta - \vartheta_w\} \, dz} E \qquad \text{P254.7}$$

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

$$C_r = \mu \, (z_d - z)^2 \qquad \text{P254.8}$$

where μ is a constant.

Discretisation

The multilayer model is the finite difference form of the mathematical model above. A schematic representation of the spatial discretisation is given in figure 1. The prognostic variables of the model are the soil moisture contents within each layer:

$$M_n = \rho_w \, \Delta z_n \, (\vartheta_n - \vartheta_r) \qquad \text{P254.9}$$

where the subscript n denotes the n^{th} soil layer, M_n is its moisture content (kg m^{-2}), θ_n is its volumetric soil moisture

$$M_n^s = \rho_w \, \Delta z_n \, (\vartheta_s - \vartheta_r) \qquad \text{P254.10}$$
$$M_n^w = \rho_w \, \Delta z_n \, (\vartheta_w - \vartheta_r) \qquad \text{P254.11}$$

concentration (m^3 H_2O (m^3 soil)$^{-1}$) and ρ_w is the density of water. Note that the residual soil moisture concentration, θ_r, has been included within the definition of M_n. Likewise the saturation soil moisture content, M_n^s, and the wilting soil moisture content, M_n^w, are:

The variable, S which appears in equations P254.3 and P254.4 therefore reduces to:

$$S_n = \frac{M_n}{M_n^s} \qquad \text{P254.12}$$

The layer soil moistures are advanced using the discretised form of equation P254.1:

$$\frac{\Delta M_n}{\Delta t} = W_{n-1} - W_n - E_n$$

where ΔM_n is the increment to M_n, Δt is the timestep, W_n is the flux from the n^{th} layer to the $(n+1)^{th}$ layer, and E_n is the evaporation which comes from the n^{th} layer (see figure 1). The fluxes, W_n , are derived from a discretised version of the Darcy equation (equation P254.2):

$$W_n = K_{n+1/2} \left\{ \frac{d\psi}{dz} \right]_{n+1/2} + 1 \Bigg\}$$

where the subscript (n+1/2) denotes the values at the boundary between the n^{th} and $(n+1)^{th}$ layers:

$$K_{n+1/2} = K(S_{n+1/2}) \qquad \text{P254.15}$$

$$\left.\frac{d\psi}{dz}\right]_{n+1/2} = \frac{2\ \{\psi(S_n) - \psi(S_{n+1})\}}{\Delta z_n + \Delta z_{n+1}} \qquad \text{P254.16}$$

The functions K(S) and Ψ(S) are given by equations P254.3 and P254.4, and the value of $S_{n+1/2}$ is determined by assuming that S varies linearly between the centres of layers n and (n+1):

$$S_{n+1/2} = \frac{S_n \Delta z_{n+1} + S_{n+1} \Delta z_n}{\Delta z_{n+1} + \Delta z_n} \qquad \text{P254.17}$$

As stated in the previous section, the evapotranspiration from each layer is calculated as a fraction of the total evpotranspiration from the root depth, by assuming that the water uptake from each depth is proportional to the product of the water content and the root density at that depth. From equation P254.7:

$$E_n = \frac{(M_n - M_n^w)\ f_n}{R_N}\ E \qquad \text{P254.18}$$

where f_n is the fraction of the roots in the n^{th} soil layer, M_n^w is the wilting soil moisture of the n^{th} layer and R_N is a normalisation factor:

$$R_N = \sum_n (M_n - M_n^w)\ f_n \qquad \text{P254.19}$$

The fraction of the roots in each layer is calculated by integrating equation P254.8:

$$f_n = \left[\frac{z}{z_d} \left\{ 3 - 3\frac{z}{z_d} + \left(\frac{z}{z_d} \right)^2 \right\} \right]_{z_{n-1}}^{z_n} \qquad \text{P254.20}$$

Numerical Method

The difficulty in solving equations P254.13 to P254.17 numerically arises from the Darcian water fluxes being strongly non linear in the soil moisture. As a result a simple explicit scheme can lead to large inaccuracies or instability unless a short timestep or large layer depths are used, neither of which is desirable. Instead some means of taking account of the sub-timestep variation in the fluxes is required. Unfortunately, a fully implicit scheme is not possible since the non linearity precludes tridiagonal inversion. The scheme used here amounts to a first order correction to the explicit flux:

$$W_n = \frac{1}{2} \left(W_n^{(0)} + W_n^{(1)} \right) \qquad \text{P254.21}$$

where $W_n^{(0)}$ is the explicit flux and $W_n^{(1)}$ is a correction term. Both of these fluxes are derived from equations P254.14 to P254.17, but whereas $W_n^{(0)}$ is calculated from the values of S_n and S_{n+1} at the previous timestep,$W_n^{(1)}$

is calculated from the sub-timestep values, S'_n and S'_{n+1}, where:

$$S'_n = \frac{M_n - W_n^{(0)} \Delta t}{M_n^s} \qquad \text{P254.22}$$

$$S'_{n+1} = \frac{M_{n+1} + W_n^{(0)} \Delta t}{M_{n+1}^s} \qquad \text{P254.23}$$

Thus $W_n^{(1)}$ is effectively the water flux that would occur at the end of the timestep (i.e. at the beginning of the next timestep) if the explicit flux, $W_n^{(0)}$, was used to transfer water from the n^{th} to the $(n+1)^{th}$ layer. The flux given by equation P254.21 can therefore be viewed as the mean of the fluxes at the beginning and end of the model timestep.

Model accuracy is also improved by updating the soil moistures sequentially beginning at the lowest layer and moving upwards. The procedure is as follows:

(i) *Calculate* W_n

(ii) *Update* M_{n+1} : $M_{n+1} \rightarrow M_{n+1} + W_n \Delta t$ P254.24

(iii) *Update* M_n : $M_n \rightarrow M_n - (W_n + E_n) \Delta t$ P254.25

(iv) *return to (i) with n replaced by* $(n-1)$

Thus for each layer n the flux to the layer below (W_n) is calculated and used to update the soil moisture contents of both layers (M_n and M_{n+1}). The layer label is then reduced by 1 and the process is repeated. Thus the flux, W_{n-1}, from layer n-1 to layer n is calculated using the value of M_n which has already been updated by W_n.

Boundary Conditions

The boundary conditions for the multilayer model are as for the existing single layer scheme:

$$W_0 = P_f + S_m + Y_s \qquad \text{P254.26}$$
$$W_N = K_N$$

where W_0 is the flux which infiltrates at the surface, W_N is the drainage from the lowest layer (N), K_N is the hydraulic conductivity of this layer, and P_f, S_m and Y_s are the fluxes of throughfall precipitation, snowmelt and surface runoff respectively. (Note that the evaporation does not appear in equation P254.26, instead it appears within the sink term (E_n) of equation P254.13 which represents the direct extraction of water by plant roots). The surface fluxes are calculated elsewhere in the land surface scheme, so W_0 is treated as an input to the soil hydrology scheme.

Parameter Values

The soil parameters of equations P254.3 and P254.4 are listed in table 1a. These are taken from table 4 of van Genuchten et al. (1991). The values used in the current scheme are shown in table 1b for comparison. Other soil parameters such as the critical soil moisture concentration below which evaporation is water limited, and the soil thermal characteristics are common to the two schemes (see Buckley and Warrilow, 1988). Note also that the "fine", "medium" and "coarse" soil textures of the UM have been identified with the "clay", "loam" and "loamy sand" classifications of van Genuchten et al.(1991).

The multilayer soil hydrology scheme is designed to use the same vertical discretisation as the soil thermodynamics model, i.e. 4 layers with thicknesses given by:

$$\Delta z_n = \xi_n \sqrt{\frac{2 \lambda_s}{\omega_1 C_s}} \qquad \text{P254.28}$$

where λ_s is the thermal conductivity of the soil, C_s is the volumetric heat capacity of the soil, ξ_n is the thickness of the n^{th} layer relative to the top layer thickness and ω_1 is a characteristic frequency. The values of ξ_n and ω_1 are those which were originally chosen to optimise the performance of the soil thermodynamics scheme (Warrilow et al., 1986):

$$\omega_1 = 3.55088 \times 10^{-4}\ s^{-1}$$
$$\xi_1 = 1.0$$
$$\xi_2 = 3.908$$
$$\xi_3 = 14.05$$
$$\xi_4 = 44.65$$

These parameters imply top layer thicknesses of 3.79, 3.45 and 3.79 cm for fine, medium and coarse soils respectively.

Code Organisation
The multilayer soil hydrology module consists of a main routine (SOIL_HYD) which calls an external subroutine (DARCY) twice for each soil layer. For clarity SOIL_HYD has been broken up into three separate sections.

Section 1 is concerned with the initialisation and redefinition of variables. It begins with the calculation of the layer thicknesses (P254.28) and the depths of the layers between boundaries. The fraction of roots in each soil layer is then evaluated (P254.20) and this is used to calculate the weighted evaporation from each layer (numerator of P254.18) and the corresponding normalisation factor (P254.19). At this stage the soil moisture variables are converted from the water which is {*available*} for evapotranspiration, M_n^{av} , to the {*total* } water which is active in the subsurface hydrology, M_n:

$$M_n = M_n^{av} + \rho_w \Delta z_n V_w \qquad \text{P254.29}$$

where ρ_w is the density of water, Δz_n is the thickness of the n^{th} soil layer, and $V_w = \theta_w - \theta_r$ is the difference between the wilting soil moisture concentration and the residual soil moisture concentration. The corresponding saturation values for each soil layer are also evaluated:

$$M_n^s = \rho_w \Delta z_n V_s \qquad \text{P254.30}$$

where $V_s = \theta_s - \theta_r$. The top boundary condition is then defined (P254.26) and the evaporation from each soil layer is normalised (P254.18)

Section 2 of SOIL_HYD is concerned with the task of calculating the Darcian fluxes between layers and updating the soil moisture contents appropriately.The procedure explained in the section on "Numerical Method" is followed. The fluxes between adjacent soil layers (P254.14) are calculated using the subroutine DARCY. For each layer boundary:
(i) the explicit flux between the $(n-1)^{th}$ and n^{th} layer ($W_{n-1}^{(0)}$) is calculated (call DARCY)
(ii) the corresponding increment to the moisture content of the $(n-1)^{th}$ layer is calculated:

$$\Delta M_{n-1} = (W_n^0 + E_{n-1})\ \Delta t \qquad \text{P254.31}$$

(iii) a check is made to ensure that the moisture content of the $(n-1)^{th}$ layer remains positive and less than the saturation value after the increment:

$$0 < M_{n-1} + \Delta M_{n-1} < M_{n-1}^s$$

(iv) the corresponding increment to the moisture content of the nth layer is calculated:

$$\Delta M_n = -W_{n-1}^0 \; \Delta t$$

(v) a check is made to ensure that the moisture content of the nth layer remains positive and less than the saturation value after the increment:

$$0 < M_n + \Delta M_n < M_n^s$$

(vi) the "half timestep" moisture contents are defined (P254.22, P254.23)
(vii) the "half timestep" flux between the (n-1)th and nth layer ($W_{n-1}^{(1)}$) is calculated (call DARCY)
(viii) the corrected flux between the (n-1)th and nth layer is calculated (P254.21)
(ix) the corresponding increment to the moisture content of the (n-1)th layer is calculated:

$$\Delta M_{n-1} = (W_{n-1} + E_{n-1}) \; \Delta t \qquad \text{P254.33}$$

(x) a check is made to ensure that the moisture content of the (n-1)th layer remains positive and less than the saturation value after the increment (see (iii))
(xi) the corresponding increment to the moisture content of the nth layer is calculated:

$$\Delta M_n = -W_{n-1} \; \Delta t \qquad \text{P254.34}$$

(xii) a check is made to ensure that the moisture content of the nth layer remains positive and less than the saturation value after the increment (see (v))
(xiii) the moisture content of the nth layer is updated (P254.24)
(xiv) the moisture content of the (n-1)th layer is updated (P254.25)
Layer boundaries are dealt with sequentially from the lower boundary upwards.

Section 3 of SOIL_HYD is concerned with the diagnosed variables. Firstly the soil moisture variables are converted back to *available* values:

$$M_n^{av} = M_n - \rho_w \Delta z_n V_w \qquad \text{P254.35}$$

The mean available moisture in the rootdepth, m, is calculated from the layer values:

$$m = \sum_n a_n M_n^{av} \qquad \text{P254.36}$$

where a_n is the fraction of the nth soil layer which is contained within the rootdepth. Finally the slow runoff is defined as the flux of water across the lower boundary of the model:

$$Y_g = W_N \qquad \text{P254.37}$$

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Tables

Soil Type	θ_s	θ_r	K_s (m/day)	b	Ψ_1 (m)	L
Clay	0.38	0.068	0.0480	11.111	1.250	0.5
Loam	0.43	0.078	0.2496	1.786	0.278	0.5
Loamy Sand	0.41	0.057	3.5020	0.781	0.081	0.5

Table 1a : Soil parameters for use in equations P254.3 and P254.4 (taken from table 4 of van Genuchten et al. (1991)).

Soil Type	θ_s	θ_w	K_s (m/day)	c
Fine	0.49	0.180	0.0312	11.0
Medium	0.46	0.105	0.1560	8.0
Coarse	0.42	0.055	0.6960	5.0

Table 1b : Soil parameters as used in the MO single layer soil hydrology scheme (Warrilow et al., 1986).

Figure 1 : A schematic representation of the discretisation used in the multilayer soil hydrology scheme. Symbols are as given in the text: M represents a layer soil moisture content, W represents a Darcian flux, and E represents an evaporative loss.