The parametrization of rainfall interception in GCMs

By A. JOHANNES DOLMAN and DAVID GREGORY

1Institute of Hydrology, Wallingford
2Hadley Centre of Climate Prediction and Research, Meteorological Office, Bracknell

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SUMMARY

Experiments with a one-dimensional version of the Meteorological Office's 11-layer GCM are used to derive a simple calibrated subgrid parametrization of rainfall interception by the Amazonian rain forest. Two interception parametrizations, both incorporating subgrid variability of precipitation within a GCM grid box are presented and tested, and compared with observations made at a single location. The 1-D model is a fully interactive atmospheric column and incorporates all the GCM physics but is forced at its boundaries by climatic data. The interception losses appear to be sensitive to what fraction of area in the GCM grid box is covered by the rain and to the precise formulation of the interception process. The implications for the parametrization of the wet area in a GCM grid box are discussed and suggestions are made as to how to parametrize the fraction of wetted area in a GCM grid box.

1. INTRODUCTION

Experiments with general circulation models (GCMs) have demonstrated that the atmospheric circulation and rainfall patterns over the continents may be influenced by the interaction of the land surface and the lowest layers of the atmosphere (Rowntree 1988; Mintz 1984). Deforestation experiments with GCMs have shown that substantial changes in atmospheric circulation and rainfall may result from changing the tropical rain forest of the Amazon into pasture or savanna (e.g. Lean and Warrilow 1989; Dickinson 1989; Shukla et al. 1990). The results of these experiments generally suggest a weakening in intensity of the hydrological cycle if large areas of forest are cleared. Since about 50% of the rainfall within the Amazon basin is derived from local evaporation (Shuttleworth et al. 1984), the prediction of the likely climatic effects of deforestation in the Amazon will ultimately depend on the realism of the particular land-surface parametrization scheme used (e.g. Henderson-Sellers 1987).

Recently a number of land-surface parametrization schemes have been proposed which incorporate a considerable degree of complexity in their description of the land-surface–atmosphere interaction (Sellers et al. 1986; Dickinson 1984; Nolihan and Planton 1989). These models are usually designed and calibrated with observations from single sites rather than with area-averaged data (e.g. Sellers et al. 1989). They are nevertheless forced with grid area-averaged output variables, such as rainfall and humidity, from the atmospheric circulation model to which they are coupled.

This may lead to problems in the parametrization of spatially heterogeneous, subgrid physical processes. For instance, rainfall interception is known to be sensitive to the temporal structure of rainfall (Lloyd 1990). In the case of the Amazon rain forest, where convection is the dominant rainfall producing mechanism, the grid rainfall rate is a composite of several storms, which may vary in surface coverage, duration and intensity. GCMs at present do not include parametrizations of this variation and tend to overestimate interception loss and underestimate dry canopy evaporation (Lean and Warrilow 1989, Abramopoulos et al. 1988) when the results are compared with observations (e.g. Shuttleworth 1988b).

Pitman et al. (1990) report on an experiment with a stand-alone version of the Biosphere Atmosphere Transfer Scheme (BATS, Dickinson et al. 1986), in which a new
rainfall interception parametrization proposed by Shuttleworth (1988b) is used to describe the subgrid variability of rainfall in a GCM grid. They claim that there is substantial sensitivity to the details of this subgrid parametrization, as is shown by the run-off-evaporation regime of tropical rain forest. However, the results of their study are of limited value as the atmospheric forcing is prescribed rather than calculated, as in a GCM.

This paper describes experiments with a 1-D version of the Meteorological Office’s 11-layer GCM exploring the sensitivity of a predicted Amazonian climate to the way in which the wet canopy evaporation from the tropical rain forest is represented. The 1-D model represents a single large area, roughly the size of a GCM grid square, and includes all the physics routines of the Meteorological Office’s GCM (Slingo 1985). At the boundaries, horizontal advection of heat, moisture and momentum is prescribed for a specific location using data from Oort (1983). The use of a 1-D model in the context of land-surface atmospheric modelling is relatively novel (Warrilow et al. 1986; Koster and Eagleson 1990). It allows investigation of the interaction of the land surface with the atmosphere and is therefore particularly useful in the development of land-surface parametrizations. Results are compared with observations obtained in the Anglo-Brazilian field experiment in the Reserva Ducke near Manaus, Amazonia (Shuttleworth 1988b).

2. 1-D MODEL

The 1-D model was developed by Warrilow et al. (1986) from the Meteorological Office’s 11-layer GCM (Slingo 1985) used for climate research. Sigma coordinates (pressure over its surface value) are used in the vertical, and all physical processes included in the GCM are modelled: interactive radiation with four cloud types (low, medium, high and convective), a mass-flux convection scheme, large-scale precipitation, and boundary-layer and surface turbulent exchange (Slingo 1985), together with a land-surface scheme, described briefly in the next section. The model is integrated forward with a time-step of 5 minutes with the radiation scheme being called every 2 hours.

The main feature of the 1-D model which distinguishes it from other similar models is its treatment of heat and moisture advection. Warrilow et al. (1986) provide a detailed description of the formulation and its derivation; a brief description is given below. The model is assumed to represent a circular area with radius \( \Delta n \). The local rate of change of a quantity \( X \), averaged over the area, due to large-scale advection can be written as

\[
\frac{\Delta X}{\Delta t_d} = \frac{|V_n| (X_r - X)}{\Delta n} - \frac{\omega \Delta X}{\Delta p}
\]

(1)

where \( |V_n| \) is the absolute value of the horizontal velocity in the direction of the positive gradient of \( X \) across the area, \( \omega \) is the vertical \( p \)-velocity, \( p \) is the pressure and \( X_r \), a reference value of \( X \) outside the area under consideration. Knowledge of \( V_n \), \( \omega \) and \( X_r \) is required if the forcing due to the large-scale advection is to be specified. A Gaussian probability distribution for these quantities is assumed, with means and standard deviation being taken from the global data-set compiled by Oort (1983). To specify the moisture reference-profile, dew-point depression, \( D \), is used. Owing to a lack of available data the standard deviation of \( D \) is linearly related to that of temperature. Because the mean values of \( X \), relate to the centre of the area under study, while the reference profile is required at the edge, a correction is applied to allow for the gradient of a quantity, \( \delta X \),
across the area. The velocity component, \( V_p \), parallel to the positive gradient of \( X \), across
the area is also specified for use in the boundary-layer scheme.

Data used in the advective forcing scheme are shown in Tables 1(a) and 1(b) for
January and July for the central Amazon. A sinusoidal interpolation procedure is used
to derive values for other months. A random-number generator is used to specify the
exact forcing on any one day, in association with the assumed Gaussian distributions
described above. Values are interpolated from day to day to prevent large step-changes
in the forcing that is being applied to the model. The only tuning of the forcing
representation is in the relationship between the standard deviation of dew-point depres-
sion and temperature, which is chosen to give the correct mean rainfall for a given area.
Initial model results indicated a lack of seasonal variation in the rainfall over the Amazon.
Vertical velocities from the Oort data were modified to apply a correction for this lack,
in accordance with the balance of observed precipitation and evaporation. The revised
figures are included in Table 1.

3. LAND-SURFACE SCHEME

The land-surface scheme currently in use in the GCM is depicted in Fig. 1. Surface
temperature is calculated using a four-layer soil-temperature model. This model was
designed to produce a reasonable surface temperature response to radiative forcing
in the frequencies from a half day to a year (Warrilow et al. 1986). It is important to stress
that the soil-vegetation layer in this representation is treated as a single solid body, with
a single surface temperature equal to the temperature of the top layer, which represents
the temperature of the soil and the vegetation. No effort is made to calculate different
surface temperatures for soil and vegetation as in some of the more complex recent land-
surface schemes. The single value of surface temperature is used in calculations of
both radiative and turbulent exchange. The main parameters used in this scheme are
summarized in Table 2.

The fluxes of latent and sensible heat are calculated from the expression:

\[
F = -\delta X |V_{11}| C_x
\]

(2)

where \( F \) is the flux of either the sensible or latent heat, \( \delta X \) the change in temperature
or humidity from the surface to the lowest atmospheric level, \( C_x \) a bulk-transfer coefficient
and \( V_{11} \) the horizontal wind velocity at the lowest model level. \( C_x \) is a complex function
of height, atmospheric stability (as expressed by the bulk Richardson number) and
surface roughness. In the case of evaporation from vegetated land surfaces the latent-
heat flux from the soil-vegetation surface is calculated from the expression:

\[
E = \frac{\delta q}{(r_a + r_s)} \]

(3)

with \( r_a \), an aerodynamic resistance, equal to \( (C_v |V_{11}|)^{-1} \), \( r_s \) the surface resistance to
water-vapour transfer, and \( \delta q \) the change in specific humidity from the saturated surface
to the lowest model layer. Equation (3) specifies the actual evaporation rate. The surface
resistance in Eq. (3) is obtained by reducing the 'unstressed' surface resistance, \( r_{su} \),
according to the relationship:

\[
r_s = \frac{r_{su}}{\beta} \]

(4)

where \( \beta \) is a reduction factor defined by the lesser of 1 and \( (\Theta - \Theta_u)/(\Theta_c - \Theta_u) \), where
### TABLE 1(a)  1-D Model-forcing data, for Manaus, Brazil (3°S, 60°W), January

<table>
<thead>
<tr>
<th>Level</th>
<th>11</th>
<th>10</th>
<th>9</th>
<th>8</th>
<th>7</th>
<th>6</th>
<th>5</th>
<th>4</th>
<th>3</th>
<th>2</th>
<th>1</th>
</tr>
</thead>
<tbody>
<tr>
<td>$T$ (°C)</td>
<td>299.0</td>
<td>296.0</td>
<td>291.0</td>
<td>283.5</td>
<td>273.0</td>
<td>260.0</td>
<td>245.0</td>
<td>229.0</td>
<td>213.0</td>
<td>194.0</td>
<td>213.0</td>
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<tr>
<td>$\sigma T$ (°C)</td>
<td>2.0</td>
<td>2.0</td>
<td>2.0</td>
<td>2.0</td>
<td>1.7</td>
<td>1.5</td>
<td>1.5</td>
<td>1.5</td>
<td>1.5</td>
<td>2.2</td>
<td>3.0</td>
</tr>
<tr>
<td>$D$ (°C)</td>
<td>4.0</td>
<td>4.0</td>
<td>5.5</td>
<td>6.5</td>
<td>12.5</td>
<td>8.0</td>
<td>8.0</td>
<td>6.0</td>
<td>8.0</td>
<td>10.0</td>
<td>30.0</td>
</tr>
<tr>
<td>$\delta T$ (°C)</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>1 ($10^{-4}$)</td>
</tr>
<tr>
<td>$\delta D$ (°C)</td>
<td>1</td>
<td>1</td>
<td>-60</td>
<td>-150</td>
<td>-270</td>
<td>-430</td>
<td>-570</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>1 ($10^{-4}$)</td>
</tr>
<tr>
<td>$V_x$ (m s$^{-1}$)</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>$V_y$ (m s$^{-1}$)</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>$\sigma V_x$ (m s$^{-1}$)</td>
<td>2</td>
<td>2.2</td>
<td>2.5</td>
<td>3.1</td>
<td>3.6</td>
<td>4.2</td>
<td>4.8</td>
<td>5.5</td>
<td>6.1</td>
<td>6.7</td>
<td>5.7 ($10^{-5}$)</td>
</tr>
<tr>
<td>$\omega$ (mb s$^{-1}$)</td>
<td>-75</td>
<td>-150</td>
<td>-300</td>
<td>-570</td>
<td>-600</td>
<td>-570</td>
<td>-465</td>
<td>-330</td>
<td>-165</td>
<td>75</td>
<td>105 ($10^{-7}$)</td>
</tr>
<tr>
<td>$\sigma \omega$ (mb s$^{-1}$)</td>
<td>1</td>
<td>2</td>
<td>4</td>
<td>7.6</td>
<td>8.0</td>
<td>7.6</td>
<td>6.2</td>
<td>4.4</td>
<td>2.2</td>
<td>1.0</td>
<td>1.4 ($10^{-4}$)</td>
</tr>
</tbody>
</table>

### TABLE 1(b)  1-D Model-forcing data, for Manaus, Brazil (3°S, 60°W), July

<table>
<thead>
<tr>
<th>Level</th>
<th>11</th>
<th>10</th>
<th>9</th>
<th>8</th>
<th>7</th>
<th>6</th>
<th>5</th>
<th>4</th>
<th>3</th>
<th>2</th>
<th>1</th>
</tr>
</thead>
<tbody>
<tr>
<td>$T$ (°C)</td>
<td>297.0</td>
<td>294.5</td>
<td>289.0</td>
<td>282.0</td>
<td>271.5</td>
<td>258.0</td>
<td>243.0</td>
<td>228.0</td>
<td>208.0</td>
<td>197.0</td>
<td>218.0</td>
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<tr>
<td>$\sigma T$ (°C)</td>
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<td>2.0</td>
<td>2.0</td>
<td>1.3</td>
<td>1.3</td>
<td>1.6</td>
<td>1.9</td>
<td>1.5</td>
<td>1.5</td>
<td>1.5</td>
<td>2.5</td>
</tr>
<tr>
<td>$D$ (°C)</td>
<td>1.0</td>
<td>3.0</td>
<td>4.5</td>
<td>10.0</td>
<td>12.0</td>
<td>12.5</td>
<td>14.0</td>
<td>15.0</td>
<td>24.0</td>
<td>14.0</td>
<td>35.0</td>
</tr>
<tr>
<td>$\delta T$ (°C)</td>
<td>27</td>
<td>14</td>
<td>4.5</td>
<td>4.5</td>
<td>4.5</td>
<td>4.5</td>
<td>4.5</td>
<td>2.4</td>
<td>0.63</td>
<td>0.63</td>
<td>0.63 ($10^{-4}$)</td>
</tr>
<tr>
<td>$\delta D$ (°C)</td>
<td>-23</td>
<td>-61</td>
<td>-120</td>
<td>-200</td>
<td>-310</td>
<td>-440</td>
<td>-620</td>
<td>2.4</td>
<td>0.63</td>
<td>0.63</td>
<td>0.63 ($10^{-4}$)</td>
</tr>
<tr>
<td>$V_x$ (m s$^{-1}$)</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
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<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>$V_y$ (m s$^{-1}$)</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>$\sigma V_x$ (m s$^{-1}$)</td>
<td>2</td>
<td>2.2</td>
<td>2.5</td>
<td>3.1</td>
<td>3.6</td>
<td>4.4</td>
<td>5.0</td>
<td>5.6</td>
<td>6.0</td>
<td>5.5</td>
<td>5.0 ($10^{-5}$)</td>
</tr>
<tr>
<td>$\omega$ (mb s$^{-1}$)</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0 ($10^{-7}$)</td>
</tr>
<tr>
<td>$\sigma \omega$ (mb s$^{-1}$)</td>
<td>0</td>
<td>1.0</td>
<td>3.0</td>
<td>4.0</td>
<td>4.2</td>
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<td>4.2</td>
<td>3.0</td>
<td>1.0</td>
<td>1.0</td>
<td>1.0 ($10^{-4}$)</td>
</tr>
</tbody>
</table>

*T*: average monthly temperature; $\sigma T$: standard deviation of temperature; $D$: average monthly dewpoint depression; $\delta T$ and $\delta D$: gradient of temperature and dewpoint depression, respectively from the edge to the 1-D model area centre; $V_x$ and $V_y$: average horizontal wind velocity components; $\omega$: vertical $p$-velocity and $\sigma \omega$ the standard deviation of the vertical $p$-velocity.
Figure 1. Schematic diagram of the land surface scheme as used in the Meteorological Office's GCM and in the original version of the 1-D model (redrawn after Warrilow and Buckley 1989). $R_w$: shortwave radiation; $R_l$: longwave radiation; $\alpha$: albedo, $R_c$: convective precipitation; $R_i$: large-scale precipitation; $E_w$: wet canopy evaporation; $E_d$: dry canopy evaporation; $H$: sensible heat flux; $T_i$: temperature of the $i$th soil layer; $D$: rooting depth; $\Theta$: soil moisture content; $Q_s$: surface run-off; $Q_g$: gravitational drainage; $C$: canopy water content; $T_f$: through-fall.

**Table 2. Parameters used in the land surface scheme for tropical forest**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface albedo (%)</td>
<td>12.25</td>
</tr>
<tr>
<td>Surface resistance (s$m^{-1}$)</td>
<td>115</td>
</tr>
<tr>
<td>Saturation storage capacity (mm)</td>
<td>0.7</td>
</tr>
<tr>
<td>Fractional vegetation cover</td>
<td>0.9</td>
</tr>
<tr>
<td>Rooting depth (m)</td>
<td>1.5</td>
</tr>
<tr>
<td>Surface roughness (m)</td>
<td>2.0</td>
</tr>
</tbody>
</table>

$\Theta$ is the soil moisture content and the subscripts c and w refer to the critical level of soil moisture and the wilting point, respectively. The unstressed surface resistance is set at 115 s$m^{-1}$ in accordance with the results of Dolman et al. (1991). This formulation is somewhat different from that used in the full GCM, where the reduction factor $\beta$ works directly upon the evaporation, the actual amount being $\beta E_p$, where $E_p$ is the potential evaporation rate. However, since the physiological response of the vegetation to an increase in soil moisture deficits is more realistically expressed as an increase in surface resistance, Eqs. (3) and (4) are employed.

Evaporation from a wet canopy is calculated from Eq. (3), with the surface resistance set to zero. The canopy store is filled by either large-scale or convective precipitation and condensation, and emptied by through-fall and evaporation. Through-fall, $T$, is modelled assuming a dependence on the depth of water on the canopy, $C$; the canopy storage capacity, $S$; and the rainfall rate, $P$, viz.

$$T = \left(\frac{C}{S}\right)P.$$  

(5)

Note that this formulation implies through-fall during the wetting-up phase of the canopy.
In the case of a canopy store not completely filled, a fraction $1 - C/S$ of the grid area is assumed to be transpiring freely, and a fraction $C/S$ to be evaporating intercepted rainfall.

4. EXPERIMENTS WITH THE STANDARD CANOPY SCHEME

Previous work on forest evaporation has emphasized the need for a separate description of wet and dry canopy evaporation (Shuttleworth 1989)—an observation which applies equally well to the Amazonian tropical rain forest. In a GCM grid area, however, a single set of meteorological variables has to represent both dry and wet conditions, and particular problems arise with respect to the averaging techniques used to calculate the dry and wet canopy contributions.

A series of 1-D model experiments was designed to compare results obtained using the present land-surface description of the GCM, with results obtained from a locally-calibrated model (Shuttleworth 1988b). Where references in this paper are made to observations of Shuttleworth (1988b), they refer to the synthesized secondary data of Shuttleworth (1988b). The results of a 1-D model experiment, in which the above description (Table 2) of the rain forest was used (A1), is compared with the results from Shuttleworth (1988b) (Table 3). In these experiments the 1-D model was integrated over one year.

The modelled and observed seasonal distribution of rainfall is shown in Fig. 2. The seasonal distribution is obtained from a 15-year record of rainfall (1971–1985) in the

![Figure 2. Comparison of observed (—–) (average of 1971–1985) and predicted (——) monthly rainfall (mm d$^{-1}$).](image)

Reserva Ducke reserve. The agreement between observed and modelled precipitation is reasonable, with rainfall underestimated in the first half of the year (by 15 to 20%) and overestimated in the second half (by 10 to 15%). Differences between model and observed values could be the result of errors in the response of the convection scheme, the random nature of the forcing or the particular climate forcing data. Within the present context, the agreement is sufficiently good to justify proceeding with the testing of the interception parametrizations. The model reproduces the results of the GCM experiments, which used essentially the same parametrization (Lean and Warrilow 1989), with the interception loss accounting for 60% of the total evaporation. The partitioning of the total evaporation between interception (wet canopy evaporation) and transpiration
(dry canopy evaporation) is shown in Fig. 3. Lean and Warrilow found wet canopy evaporation to account for 71% of total evaporation, which was overestimated by 15% compared with observations. The 1-D model simulations underestimate total evaporation by about 10%, but this difference is probably insignificant when interannual variation in precipitation and evaporation are taken into account. The model uses about 90% of the available energy for evaporation, which agrees with the results of Shuttleworth (1988b); approximately 50% of the annual rainfall is returned to the atmosphere by the process of evaporation, in agreement with the observations. It is worth noting that in the 1-D model experiment the value of the saturation storage capacity was reduced from 2.5 mm (Lean and Warrilow 1989) to 0.7 mm. This did not produce a substantially different interception loss—27% and 26% respectively, when the model was integrated for one month. Shuttleworth (1988b) and Lloyd et al. (1988), however, report considerable sensitivity of the modelled interception loss to changes in the value of the saturation storage capacity. The reduction in storage capacity in the 1-D experiment is apparently not sufficient to allow the canopy to become more saturated, and to start producing through-fall by canopy drip.

![Figure 3. Partitioning of total evaporation into interception, \( E_i \), and transpiration, \( E_t \), for the original parametrization. Modelled rainfall is also shown (---).](image)

5. EXPERIMENTS WITH REVISED CANOPY SCHEMES

The overestimation of interception loss is likely to be caused by the neglect of spatial variability in the convective rainfall pattern over a single GCM grid. The time-distribution of grid-averaged rainfall can be compared directly with the observed time-distribution at Manaus, as this comparison involves only the mean quantities and does not involve higher-order moments (Fig. 4). The modelled distribution agrees very well with the observations over a 2-year period at the Reserva Ducke site (Lloyd 1990). The agreement suggest that the 1-D model reproduces correctly both the average forcing and the dominant convective nature of rainfall in the central Amazon basin. However, to allow for the spatial variation in this rainfall and for the fact that at any one time the GCM grid is unlikely to be fully covered by a single storm, it is necessary to take these factors into account in the formulation of canopy interception.

Shuttleworth (1988a) presents a parametrization of this kind. The local rainfall rate in the rain-covered part of the grid is assumed to follow a negative exponential probability
distribution e.g.

\[ f(P_1) = \left( \frac{\mu}{P} \right) \exp \left( -\frac{\mu P_1}{P} \right) \]  \hspace{1cm} (6)

with \( P \) and \( P_1 \) the model’s grid-average rainfall rate and the local rainfall rate respectively, and \( \mu \) the proportion of the grid covered with rain. Within the area the local through-fall is taken to be

\[ T_1 = \begin{cases} P_1 - C_m & \text{if } P_1 \Delta t > S - C \\ 0 & \text{if } P_1 \Delta t < S - C \end{cases} \]  \hspace{1cm} (7a)

where \( C_m = (S - C)/\Delta t \) and \( \Delta t \) is the time-step. It can be seen that no through-fall occurs locally until the canopy is saturated. It should be noted that this description differs in that respect from that used by Warrilow et al. (1986). Integrating over the area where rain occurs and averaging over the grid-box, Shuttleworth (1988a) demonstrated that

\[ T = P \exp \left( -\frac{\mu C_m}{P} \right). \]  \hspace{1cm} (7b)

Recently an analogous scheme, based on the original Warrilow et al. (1986) assumption of the dependence of the through-fall on rainfall rate (Eq. 5), has been developed and included in the new Meteorological Office Unified Modelling (UM) system (Gregory and Smith 1990) i.e.

\[ T_1 = \begin{cases} P_1 - C_m & \text{if } P_1 \Delta t > S \\ P_1 \left( \frac{C}{S} \right) & \text{if } P_1 \Delta t < S \end{cases} \]  \hspace{1cm} (8a)

with the grid-box mean through-fall given by

\[ T = P \left( 1 - \frac{C}{S} \right) \exp \left( -\frac{\mu S}{P \Delta t} \right) + P \frac{C}{S} \]  \hspace{1cm} (8b)

The results of 1-D model experiments with both these schemes are also given in Table 3. Cases A2–A4 refer to the WJS scheme (Eq. 7), whilst cases A5–A7 refer to the new
### Table 3. 1-D Model Interception and Transpiration Results

<table>
<thead>
<tr>
<th>Model version</th>
<th>'Observed'</th>
<th>A1</th>
<th>A2</th>
<th>A3</th>
<th>A4</th>
<th>A5</th>
<th>A6</th>
<th>A7</th>
</tr>
</thead>
<tbody>
<tr>
<td>Precipitation (mm)</td>
<td>2593</td>
<td>2418</td>
<td>2347</td>
<td>2367</td>
<td>2332</td>
<td>2215</td>
<td>2359</td>
<td>2332</td>
</tr>
<tr>
<td>Interception (mm)</td>
<td>373</td>
<td>766</td>
<td>316</td>
<td>437</td>
<td>127</td>
<td>222</td>
<td>406</td>
<td>125</td>
</tr>
<tr>
<td>Percentage</td>
<td>14</td>
<td>32</td>
<td>14</td>
<td>19</td>
<td>6</td>
<td>10</td>
<td>17</td>
<td>5</td>
</tr>
<tr>
<td>Transpiration (mm)</td>
<td>1020</td>
<td>512</td>
<td>873</td>
<td>760</td>
<td>1036</td>
<td>950</td>
<td>799</td>
<td>1038</td>
</tr>
<tr>
<td>Percentage</td>
<td>39</td>
<td>21</td>
<td>37</td>
<td>32</td>
<td>44</td>
<td>43</td>
<td>34</td>
<td>45</td>
</tr>
</tbody>
</table>

Models:  
A1: 'Old' GCM Parametrization  
A2: WJS with $\mu = 0.1/0.5$  
A3: WJS with $\mu = 0.3/1.0$  
A4: WJS with variable wetted area  
A5: Unified model parametrization, $\mu = 0.1/0.5$  
A6: Unified model parametrization, $\mu = 0.3/1.0$  
A7: Unified model parametrization with variable wetted area  
($\mu = 0.1$ or $\mu = 0.3$ applies to convective rainfall, $\mu = 0.5$, $\mu = 1$ applies to large-scale rainfall)

UM scheme (Eq. 8). Three situations, related to the fractional area covered by a rainstorm, are considered for each parametrization. Cases A2 and A5 refer to 30% coverage for a convective precipitation event, and 100% for large-scale rain. Cases A3 and A6 refer to 10% and 50% coverage.

Cases A4 and A7 use a variable wetted area related to the average intensity of rainfall as observed in the Reserva Ducke experiment. As can be seen from Fig. 4, the occurrence of convection shows a marked diurnal variation, so the constant value which has been given to the wetted area will necessarily ignore this variation and produce either too weak a rainfall over too large an area, or too intense a rainfall over too small an area, depending on whether the fixed area refers to night-time or mid-day convective activity. A more physically based estimate of the fraction can be given by assuming that the diurnal variation can be modelled by incorporating an explicit dependence of the fraction of the grid covered with rain on the grid average rainfall rate. This can be written as

$$\mu = \frac{P}{P_r}$$

where $P$ is the grid average rainfall rate produced by the model and $P_r$ an observed average rainfall rate as measured in the Reserva Ducke experiment (Lloyd 1990). From measurements taken at Manaus, $P_r$ is given as 5.2 m h$^{-1}$.

The parametrization of interception using Eqs. (7) and (8) suffers from a dependence of the model on the time-step through the incorporation of the rainfall amount in the exponential terms. As a consequence the through-fall decreases when the time-step of the model becomes very small. Typical values for interception loss range from 23% to 17% for a time-step of 5 minutes and 20 minutes respectively, with the values changing less rapidly with increasing time-step. To avoid this problem, the calculation of through-fall was made using amounts adjusted to represent hourly values, and the actual through-fall amounts were adjusted back again at each time-step, just before the canopy water budget was drawn up. Another possibility may be to allow $\mu$ to vary with the time-step, in which case both the time-steps in Eqs. (7b) and (8b) will cancel out. Both methods ensure that the through-fall percentages remain independent of the time-step, and take into account only spatial variability in rainfall.

A further adjustment was made in the models to ensure that evaporation from a wet canopy occurred only from the actual wetted area, with the remainder of the grid...
transpiring freely. In this case evaporation from wet canopies took place in only a fraction, \( \mu \), of the grid-box area.

From Table 3 it becomes clear that there is substantial variation in the predicted interception loss—and corresponding partitioning of evaporation between the parametrization of wet and dry canopy interception. All parametrizations which allow for a spatial distribution of rainfall show a marked improvement in the prediction of interception and transpiration. Both parametrizations show a decreased interception loss when the area covered with rain is reduced from 30% to 10% (Cases A2, A3 and A5, A6). The Shuttleworth (1988a) parametrization appears to give a slight overestimation of interception loss when compared with the value 9% ±3.6 quoted by Lloyd et al. (1988), when a value of \( \mu = 0.1 \) is used (A2). The result obtained by Shuttleworth (1988b), using a calibrated model, indicates a loss of 14%. Indeed the model predictions obtained by Lloyd et al. (1988) indicate a slight inability to model the observed interception loss, as they predict 12.3% and 11.3%, respectively, for the two tested models; both these values are however within measurement error of the observations. The value 14% (Shuttleworth 1988b) is probably best regarded as an upper limit for interception loss in this area. The Unified Model parametrization predicts a 10% interception loss for a value of \( \mu = 0.1 \), which is in agreement with observations (A5). The partitioning of evaporation for this case is shown in Fig. 5, which may be compared with Fig. 3 to assess the improvement in the evaporation prediction of the model. However, relating the value of \( \mu \) to the rainfall rate results in a substantial underestimation of the interception loss for both the Shuttleworth (A4) and UM (A7) parametrizations. This suggests that the parametrization of \( \mu \) gives too small a wetted area in this case. Both parametrizations give remarkably similar results in this case.

To test whether the new parametrization improves the sensitivity of the interception loss to a change in saturation storage capacity, the storage capacity was changed by 50%. This produced a change in interception loss by 15% when using the new UM parametrization, with \( \mu = 0.1/0.5 \), which can be compared with a change of 4% for the original scheme, and a change of 25% observed in the stand-alone interception modelling by Shuttleworth (1988b) and Lloyd et al. (1988). This indicates that the new parametrization simulates the interception of tropical rainforest on a regional scale much more accurately than the previous parametrization in which no allowance was made for the spatial distribution of rainfall.

![Figure 5](image-url)  
**Figure 5.** Partitioning of total evaporation into interception, \( \mathbb{I} \), and transpiration, \( \mathbb{E} \), for the UM parametrization with \( \mu = 0.1/0.5 \) (run A5). Modelled rainfall is also shown (---).
6. Discussion

The 1-D model experiments on wet canopy evaporation show the importance of including the effect of spatial inhomogeneity in the parametrization of canopy processes in GCMs. Although some of the previous GCM deforestation experiments predicted the correct total daily evaporation, the partitioning between dry and wet canopy evaporation appeared to be incorrect. Incorporation of a more physically realistic subgrid parametrization reproduces the correct proportions of dry and wet canopy evaporation, and increases the sensitivity to changes in storage capacity.

The percentage total evaporation is relatively constant in all 1-D model experiments (approximately 50%). This suggests that the total amount of modelled evaporation is strongly controlled by the radiation input and that changes in the parametrizations of the interception scheme affect only the partitioning between dry and wet canopy evaporation. Although the total amount of moisture returned to the atmosphere in this way remains the same, the timescale in which this occurs is different. Transfer of moisture through transpiration is more gradual than transfer through interception and involves longer timescales, because of the delay in soil moisture storage. This may in turn affect the intensity of run-off or flooding. From this perspective alone it is important to model the balance between wet and dry canopy evaporation correctly.

Shuttleworth (1988b) makes the observation that total evaporation is relatively constant throughout the year at about 110 mm monthly. The 1-D model reproduces this phenomenon very clearly, although the almost complete lack of seasonal variation may have been amplified by the slight inability of the model to reproduce the observed rainfall pattern.

The 1-D model experiments show that caution must be exercised when applying subgrid parametrizations straightforwardly, without trying to relate the model results to actual measurements. A similar comment applies to the use of 1-D vegetation atmosphere models with a prescribed atmospheric forcing. With an interception scheme similar to that used in experiments A2–A4, Pitman et al. (1990) showed that a change in \( \mu \) from 0.5 to 0.1 (for convective storms) gave a 50% reduction in total evaporation. However, important feedback effects can be prevented by prescribing the atmospheric forcing. Pitman et al. (1990) comment that feedback effects may have a strong modifying impact on the sensitivity of their model. The use of a fully 1-D model here demonstrates the importance of such feedback on the control of surface fluxes: only a 1% increase in total evaporation is seen between experiments A2 and A3 when \( \mu \) is increased from 0.1 to 0.3. A much larger change is found in the partitioning between wet and dry canopy evaporation.

The parametrization in which through-fall is allowed to occur before saturation gave slightly better results than the Shuttleworth (1988b) scheme. In point models of interception loss for tropical rainforest, through-fall before canopy saturation is generally not allowed, and is only observed in very small amounts in the field. Indeed, owing to the high intensity of rainfall in the humid tropics and the relatively short time required to saturate a canopy, the through-fall from unsaturated canopies is usually very small. The somewhat surprisingly good performance of the UM scheme can be explained by realizing that it effectively allows a distribution of the canopy storage over the grid from saturated to unsaturated, while the grid average canopy storage may be unsaturated. By allowing a certain part of the grid to be fully saturated, and the other part not, through-fall from the grid canopy can still occur; whereas in the Shuttleworth scheme, this would not be possible until the complete grid average canopy storage is saturated.

The present experiments show the complexity involved in the modelling of the
subgrid variability in GCM grids and underline the need to compare GCM model predictions with observations. Such observations could include image analysis of rainfall or cloud cover from satellite data. It may be possible in the future to relate the fractional coverage to convective cloud amount or cloud liquid-water content, but the difficulty of doing so, should not be underestimated; it may be difficult to assess the relation of model variables like rainfall rate to wetted area in a GCM grid.

It is speculative to predict the consequences of the incorrect partitioning between dry and wet canopy evaporation for the interpretation of past GCM deforestation experiments. The present results indicate that the likely climatic effects, such as a weakening of the intensity of the hydrological cycle, may have been amplified in the GCM experiments by the overestimation of the interception component of total evaporation. To what extent the compensating effect of dry and wet canopy evaporation will disguise this effect can, however, only be resolved by new, more realistic GCM experiments which use calibrated land-surface models.

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REFERENCES


