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# Modelling rainfall interception by a lowland tropical rain forest in northeastern Puerto Rico

J. Schellekens[a,*], F.N. Scatena[b], L.A. Bruijnzeel[a], A.J. Wickel[a,1]

[a]Faculty of Earth Sciences, Vrije Universiteit, De Boelelaan 1085, 1081 HV Amsterdam, Netherlands
[b]United States Department of Agriculture, Forest Service, International Institute of Tropical Forestry, P.O. Box 25000, San Juan, PR 00928-5000, USA

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**Abstract**

Recent surveys of tropical forest water use suggest that rainfall interception by the canopy is largest in wet maritime locations. To investigate the underlying processes at one such location—the Luquillo Experimental Forest in eastern Puerto Rico—66 days of detailed throughfall and above-canopy climatic data were collected in 1996 and analysed using the Rutter and Gash models of rainfall interception. Throughfall occurred on 80% of the days distributed over 80 rainfall events. Measured interception loss was 50% of gross precipitation. When Penman–Monteith based estimates for the wet canopy evaporation rate (0.11 mm $h^{-1}$ on average) and a canopy storage of 1.15 mm were used, both models severely underestimated measured interception loss. A detailed analysis of four storms using the Rutter model showed that optimizing the model for the wet canopy evaporation component yielded much better results than increasing the canopy storage capacity. However, the Rutter model failed to properly estimate throughfall amounts during an exceptionally large event. The analytical model, on the other hand, was capable of representing interception during the extreme event, but once again optimizing wet canopy evaporation rates produced a much better fit than optimizing the canopy storage capacity. As such, the present results support the idea that it is primarily a high rate of evaporation from a wet canopy that is responsible for the observed high interception losses. 

*Keywords*: Interception loss; Modelling; Tropical rain forest; Puerto Rico

## 1. Introduction

A survey of tropical forest water use (evapotranspiration, ET) reveals that the highest values (2000–2400 mm $y^{-1}$) are observed at continental edge and island locations of high rainfall (Bruijnzeel, 2000). Much lower values (ca. 1200–1450 mm $y^{-1}$) occur at mid-continental equatorial sites. Closer scrutiny of the data reveals that at wet ‘maritime’ sites, rainfall interception (evaporation from a wet canopy, $E_i$) may constitute a very large portion—up to 70%—of ET compared to only 20–25% at more ‘continental’ sites (Bruijnzeel, 2000). Similar contrasts have been reported for coastal and inland locations at temperate latitudes (Pearce et al., 1976; Calder, 1977; Gash and Stewart, 1977). Such findings reinforce the contention that deforestation at continental edge and island locations is likely to have a greater effect on river flow

* Corresponding author. Fax: +31206462457.
*E-mail addresses:* schj@geo.vu.nl (J. Schellekens), f_scatena@upr1.upr.clu.edu (F.N. Scatena), brul@geo.vu.nl (L.A. Bruijnzeel), bwickel@hydrolab.arsusda.gov (A.J. Wickel)
[1] Present address: USDA-ARS-Hydrolab, 104 BLDG 007 BARC-W, Beltsville, MD 20705, USA.

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than in the case of mid-continental sites (Shuttleworth, 1989).

Although evidence for high to very high values of $E_i$ ($\geq 30\%$ of incident rainfall) in tropical maritime conditions is increasing steadily (Clements and Colon, 1975; Gilmour, 1975; Scatena, 1990; Cavelier et al., 1997; Clark et al., 1998; Hafkenscheid et al., 1998), very few of these studies have used high resolution recording throughfall gauges in combination with above-canopy climatic observations to explain these high values. In the three studies for which detailed observations are available (Calder et al., 1986; Hafkenscheid et al., 1998; Schellekens et al., submitted), the average wet canopy evaporation rate ($E_w$) inferred from measurements of incident rainfall ($P$), throughfall ($TF$) and stemflow ($SF$) greatly exceeded the rates predicted by the Penman–Monteith equation (Monteith, 1965). Interestingly, no such discrepancies were observed in similar studies in more continental tropical lowland rain forests (Lloyd et al., 1988; Hutjes et al., 1990).

Various solutions have been proposed to bridge the gap between measured and predicted $E_i$. For example, in West Java Calder et al. (1986) not only used a reduced (optimized) value for the aerodynamic resistance to evaporation $r_a$ ($5\,s\,m^{-1}$), but they also proposed a much larger value for the canopy saturation value ($S$, 4–5 mm) than the 0.75–1.3 mm range that is usually obtained for tropical rain forests using more traditional techniques (e.g. Jackson, 1975; Murdiyarso, 1985; Lloyd et al., 1988; Scatena, 1990; Elsenbeer et al., 1994; Jetten, 1996). Calder et al. (1986) based their choice of a high value of $S$ on laboratory experiments by Herwitz (1985). Lloyd et al. (1988), on the other hand, considered such high values artefacts and reverted to traditionally derived values of $S$ for their forest in central Amazonia.

Several studies have documented exceedingly high annual rainfall interception totals—up to 50% of gross precipitation—in a lowland rain forest in maritime northeastern Puerto Rico (Scatena, 1990; Schellekens et al., submitted). To gain more insight in the underlying causes and mechanisms, the present paper examines the performance of two widely used rainfall interception models—the Rutter model (Rutter et al., 1971; 1975 and the analytical model (Gash, 1979; Gash et al., 1995)—using detailed measurements of rainfall, throughfall and above-canopy weather conditions for the 66-day period between 5 May and 9 July 1996. Particular attention is paid to the effect of variations in the magnitude of $S$ and $r_a$ on predicted $E_i$ with a view to examine the suitability of the solutions proposed by Calder et al. (1986) under the extreme climatic conditions prevailing at the study site.

## 2. The study site

The 6.34 ha Bisley II catchment—in which the present study was conducted—is located on the northeastern slopes of the Luquillo Experimental Forest (LEF), northeastern Puerto Rico. The catchment is situated at 18°18′N, 65°50′W, between 265 and 456 m above mean sea level and is covered with Tabonuco (*Dacryodes excelsa*) type forest. The Atlantic Ocean is located at less than 10 km from the outlet of the catchment. The forest was hit by Hurricane Hugo in September 1989 (Scatena and Larsen, 1991). As a result, the Bisley forest has an irregular upper canopy, with the highest Tabonuco trees located on ridge tops; an understory of palms and woody vegetation, and ground level herbs and scrubs (Lugo and Scatena, 1995). However, long-term observations of throughfall, litterfall and canopy biomass development (Scatena et al., 1993, Scatena, unpublished data) as well as recent leaf area index (LAI) estimates (estimated at 5.9 around the *TF* gutters on the basis of light attenuation measurements; Holwerda, 1997) indicated that conditions in May–July 1996 were comparable to those prevailing before the hurricane disturbance. The average *LAI* of a similar forest nearby was estimated at 6.4 although it ranged from 1.95 in ravines to 12.1 on ridge tops and slopes carrying tall emergents (Odum et al., 1970).

The climate at the site is maritime tropical, type A2m according to the Köppen classification. The annual rainfall is about 3500 mm of which about 70% is brought by the north-easterly trade-winds (García-Martinó et al., 1996). Rainfall is distributed fairly evenly within the year. In general, May and November are the wettest months with about 385 mm each while the January–March period is relatively 'dry' with 200 mm per month on average. Rainfall events at nearby El Verde (450 m a.s.l.) are generally small (median of daily rainfall 3 mm) but numerous (267 rain days per year) and of relatively low intensity ($<5$ mm h$^{-1}$; Brown et al., 1983). Mean monthly temperatures in the Bisley area vary little during the year (24°C in December–February, 27.5°C in July–August). Average humidity is high at 84–90%. Average monthly wind speeds are $<2$ m s$^{-1}$ with, again, little seasonal variation (Brown et al., 1983).

## 3. Instruments

The period under investigation lasted from 5 May until 9 July 1996. Precipitation (*P*) was measured every 5 min above the canopy at 26 m on a scaffolding tower using a Texas Instruments tipping bucket rain gauge (TE525LL-L; 0.254 mm per tip). Back up was provided by an adjacent totalizing rain gauge that was emptied once a week. *TF* was measured throughout the catchment with 20 randomly positioned but non-roving collectors (143 cm$^2$ surface area) which were emptied weekly. These fixed gauges have been in operation since June 1987 (Scatena, 1990). A roving gauge technique (Lloyd and Marques-Filho, 1988) had been adopted in the beginning but after initial comparison of the performance of the fixed and roving gauges did not show significant differences, only the fixed network was maintained (cf. Brouwer, 1996). In addition, *TF* was recorded continuously using three flat-bottomed, sharp-rimmed steel gutters ($6 \times 300$ cm) placed at a steep angle to minimize losses via splash. Each gutter was equipped with a tipping bucket cum logger system manufactured at the Vrije Universiteit, Amsterdam. To minimize wetting losses and prevent clogging by debris, the gutters were cleaned and sprayed with a silicone solution every week. The records obtained with the gutters were converted to areal averages every time the manual gauges were emptied, using a weighting procedure based on the relative magnitude of the surface areas of the respective gauge types. This correction also included any changes in the calibration of the tipping buckets (Marsalek, 1981). Stemflow was not measured in this study. An average value of 2.3% of *P* as derived in the earlier study (Scatena, 1990) was used throughout.

Wind speed and direction were measured at 26 m using a Met-One 014A Wind Set. Incoming short-wave radiation (350–1100 nm) was measured by a Li-Cor LI-200X pyranometer. Air temperature and humidity at 26 m were determined using a Vaisala HMP35C probe which was protected against direct sunlight and precipitation by a Model 41002 radiation shield. All climatic data other than rainfall were stored at hourly intervals in a Campbell Scientific Ltd 21X data logger and retrieved weekly for further processing.

## 4. Methods

### 4.1. Derivation of canopy parameters

Both the Rutter and the analytical (Gash) model of interception require knowledge of the canopy structure as described by the following parameters: *S* (mm) is the canopy storage capacity—the amount of water left on the canopy when rainfall and *TF* have ceased; *p* the free *TF* coefficient—the proportion of the rain which falls to the ground without striking the canopy; $S_t$ (mm) the trunk water capacity; and $p_t$ the proportion of rain that is diverted to stemflow. Earlier work on stemflow in the Tabonuco forest at El Verde by Clements and Colon (1975) suggested a trunk water capacity value of 0.01 mm. Given the much higher tree biomass at El Verde compared to Bisley (Odum et al., 1970; Scatena et al., 1993), $S_t$ at Bisley will be negligibly small. Therefore, evaporation from the trunks was not considered further and a value equal to the average stemflow fraction of 0.023 was used for $P_t$ (Scatena, 1990).

The canopy storage capacity (*S*) of the forest was determined using the methods of Jackson (1975), Gash and Morton (1978), and Rowe (1983). The Jackson (1975) approach also allows the derivation of the free *TF* coefficient (*p*) as the slope of the regression between *P* and *TF* for storms that are too small to fill the canopy storage. It is assumed that evaporation losses during these small storms are negligible. The *TF* vs. *P* graph steepens beyond the inflexion point where canopy saturation is reached. In the Jackson approach *S* is determined as:

$$S = P_{\text{inflection}} - TF_{\text{inflection}} \tag{1}$$

where the inflection point is defined as the intersection of the two regressions between *TF* and *P* for storms that do and those that do not fill the canopy storage (Jackson, 1975).

Gash and Morton (1978) drew a straight line of near-unit slope ($1 - p_t = 0.0977$ in this case) through the upper points of the *TF* vs. *P* graph for storms $\geq 1.5$ mm, assuming the highest points to represent conditions with minimal evaporation losses. *S* is then obtained as the negative intercept of the line with the *TF* axis.

Rowe (1983) proposed a simple event-based canopy water balance approach to estimates. Assuming that the canopy storage will be filled by storms $\geq 2$ mm and that evaporation losses will be negligible for storms $\leq 4$ mm, *S* can be estimated from:

$$S = P - TF - SF \tag{2}$$

### 4.2. The Rutter model

Rutter et al.(1971, 1975) developed a numerical model of rainfall interception based on a running water balance of the canopy (and trunks, if required). Originally developed for a Corsican pine forest in Great Britain, it has been applied to numerous forest types around the world, including tropical forests (Calder et al., 1986; Lloyd et al., 1988; Hutjes et al., 1990). A brief description of the model is given below.

The change in amounts of water stored on the canopy is determined by the proportion of the rain that hits the canopy, the drainage from the canopy, and evaporation of intercepted water:

$$\mathrm{d}C/\mathrm{d}t = (1 - p - p_t)R - E_w - D \qquad \text{when } C \geq S$$

$$\mathrm{d}C/\mathrm{d}t = (1 - p - p_t)R - (C/S)E_w - D$$

$$\text{when } C < S \tag{3}$$

where *C* (mm) is the amount of water on the canopy, *R* (mm min$^{-1}$) the rainfall intensity, $E_w$ (mm min$^{-1}$) the evaporation rate from the wet canopy, *D* (mm min$^{-1}$) the drainage rate of water stored on the canopy with *S*, *p*, and $p_t$ defined as before. $E_w$ is usually estimated using the Penman–Monteith equation with the canopy resistance parameter ($r_s$) set to zero (Monteith, 1965):

$$\lambda E_w = \frac{\Delta A + \rho C_p VPD/r_a}{\Delta + \gamma} \tag{4}$$

where $\lambda E_w$ (W m$^{-2}$) is the latent heat flux from the wet canopy, *A* (W m$^{-2}$) the available energy, $\Delta$ (Pa K$^{-1}$) the slope of the temperature–vapour pressure relationship at temperature *T*, $\gamma$ (Pa K$^{-1}$) the psychrometric constant, VPD (Pa) the vapour pressure deficit, $r_a$ (s m$^{-1}$) the aerodynamic resistance, $\rho$ (kg m$^{-3}$) the density of dry air and $C_p$ (J kg$^{-1}$ K$^{-1}$) the specific heat of air at constant temperature *T*. The aerodynamic resistance ($r_a$) is generally calculated from the wind speed and the surface roughness, using a relationship which assumes that effects of atmospheric stability and bluff body forces are either compensating or negligible (Thom, 1975):

$$r_a = \frac{\left(\ln \frac{z - d}{z_0}\right)^2}{k^2 u_{(z)}} \tag{5}$$

where *k* is von Kármán's constant, 0.41, *z* (m) the measurement height above the ground surface, *d* (m) the zero plane displacement height, $z_0$ (m) the roughness length and $u_{(z)}$ (m s$^{-1}$) the wind speed at height *z*. The displacement length *d* was taken as $0.86h_v$ ($h_v$ being the vegetation height in m) and the roughness length $z_0$ as $0.06h_v$ ($h_v$ was 20 m, see Schellekens et al., submitted, for details).

When the canopy is calculated as being only partially wet ($C < S$), the estimated evaporation rate is reduced such that (see Eq. (3)):

$$E_{\mathrm{reduced}} = (C/S)E_w \tag{6}$$

An exponential function is used to calculate drainage from the canopy:

$$D = D_0 e^{b(C-S)} \tag{7}$$

where *D* (mm min$^{-1}$) is the drainage rate, $D_0$ (mm min$^{-1}$) the drainage rate when $C = S$ (saturated canopy) and *b* an empirical parameter. To be consistent with the definition of *S* given earlier, *D* is set to zero when $C < S$. This also avoids the problem of a small but finite drainage when $C = 0$ (Calder, 1977). The Corsican pine stand to which the Rutter model was applied originally, had a value for $D_0$ of 0.0019 mm min$^{-1}$. For other canopies with a different LAI, this would be expected to be $0.0019\mathrm{LAI}/\mathrm{LAI_c}$, where $\mathrm{LAI_c}$ is the leaf area index of Rutter's Corsican pine forest. Assuming *S* is directly proportional to LAI, the saturated drainage rate for other canopies can be

Table 1
Formulation of the components of interception loss according to Gash (1979)

| Component of interception loss | Formulation |
|---|---|
| For $m$ small storms ($P_g < P'_g$) | $(1 - p - p_t)\sum_{j=1}^{m} P_{g,j}$ |
| Wetting up the canopy in $n$ large storms ($P_g \geq P'_g$) | $n(1 - p - p_t)P'_g - nS$ |
| Evaporation from saturated canopy during rainfall | $\bar{E}/\bar{R}\sum_{j=1}^{n}(P_{g,j} - P'_g)$ |
| Evaporation after rainfall ceases for $n$ large storms | $nS$ |
| Evaporation from trunks in $q$ storms that fill the trunk storage | $qS_t$ |
| Evaporation from trunks in ($m + n - q$) storms that do not fill the trunk storage | $p_t\sum_{j=1}^{m+n-q} P_{g,j}$ |

obtained from the value of $S$ (1.05) for the Corsican pine stand of Rutter et al. (1975) (Rutter and Morton, 1977; Lloyd et al., 1988; Jetten, 1996). A similar procedure can be followed to adjust the value of $b$. Thus, when recalculating to a 'standard' $S$ of 1 mm, we obtain:

$$D_0 = 0.0019/1.05 \times S = 0.0018S \tag{8}$$

$$b = 3.7 \times 1.05/S = 3.86S \tag{9}$$

In the present study, the set of equations for the Rutter model was solved using an explicit finite difference approximation of $\mathrm{d}C/\mathrm{d}t$.

### 4.3. The analytical (Gash) model

The analytical model of rainfall interception is based on Rutter's numerical model (see Gash, 1979; Gash et al., 1995, for a full description). The simplifications that Gash (1979) introduced allow the model to be applied on a daily basis, although a storm-based approach will yield better results in situations with more than one storm per day (Pearce and Rowe, 1981). The amount of water needed to completely saturate the canopy ($P'$: Gash, 1979) is defined as:

$$P' = \frac{-\bar{R}S}{\bar{E}_w}\ln\left[1 - \frac{\bar{E}_w}{\bar{R}}(1 - p - p_t)^{-1}\right] \tag{10}$$

where $\bar{R}$ (mm h$^{-1}$) is the average precipitation intensity on a saturated canopy, $\bar{E}_w$ (mm h$^{-1}$) the average evaporation from the wet canopy and with the vegetation parameters $S$, $p$ and $p_t$ as defined previously. The model uses a series of expressions to calculate the interception loss during different phases of a storm (Table 1). An analytical integration of the total evaporation and rainfall under saturated canopy conditions is then done for each storm to determine average values of $\bar{E}_w$ and $\bar{R}$. The total evaporation from the canopy (i.e. the total interception loss $E_i$) is calculated as the sum of the components listed in Table 1. Interception losses from the stems are calculated for days with $P \geq S_t/p_t$ (Gash, 1979). Because $p_t$ and $S_t$ are small in the present case (Section 4.1), evaporation from the stems was neglected.

In applying the analytical model, saturated conditions are assumed to occur when the hourly rainfall exceeds a certain threshold. Often a threshold of 0.5 mm h$^{-1}$ is used (Gash, 1979; Gash et al., 1980; Lloyd et al., 1988). $\bar{R}$ is calculated for all hours when the rainfall exceeds the threshold to give an estimate of the mean rainfall rate onto a saturated canopy. $\bar{E}_w$ is then calculated using Eqs. (4) and (5) (as in the Rutter model) during the same hours.

Gash (1979) has shown that in a regression of interception loss on rainfall (on a storm basis) the regression coefficient should equal $\bar{E}_w/\bar{R}$. Assuming that neither $\bar{E}_w$ nor $\bar{R}$ vary considerably in time, $\bar{E}_w$ can be estimated in this way from $\bar{R}$ in the absence of above-canopy climatic observations (Rowe, 1983; Bruijnzeel and Wiersum, 1987; Dykes, 1997). Values of $\bar{E}_w$ derived in this way generally tend to be (much) higher than those calculated with Eqs. (4) and (5) (Gash et al., 1980; Bruijnzeel and Wiersum, 1987; Waterloo et al., 1999).

## 5. Results

### 5.1. Measured rainfall, throughfall and interception

Between 5 May and 9 July 1996, a total of 852 mm of rain was received, of which 227 mm occurred in an extreme event on 13 May. During this period 53 days had rain, of which 33% fell at night. The average amount of rainfall per event was 10.7 mm and the average duration was 3 h 34 min, resulting in an average rainfall intensity of 3.0 mm h$^{-1}$. However, the distribution of the 80 events was strongly skewed

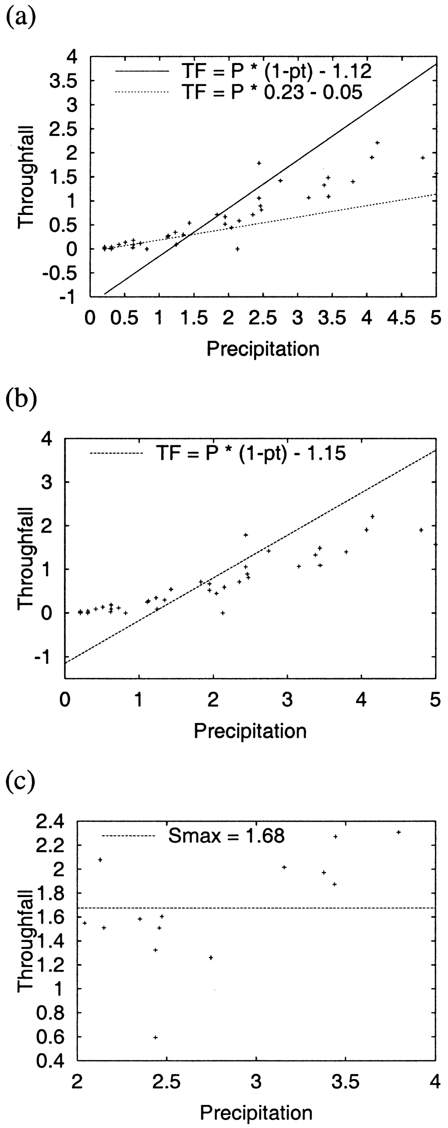



Fig. 1. Estimation of the canopy parameters $S$ and $p$ using the methods of: (a) Jackson (1975) $S = 1.12$, $p = 0.23$; (b) Gash and Morton (1978) $S = 1.15$; and (c) Rowe (1983) $S = 1.68$. *Y*-axes show throughfall (*TF*, in mm) and *x*-axes show rainfall (*P*, in mm).

(Schellekens et al., submitted) and therefore the use of median rather than average values is more appropriate. Events were separated by 3 h without rain to allow for the complete drying up of the canopy. Median rainfall, duration and intensity—based on event analysis—were 3.3 mm, 2 h 25 min and 1.85 mm h$^{-1}$, respectively. However, the choice of the temporal resolution of the precipitation data used for the determination of the intensity greatly influenced the results. For example, the average rainfall intensity derived from the 5-min records—using only intervals with recorded precipitation—was 7.2 mm h$^{-1}$ (median 2.5 mm h$^{-1}$) while the average rainfall intensity derived from hourly values using the same method was only 2.8 mm h$^{-1}$ (median 1.0 mm h$^{-1}$) i.e. closer to the event-based estimate of 3.0 mm h$^{-1}$ (median 1.85 mm h$^{-1}$). Wadsworth (1949) estimated the number of storms per year in the LEF at about 1600 with an average storm duration of 19 min. These figures were also derived from high temporal resolution data. The implications of such differences are discussed later.

A total amount of 388 mm of *TF* was recorded, implying a total interception loss ($E_i$) of 444 mm (52% of P) when assuming an average stemflow fraction of 2.3%. There was no significant difference in the *TF* vs. *P* relationships for day-time and night-time events (Schellekens et al., submitted).

### 5.2. Forest structural parameters

The results obtained with the various methods for determining the canopy capacity *S* and the free *TF* coefficient *p* are summarized in Fig. 1. Values of *S* derived with the methods of Jackson (1975) and Gash and Morton (1978) were very similar at 1.12 and 1.15 mm, respectively. The less refined approach of Rowe (1983) gave a value of 1.68 mm which may well have been influenced by evaporation during the storms. A value for *S* of 1.15 mm was adopted for further use in the models. The value established for *p* (0.23) seemed rather high in view of the dense forest of the study plot, but was used throughout the calculations.

### 5.3. Application of the Rutter model

Fig. 2 shows the cumulative *TF* pattern predicted by the Rutter model for the 66-day study period. Various runs were made. In the first run (model A, top) a value of *S* of 1.15 mm was used in combination with a value for $E_w$ according to Eqs. (4) and (5) (0.11 mm h$^{-1}$). It is immediately apparent from these graphs that *TF* is seriously overestimated (i.e. $E_i$ is underestimated), particularly during the extreme event of 13 May. Total *TF* predicted by model A was

Table 2
Sensitivity analysis of the Rutter model to changes in the main parameters (model scenario B, ‘optimum fit’, excluding the big storm of 13 May 1996) showing the changes in predicted *TF* after increasing or decreasing a parameter by 10%

| Parameter (initial value) | Resulting *TF* | | | | |
|---|---|---|---|---|---|
| | Measured | + 10% | − 10% | % change + | % change − |
| $E_w$ (2.8 mm $h^{-1}$) | 296.3 | 287.6 | 305.9 | −2.9 | 3.3 |
| $S$ (1.15 mm) | 296.3 | 292.2 | 300.8 | −1.4 | 1.5 |
| $p$ (0.23) | 296.3 | 301.9 | 290.9 | 1.9 | −1.8 |
| $p_t$ (0.023) | 296.3 | 295.4 | 297.2 | −0.3 | 0.3 |

752 mm, an overestimation of 94%. Excluding the extreme event and repeating the exercise still produced an overestimation of 79% (*TF* total of 532 mm, Fig. 2a). In model B, the value of $S$ was retained at 1.15 mm, but $E_w$ was optimized—using a trial and error procedure—to a (very high) value of 2.80 mm $h^{-1}$ (Fig. 2b). The effect of a five-fold increase of $S$ to 5.75 mm (cf. Calder et al., 1986) but retaining the value of $E_w$ as estimated via Eqs. (4) and (5) (model C) was examined next. As shown in Fig. 2c, the prediction by model C is somewhat better than in the case of model A but total *TF* is still overestimated considerably (estimated total *TF* 496 mm, an overestimation of 30%). An optimized value for $E_w$ of 1.1 mm $h^{-1}$ was required to fit measured and predicted *TF* totals when using the increased value for the storage capacity (5.75 mm, model D). The sensitivity of the present application of the Rutter model to changes in the magnitudes of $E_w$, $S$, $p$ and $p_t$ is explored in Table 2. The sensitivity of the model to changes in $E_w$ in particular is clearly borne out by these simulations.

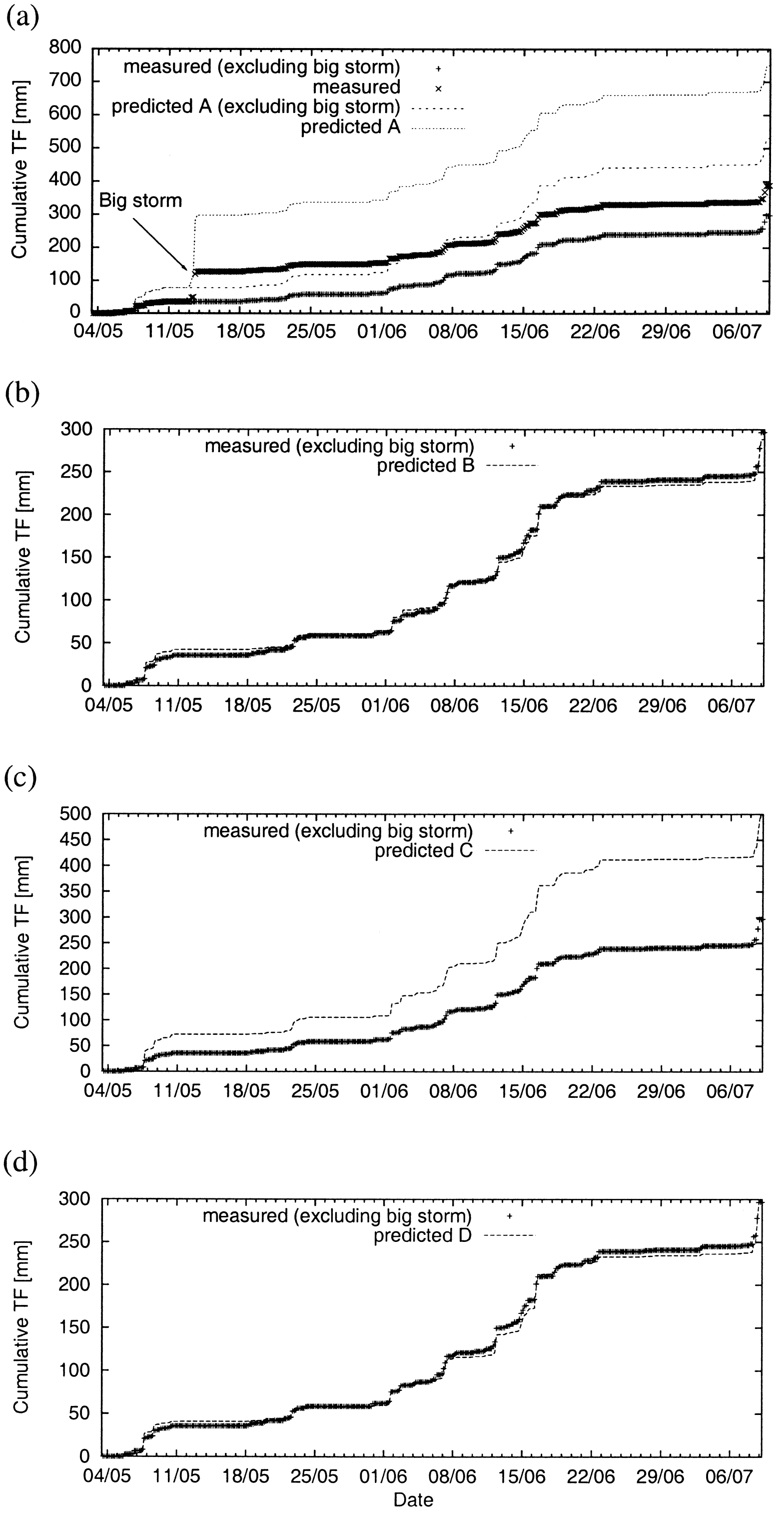
A more detailed look at the performance of the Rutter model during four individual storms is presented in Figs. 3 and 4. The first storm (no. 29, 8.55 mm of precipitation, 3 h duration) occurred on 30 May 1996 (Fig. 3a and b), the second one (no. 56, 54.8 mm of precipitation in 10 h 30 min) on 16 June 1996 (Fig. 3c and d). The third and fourth storms occurred on 18 June 1996 (no. 58, 17.3 mm of precipitation in 6 h, Fig. 4a and b) and on 8 July 1996 (no. 78, 62.7 mm in 10 h Fig. 4c and d), respectively. The parts (a) and (c) of each figure depict the actually measured patterns of $P$ and *TF*, plus predicted *TF* according to models B ($S = 1.15$ mm; optimized $E_w = 2.80$ mm) and D ($S = 5.75$ mm; optimized $E_w = 1.1$ mm) given earlier. Although neither of the two scenarios gives a perfect fit to the measured data, it is clear that model B (original $S$) conforms better to reality than model D (increased $S$). As shown in parts (a) and (c) of Figs. 3 and 4, the *TF* predicted by model D reacts too slowly to changes in rainfall intensity, which leads to serious overestimation of *TF* after rainfall has ceased. Furthermore, as illustrated in parts (b) and (d) of Figs. 3 and 4, the amount of water stored on the canopy ($C$) predicted by model D sometimes attains unrealistically high values (up to 10 mm). Modelling efficiencies as calculated with the method developed by Nash and Sutcliffe (1970) (a value of 1 indicates a perfect fit) for model B were determined at 0.56 (storm 26), 0.50 (storm 56), 0.67 (storm 58) and 0.74 (storm 78) while those for model D were determined at 0.65 (storm 26), 0.45 (storm 56), 0.63 (storm 58) and 0.73 (storm 78).

### 5.4. Application of the analytical model

Measured and predicted totals of *TF* as determined by the analytical model are presented in Fig. 5. Initially the model was run both on a daily and on an event basis. Because the difference between the two respective predictions was very small, daily rainfall was used in the various simulations discussed below.

Model A employed a value of $S$ of 1.15 mm and an

Fig. 2. Measured and modelled cumulative throughfall for the Rutter model using four scenarios: (a) Model A, $S = 1.15$ mm, $E_w = 0.11$ mm $h^{-1}$; Model B, $S = 1.15$ mm, $E_w = 2.80$ mm $h^{-1}$; Model C, $S = 5.75$ mm, $E_w = 0.11$ mm $h^{-1}$; Model D, $S = 5.75$ mm, $E_w = 1.10$ mm $h^{-1}$.


(a)
measured (excluding big storm)
measured
predicted A (excluding big storm)
predicted A
Big storm
Cumulative TF [mm]
0
100
200
300
400
500
600
700
800
04/05
11/05
18/05
25/05
01/06
08/06
15/06
22/06
29/06
06/07
(b)
measured (excluding big storm)
predicted B
Cumulative TF [mm]
0
50
100
150
200
250
300
04/05
11/05
18/05
25/05
01/06
08/06
15/06
22/06
29/06
06/07
(c)
measured (excluding big storm)
predicted C
Cumulative TF [mm]
0
50
100
150
200
250
300
350
400
450
500
04/05
11/05
18/05
25/05
01/06
08/06
15/06
22/06
29/06
06/07
(d)
measured (excluding big storm)
predicted D
Cumulative TF [mm]
0
50
100
150
200
250
300
04/05
11/05
18/05
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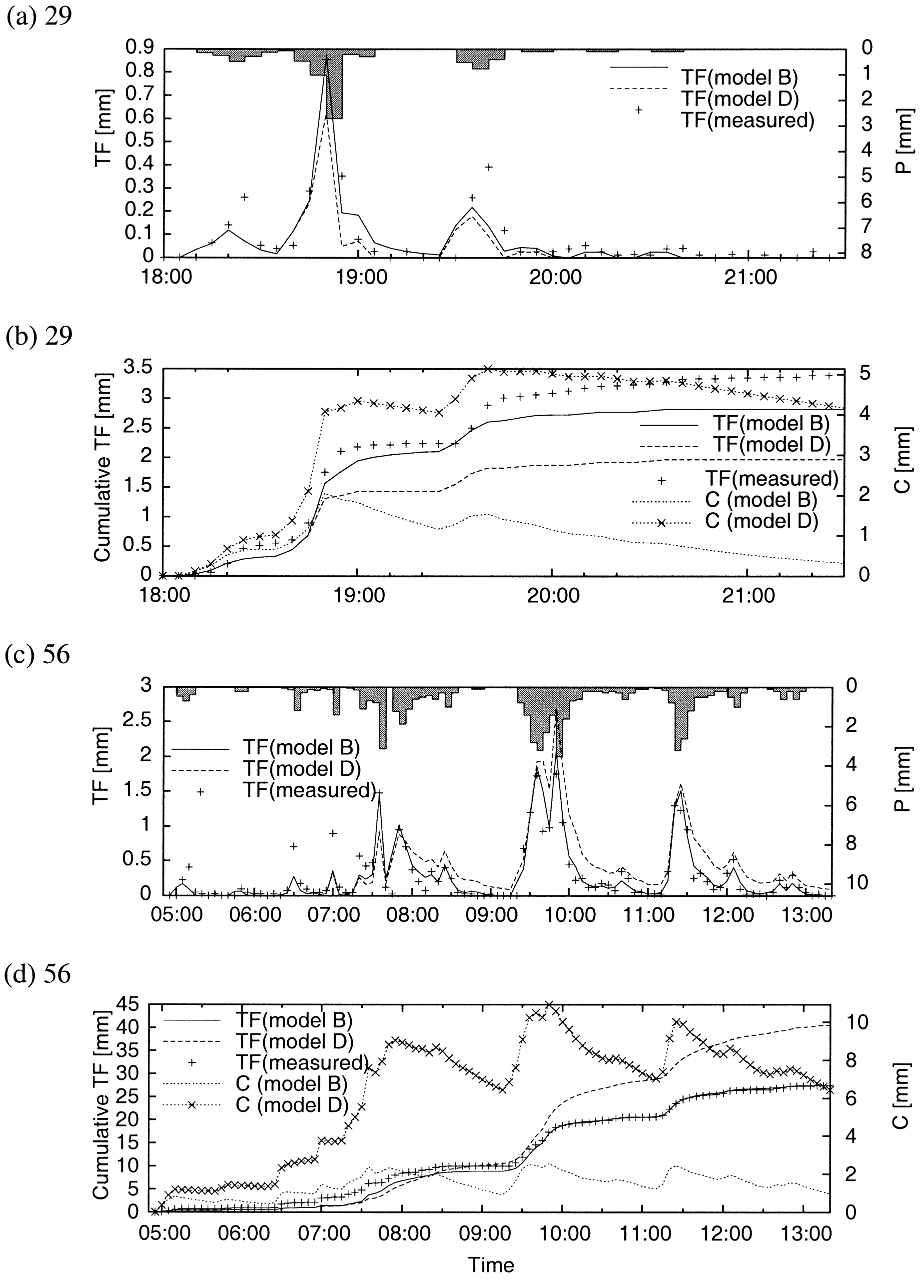


Fig. 3. Measured and modelled throughfall patterns predicted by the Rutter model for: ((a) and (b)) storm no. 29 (8.5 mm) and ((c) and (d)) storm no. 56 (54.8 mm). Also indicated in (b) and (d) are the predicted amounts of water stored on the canopy.

$\bar{E}_w/\bar{R}$ of 0.06 as based on the estimated evaporation using Eqs. (4) and (5) and recorded rainfall intensity ($1.85\ \mathrm{mm\ h^{-1}}$). As shown in Fig. 5, model A severely underestimated the measured interception loss (cf. Table 3). An optimum fit (model B, including the 13 May event) was obtained when using the original estimate of $S$ and an optimized value for $\bar{E}_w/\bar{R}$ of 0.51 as

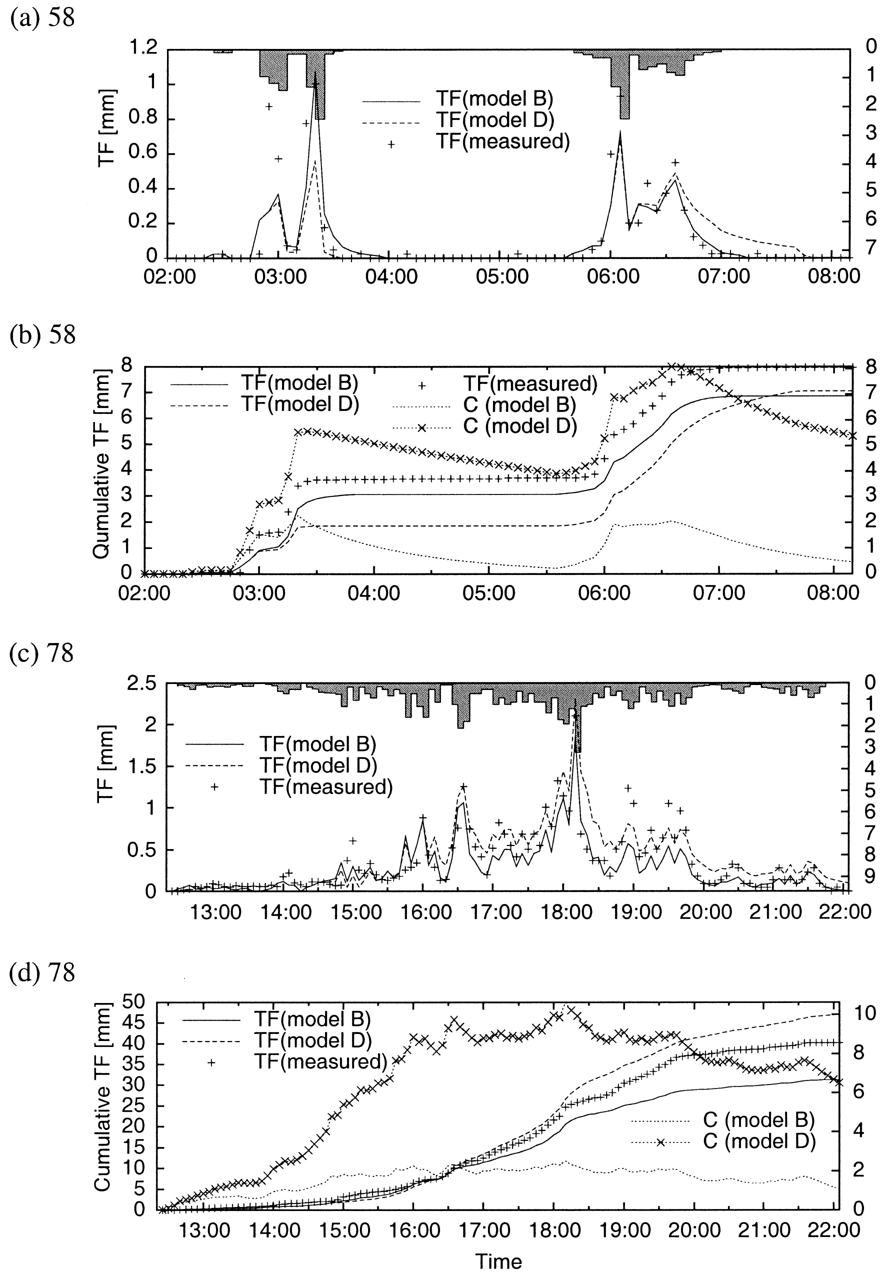


Fig. 4. Measured and modelled throughfall patterns by the Rutter model for: ((a) and (b)) storm no. 58 (17.3 mm) and ((c) and (d)) storm no. 78 (62.7 mm). Also indicated in (b) and (d) are the predicted amounts of water stored on the canopy.

derived from the regression of interception loss vs. gross precipitation (Gash, 1979, cf. Table 3). Increasing $S$ to 5.75 mm (model C) produced a somewhat better result than model A but to establish an optimum fit $\bar{E}_w/\bar{R}$ still had to be increased to 0.395 (model D).

The importance of $\bar{E}_w/\bar{R}$ to the outcome of the analytical model is further highlighted by the

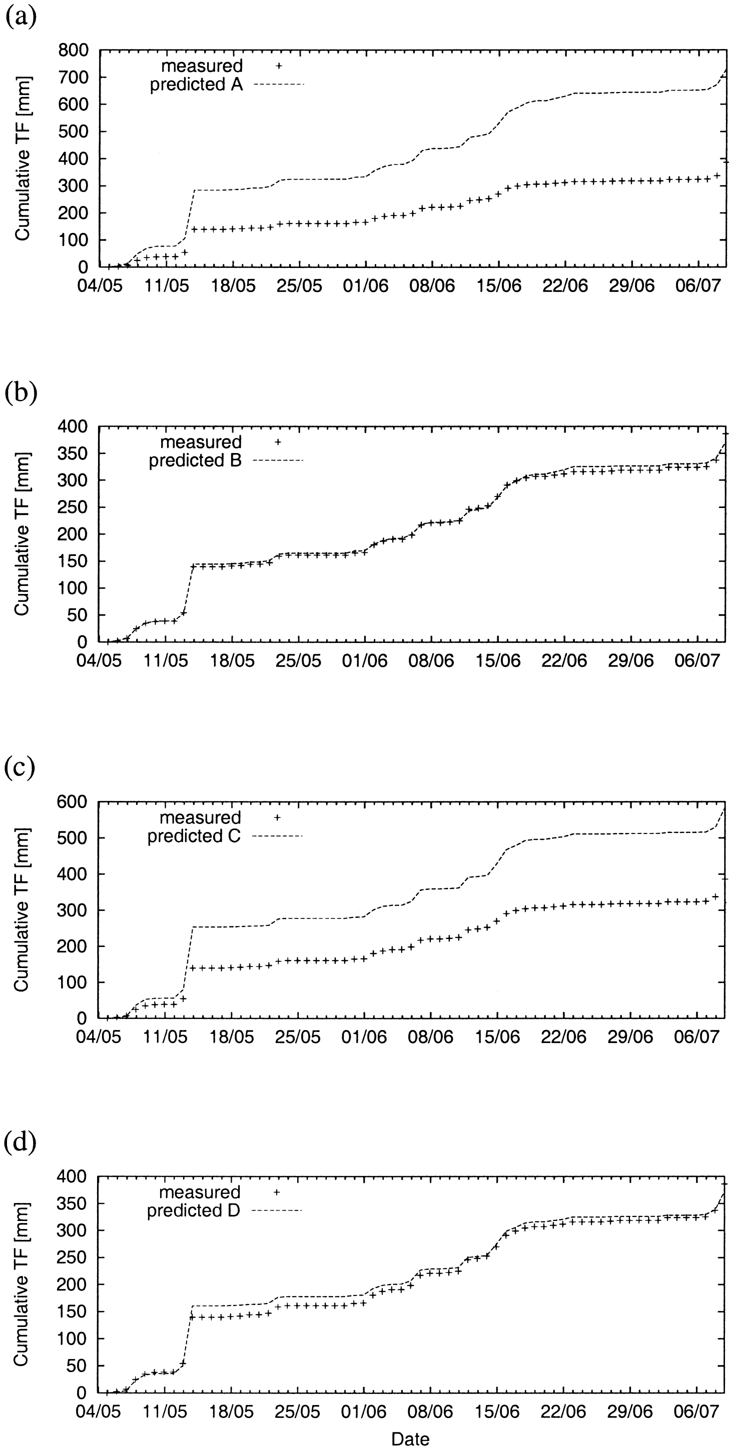


Fig. 5. Observed and predicted cumulative throughfall according to the analytical model for four scenarios: (a) model A, $S = 1.15$ mm, $\bar{E}_w/\bar{R} = 0.06$; (b) Model B, $S = 1.15$ mm, $\bar{E}_w/\bar{R} = 0.51$; (c) Model C, $S = 5.75$ mm, $\bar{E}_w/\bar{R} = 0.06$; (d) Model D, $S = 5.75$ mm, $\bar{E}_w/\bar{R} = 0.395$.

Table 3
Amounts of throughfall and components of interception loss for four scenarios used in an application of the Gash model. A: $S = 1.15$, $\bar{E}_w/\bar{R} = 0.06$; B: $S = 1.15$, $\bar{E}_w/\bar{R} = 0.51$; C: $S = 5.75$, $\bar{E}_w/\bar{R} = 0.06$, D: $S = 5.75$, $\bar{E}_w/\bar{R} = 0.395$. Model efficiency values according to the method of Nash and Sutcliffe (1970)

| Component of interception loss | Model | | | |
|---|---|---|---|---|
| | A (mm) | B (mm) | C (mm) | D (mm) |
| Total gross rainfall | 850 | 850 | 850 | 850 |
| Wetting-up of the canopy | 4.7 | 38.6 | 56.5 | 162.3 |
| Evaporation from a saturated canopy | 46.8 | 378.2 | 34.0 | 198.3 |
| Evaporation from canopy after rainfall ceases | 47.2 | 42.6 | 155.3 | 97.8 |
| Total calculated interception loss | 99 | 459 | 246 | 458 |
| Total calculated throughfall | 732 | 371 | 585 | 372 |
| Total measured throughfall | 386 | 386 | 386 | 386 |
| Model efficiency | 0.11 | 0.96 | 0.12 | 0.91 |

sensitivity analysis given in Table 4. A 10% change in the magnitude of $\bar{E}_w/\bar{R}$ produced an equally large change in predicted *TF*. As shown in Table 3, evaporation from a saturated canopy contributed the bulk (82%) of the total predicted interception loss in model B ($S = 1.15$ mm, optimized $\bar{E}_w/\bar{R}$), with near-equal contributions by evaporation during the wetting up and drying of the canopy (8 and 9%, respectively) and only minor contributions associated with small storms. Naturally, raising the value of $S$ to 5.75 mm greatly increased the predicted importance of the wetting and drying components (models C and D; Table 3). Table 3 also lists the respective model efficiencies as calculated using the Nash and Sutcliffe (1970) method indicating a slightly better performance for model B compared to model D.

Based on the $\bar{E}_w/\bar{R}$ value derived from the regression of $E_i$ on $P$ (0.51), an apparent average wet canopy evaporation rate of 0.94 mm h$^{-1}$ is obtained when inserting the median rainfall intensity of 1.85 mm h$^{-1}$ in the case of $S = 1.15$; the corresponding value of $E_w$ for $S = 5.75$ mm is 0.73 mm h$^{-1}$. As indicated earlier, $\bar{E}_w$ based on Eqs. (4) and (5) was only 0.11 mm h$^{-1}$.

## 6. Discussion

### 6.1. Forest structural parameters

The presently adopted value for the canopy capacity ($S$, 1.15 mm) compares favourably with earlier (pre-hurricane Hugo) estimates for the Bisley forest by Scatena (1990). His estimates ranged from 0.08 mm for open-canopy, palm-filled ravine sites to 1.3 mm for well-stocked ridges. Clements and Colon (1975) reported values between 0.76 and 1.27 mm for a similar forest at a slightly higher elevation (450 m) at nearby El Verde. Otherwise, the present value is somewhat higher than reported for a number of forests in Amazonia (0.74–1.03 mm; Lloyd and Marques-Filho, 1988; Elsenbeer et al., 1994; Jetten, 1996;

Table 4
Sensitivity analysis of various parameters in the 'model B' scenario of the analytical model showing the change in predicted *TF* after increasing or decreasing a parameter by 10%

| Parameter (initial value) | Resulting *TF* | | | | |
|---|---|---|---|---|---|
| | Original | +10% | −10% | % change + | % change − |
| $\bar{E}_w/\bar{R}$ (0.51) | 371.3 | 331.7 | 411.3 | − 10.7 | 10.8 |
| $S$ (1.15 mm) | 371.3 | 369.0 | 373.6 | − 0.6 | 0.6 |
| $p$ (0.23) | 371.3 | 372.5 | 370.1 | 0.3 | −0.3 |
| $p_t$ (0.023) | 371.3 | 369.5 | 373.1 | − 0.5 | 0.5 |

Ubarana, 1996) East Africa (0.89 mm; Jackson, 1975), or South-east Asia (ca. 0.9 mm; Murdiyarso, 1985; Dykes, 1997) which is in line with the higher interception loss at the study site (cf. Schellekens et al., submitted).

On the other hand, the value for the free *TF* coefficient ($p$, 0.23) is unexpectedly high considering the nature of the forest and the high $E_i$. Lloyd et al. (1988) and Ubarana (1996) reported $p$-values of 0.03–0.08 for Amazonian forest based on anascope readings. Elsenbeer et al. (1994) obtained a value of 0.32 for a forest in western Peru, whereas Jackson (1975) found a $p$ value of 0.24 for a tall montane forest of exceptional biomass in Tanzania (Lundgren, 1978). The latter two interception studies based their estimates of $p$ on an analysis of small storms. It is possible, therefore, that these estimates do not represent 'true' free throughfall but rather 'indirect' throughfall occurring before the canopy is fully saturated. The foliar structure of the Tabonuco forest which features pointed leaves with a waxy surface that are pointed downwards (cf. Ubarana, 1996) would certainly facilitate this. Such characteristics would also tend to speed up drainage from the canopy and so limit canopy water retention. This is also confirmed by our application of the Rutter model which clearly showed that the use of a large value of $S$ seriously overestimates *TF* after rainfall has ceased (cf. Fig. 3a and c and Fig. 4a and c). As such, the claims of very large storage capacities for tropical rain forests (Herwitz, 1985; Calder et al., 1986) are not supported by the present analysis. However, extra storage on and subsequent evaporation from trunks and branches during high intensity rain cannot be ruled out entirely, and further work on the stemflow dynamics at Bisley is required.

### 6.2. Performance of the Rutter and analytical models

Despite the fact that the Rutter model is more sophisticated than the analytical model, its performance was not better (cf. Figs. 2 and 5). In order to match measured and predicted *TF* totals, the (average) value of $E_w$ had to be increased to 2.8 mm h$^{-1}$. Nevertheless, a comparison of the detailed *TF* records with the model predictions for individual storms (Figs. 3 and 4) indicates that high evaporation is more likely than an increased storage capacity. In addition, when using a value of $S$ of 5.75 mm, the tendency of the Rutter model to overestimate *TF* for large storms and to underestimate *TF* for small storms remains. Also, the model failed completely to predict the *TF* total associated with the major event occurring on 13 May (Fig. 2). The main reason for this (and why the simpler analytical model does not have this problem; Fig. 5) lies in the variable character of the rainfall at Bisley. As shown in Figs. 3a and c and 4a and c, several intervals of relatively high rainfall and throughfall intensity may occur within a storm. In particular, the relation between 5-min *TF* and $P$ totals at Bisley is almost linear. This is where the Rutter model breaks down (because $E_w$ in the model does not keep up with rainfall intensity). This is less of a problem with the analytical model where variations in rainfall intensity are evened out by the use of $\bar{R}$ for entire storms. A similar problem was encountered by Jetten (1996) for a forest in Guyana. He therefore proposed a multi-layer version of the Rutter model (Cascade) whose main advantage is its ability to store larger quantities of water during large, high intensity events. The Cascade model was also applied to the present data. However, although it solved the problem of overestimation of *TF* for large events to some extent, the results were not satisfactory on a 5-min time scale. In particular, the *TF* pattern predicted by Cascade after rainfall had ceased, resembled that produced by the Rutter model with increased $S$ (cf. Figs. 3a and c and 4a and c). An alternative, stochastic, approach was advanced by Calder (1986) in which rainfall interception is described in terms of the stochastic manner in which individual elements of a canopy are struck and wetted by individual raindrops. An application of the stochastic model using the present data is in preparation.

The drainage function (Eq. (7)) is an important component of the Rutter model. As indicated in the previous section, it seems plausible that drainage at Bisley also occurs before $C = S$. A potential improvement of the Rutter model for application at the study site and possibly other rain forests with similar characteristics, therefore, would be to adjust the drainage function accordingly and to reduce the value of $P$ to the average gap fraction as estimated, for example, using an anascope (Gash and Morton 1978; Ubarana, 1996). However, the difficulties associated with the proper quantification of $E_w$ at Bisley (see below) tend to overshadow such refinements.

The analytical model performed very well on our data as long as the value of $\bar{E}_w/\bar{R}$ was derived from the regression of $E_i$ on $P$. By contrast, the use of the much smaller value of $\bar{E}_w/\bar{R}$ based on Eqs. (4) and (5) resulted in a highly overestimated amount of *TF* (Fig. 5 and Table 3). Good results with 'measured' $\bar{E}_w/\bar{R}$ were obtained by Lloyd et al. (1988) for a forest in Central Amazonia and by Ubarana (1996) using measured $E_w$ in the Rutter model in Western Amazonia, i.e. at mid-continental equatorial locations with low wind speeds. Under more maritime tropical conditions, regression-based estimates of $\bar{E}_w/\bar{R}$ tend to be (much) higher than those based on the Penman–Monteith equation (cf. Bruijnzeel and Wiersum, 1987; Dykes, 1997; Waterloo et al., 1999). However, caution is needed when comparing regression-based values of $\bar{E}_w/\bar{R}$ and Penman–Monteith based values of $\bar{E}_w$. The cited maritime tropical studies all used mean rather than median rainfall intensities to represent $\bar{R}$. In the present case, the use of an average rainfall intensity would yield an apparent $\bar{E}_w$ of 1.52 vs. 0.94 mm h$^{-1}$ for the median rainfall intensity. Just as important is the temporal resolution of the rainfall data used to estimate $\bar{R}$. For Bisley, rainfall intensity estimated from 5-min, hourly and event-based data amounted to: 7.2 (median 2.5), 2.8 (median 1.0) and 3.0 mm h$^{-1}$ (median 1.9 mm h$^{-1}$), respectively. Consequently, the variation in $\bar{E}_w$ derived from inserting the respective values of $\bar{R}$ in $\bar{E}_w/\bar{R}$ would be just as large.

### 6.3. Wet canopy evaporation

Both the Rutter and the analytical model could be adjusted in such a way as to predict observed *TF* totals at Bisley satisfactorily under most circumstances (Figs. 2 and 5, respectively). However, this required adjusting the rate of evaporation from the wet canopy $E_w$ as predicted by Eqs. (4) and (5) by at least one order of magnitude. Interestingly, the optimized value for $E_w$ was markedly different for the two models, with the largest value needed for the Rutter model. This is due to the problem of variable rainfall intensities within storms referred to already.

Schellekens et al. (submited) reported a total evapotranspiration (ET) for the forest at Bisley over the two years 1996 and 1997, that was approximately 42% higher than the evaporation equivalent of the total net radiant energy input ($R_n$). The contrast becomes even larger when considering evaporation from a wet canopy. The corresponding radiant input only amounts to 0.11 mm h$^{-1}$ (vs. optimized $E_w$ values of 1.1–2.8 mm h$^{-1}$), suggesting that during wet canopy conditions typically <10% of the required energy at Bisley is supplied in the form of radiant energy. A similar, though less extreme, case has been documented by Shuttleworth and Calder (1979). Here, the canopy of a spruce plantation at Plynlimon, Wales, UK, was shown to receive typically 20% of its total energy input during wet canopy conditions in the form of radiant energy, whereas the average annual ET from the forest exceeded the total radiant energy input by about 12% (Shuttleworth and Calder, 1979). High values of $E_w$ (0.5–0.8 mm h$^{-1}$; Bruijnzeel and Wiersum, 1987; Dykes, 1997; Waterloo et al., 1999) have also been inferred from rainfall and *TF* measurements at other maritime tropical locations but much lower values (typically about 0.2 mm h$^{-1}$ or less; Lloyd et al., 1988; Hutjes et al., 1990; Asdak et al., 1998; Waterloo et al., 1999) are usually calculated from above-canopy climatic measurements using Eq. (4). Similarly, Gash et al. (1980) derived an average value of 0.13 mm h$^{-1}$ for $E_w$ in the case of the Plynlimon spruce forest referred to earlier when using the climatic data, whereas a value of 0.24 mm h$^{-1}$ was inferred from measured interception. Another strong argument for the importance of non-radiant energy for evaporation at Bisley is the observation of Schellekens et al. (submitted) that *TF*/*P* ratios did not differ significantly for day-time and night-time events.

However, the origin of the energy needed in surplus of radiant energy is still largely a matter of speculation. Large-scale advection of warm air from the nearby Atlantic ocean is a likely source (cf. Shuttleworth and Calder, 1979), although, unlike the situation in Wales, wind speeds in the study area are rather low (typically 1.1–2.3 m s$^{-1}$; Holwerda, 1997; Schellekens et al., 1998). Another possible energy source is heat released upon condensation of water vapour in the air above the forest. The latter possibility seems to find support from the observation that larger rainfall events at Bisley are accompanied by higher $E_i$ (i.e. the *TF*/*P* ratio is independent of storm size; Schellekens et al., submitted) suggesting a positive feedback of rainfall amount (and thereby condensation) on the magnitude of $E_i$.

Accepting that the presently measured $E_w$ requires the total available energy $A$ to be much higher than measured $R_n$ also has consequences for the magnitude of $r_a$. Typical values of $r_a$ at Bisley as computed with Eq. (5) are about $20\ \mathrm{s\,m^{-1}}$ (Holwerda, 1997; Schellekens et al., 1998). Schellekens et al. (submitted) used an inverse application of Eq. (4) to estimate $r_a$ from measured interception losses (equating $A$ with $\lambda E_w$), and obtained an average value of $2.1\ \mathrm{s\,m^{-1}}$ (a very low value). Using a similar approach, Asdak et al. (1998) obtained an average value of $3.2\ \mathrm{s\,m^{-1}}$ for a forest in central Kalimantan, Indonesia whereas Calder (1977) derived an optimum value of $3.5\ \mathrm{s\,m^{-1}}$ for the Plynlimon forest vs. a theoretical value of $5.4\ \mathrm{s\,m^{-1}}$ when using Eq. (5). However, when taking into account the sensitivity of $r_a$ (and thus $E_w$) as determined with Eq. (5) to variations in the magnitude of $z_0$—particularly at high values of $d/h_v$ (Gash et al., 1980)—the latter parameters would have to be reduced to unrealistically low values to bring the outcome of Eq. (5) in agreement with $r_a$ as estimated from measured interception losses. Thus, the question remains how the $r_a$ at Bisley can be so low compared to values predicted by Eq. (5). Calder (1990) drew attention to the possibility of enhanced upward transport of evaporated moisture by gusts and eddies, even during neutral or stable conditions, although he added that the 'extent to which these gusts are associated with thermal- or humidity driven plumes, local topography or other factors is still unknown'. More recently, McNaughton and Laubach (1998) showed that such enhanced upward moisture transport increases with a relative increase of the standard deviation of the wind speed compared to the mean wind speed. This confirms the contention of Calder (1990) that gusts may indeed be important, and although further study of the instantaneous wind data at Bisley is needed, it is probable that gusts are relatively important because of the prevailing low mean wind speed. In addition, the aerodynamic roughness of the Bisley forest may be increased further by its irregularly shaped canopy as a result of hurricane damage in the past (Scatena et al., 1993). Further work is necessary to improve our understanding of the transport mechanism of evaporated moisture under conditions of low wind speed during rain events. Clearly, at Bisley such transport is much more efficient than suggested by Eq. (5).

## 7. Conclusions

The use of a wet canopy evaporation rate based on the Penman–Monteith equation can result in severely underestimated interception losses as predicted by the Rutter and the analytical models. A detailed analysis of the results obtained with the Rutter model for four storms of variable magnitude showed that the results were improved more by increasing the wet canopy evaporation component than by increasing the canopy storage capacity to fit measured *TF*. Using an optimized value for the wet canopy evaporation component, both models were able to predict measured *TF* amounts satisfactorily. However, the Rutter model encountered severe problems for a large event with variable rainfall intensity. The analytical model experienced no such problems because it uses an averaged rainfall rate. Our results support the idea that it is primarily a high rate of evaporation from a wet canopy that is responsible for the high interception losses observed under wet maritime tropical conditions.

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