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AMAZONIAN EVAPORATION

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RESUMO

Medições de evaporação da cobertura vegetal seca e perdas por intercepção obtidas durante um estudo de dois anos de evaporação na floresta tropical no centro da Amazônia são utilizados para calibrar um modelo micrometeorológico de evaporação. A evaporação total da floresta é calculada usando este modelo para um período de 25 meses de medições médias horárias das variáveis meteorológicas feitas acima da cobertura vegetal e medições regulares de tensão d'água no solo. A perda mensal por evaporação durante este período é comparada com as medidas de precipitação, radiação e taxas de evaporação potencial calculadas.

ABSTRACT

Measurements of dry canopy evaporation and interception loss obtained during a two year study of tropical rain forest evaporation in the central Amazon are used to calibrate a micrometeorological model of evaporation.
The total forest evaporation is calculated using this model for a 25 month period from hourly average measurements of meteorological variables made above the canopy and regular measurements of soil water tension. The monthly evaporation loss over this period is compared with measured precipitation, radiation input and calculated potential evaporation rates.

1. INTRODUCTION

A previous paper (Shuttleworth *et al.*, 1984a) reported measurements of evaporation and sensible heat loss made above the canopy of undisturbed rain forest at an experimental site near the city of Manaus in the central Amazon. Results were presented for a sample of eight dry days in September 1983. Parallel papers (Lloyd and Marques, 1988, Lloyd *et al.*, 1988) describe in detail the measurement and modelling of rainfall interception for a two year period (September 1983 to August 1985) at this site. These results were supplemented with an additional extensive set of measurements of dry canopy evaporation made during intensive experimental campaigns from June to September 1984 and March to August 1985, see Shuttleworth (1988). This paper reiterates and reviews these important results with a view to making them more readily available to the Brazilian research community.

The evaporation is here modelled with a version of the Rutter interception model (Rutter *et al.*, 1971; 1975) modified to include a calculation of the transpiration when the canopy is dry. Evaporation is calculated from the Penman-Monteith equation (Monteith, 1965). The aerodynamic resistance (r_a) and surface resistance (r_s) used in the dry canopy description are derived from measurements in dry conditions over the two year experimental period. The surface resistance exhibits an apparent dependence on measured soil water tension and this dependence is incorporated into the model. In addition, the wet canopy description requires values for a canopy storage parameter, S, and a direct throughfall parameter, p, whose determination is described elsewhere (Lloyd *et al.*, 1988).

An almost continuous set of hourly average meteorological measurements made with two automatic weather stations mounted on a tower above the canopy, and regular measurements of soil tension, made with two sets of soil tensiometers, provide the driving variables used as input to the model. These measurements were available for twenty five months, from September 1983 to September 1985, a period in which the precipitation measured at this site is representative of the 16 year average.

Sensitivity analyses were carried out on the parameters r_a, S and p used in the model, and these are described in the following discussion. They indicate that the *total* evaporation loss shows only limited sensitivity to their assigned values, although the separate wet canopy and dry canopy components in some cases exhibit a more significant but complementary sensitivity.

In discussion the average monthly total evaporation calculated from the model is compared with the precipitation input, the input of radiant energy to the forest, and with estimates of potential evaporation made with the Penman (Penman, 1948) and Priestley-Taylor equations (Priestley and Taylor, 1972).

2. EXPERIMENTAL SYSTEMS AND ANALYSIS

The experimental site, instrumentation and routine analysis procedures used in the collection of data reported are described in detail in previous and parallel publications, see Shuttleworth *et al.*, (1984a), Lloyd and Marques (1988), Lloyd *et al.*, (1988), Roberts *et al.*, (1988) and Moore and Fisch (1986). This section serves merely to provide a summary, and to draw attention to particular aspects.

(a) Flux Measurements

The measurements of evaporation, sensible heat flux and momentum transfer were made using the 'Hydra', the Institute of Hydrology's eddy correlation flux measuring instrument, in its Mark 1 form. Tests of this device made at this site and elsewhere suggest its performance is satisfactory above tall vegetation, see for example Shuttleworth *et al.*, (1984a) and Lloyd *et al.*, (1984). The operational and analysis procedures used are identical to those used previously (Shuttleworth *et al.*, 1984a), except that a new routine correction procedure was incorporated into the off-line analysis. This is described in detail by Moore (1986). It makes a first order adjustment for the flux loss which results from the finite size and separation of the component sensors and the finite frequency response of the on-line analysis. The magnitude of this correction is different for each flux measured and varies with atmospheric conditions, but is generally in the order of 3 to 8 per cent. Results previously reported for this site (Shuttleworth *et al.*, 1984a) have been adjusted by this correction, and are included in their modified form in the present analysis.

The systematic and random errors involved in the measured fluxes provided by this prototype Hydra instrument, which may in part respond to its operational environment, were estimated for this site by Shuttleworth *et al.*, (1984). Comparison was made between the integrated energy fluxes measured by the Hydra and the daily integral of measured net radiation on the one hand, and between Hydra measurements of Bowen ratio and above canopy temperature and humidity profiles on the other. These suggest a possible systematic error optimistically estimated as 5 per cent, and pessimistically estimated as 10 per cent, in each of the measured energy fluxes. To maintain internal consistency between the radiation and heat flux measurements, and thereby remove this uncertainty from the discussion given later, the individual, hourly-average energy fluxes measured by the Hydra were corrected to maintain an integrated energy balance with the integrated radiant energy input for all hours when data were simultaneously available. The net effect of making this correction was to increase the calculated total evaporation rates given later by 5 per cent.

In a recent publication, Gash (1986) describes a simples equation for estimating the effective fetch of micrometeorological measurements over crops of different roughness. Applying this here, assuming the surface roughness characteristics described in section 3(b), indicates that 90 per cent of the measured flux originates within 1800 m of the experimental tower.

(b) Routine Meteorological Measurements

Two automatic weather stations (Didcot Instruments) were used continuously to provide hourly average measurements of net and solar radiation, air temperature, (aspirated) wet bulb depression, wind speed and direction, and rainfall. At least one station was operational for 730 complete days between 1 September 1983 and 30 September 1984. For eighty per cent of the time both automatic weather stations were functional, and the average values from the two stations have been used. When one of the stations was not operational, data were taken from the other. Some data were lost during September 1984 (13 days lost), February 1985 (10 days lost), March 1985 (4 days lost), and April 1985 (4 days lost). In the course of the following analysis, model calculations of total evaporation and their comparison with potential evaporation, net radiation and rainfall for months in which such data loss occurred, are made assuming that the sub-sample of available data is representative of the remainder of the month. The monthly average values presented are obtained by scaling calculations for days when data were available to take account of the appropriate number of missing days. In consequence the results for these four months are less reliable than those at other times.

For an extended period, 26 March 1984 to 22 June 1984, the polythene domes on both the net radiometers initially clouded and later cracked through prolonged exposure to solar radiation. The net radiation measurement over this period is not considered reliable, and it was substituted by an estimate of net radiation based on data taken through the remainder of this study as follows:

$$R_N = \{0.782 - 0.028 \ (2\pi \ [N_D - 31]/365)\} S_T - $$
$$- 16 + 7 \cos (2\pi \ [N_D - 31]/365) \qquad (1)$$

where R_N is the net radiation, S_T is the measured solar radiation, and N_D is the day number in the year.

Equation (1) is merely an empirical description of the relationship between net and solar radiation as measured by these automatic weather stations at this site, and its derivation is described by Lloyd *et al.*, (1988).

(c) Rainfall Interception Measurements

The measurement and modelling of rainfall interception by the forest canopy at this experimental site is described in detail elsewhere (Lloyd and Marques, 1988; Lloyd *et al.*, 1988).

In outline, the interception loss is deduced as the difference between rainfall measured above the canopy and throughfall measured with initially 16 and, after July 1984, 36 collection gauges, which were randomly relocated on a 100 m by 5 m grid at weekly, and occasionally daily, intervals. The stemflow component, which was small (in the order 2 per cent of rainfall) was measured with initially 6 and, after 10 August 1984, 19 stemflow gauges.

A detailed analysis of errors (Lloyd and Marques, 1988) indicates that the sampling error involved in the measurement of throughfall, and therefore (by difference) interception loss, is considerably greater than that commonly occurring in studies of temperate forest stands. This is a consequence of the horizontal movement of water in the forest canopy with some concentration of throughfall in widely spaced, and therefore only occasionally sampled, 'drip points', and depletion elsewhere. The presence of such large sampling errors means that comparisons between experimental measurements and model estimates of the interception component are only useful in the form of cumulative totals, with many gauges frequently and randomly moved over a long period (see Lloyd and Marques, 1988). Expressed in this form, the quasi-random sampling errors cancel over many samples. The present experiment defined the integrated fractional interception loss over the whole study period as 9.1 ± 3.5 per cent of gross rainfall.

(d) Soil Tension Measurements

The experimental system and resulting observations of soil water tension in this study were measured with tensiometers of conventional design. Two soil tension profiles were measured, 4 m apart, with tensiometers installed in the soil at 10 cm intervals to a depth of 50 cm, and then at 20 cm intervals to a depth of 1.9 m. In general most of the change in total soil water potential was observed to occur in the top 1 m of soil, and occurred most rapidly in the first 30 cm. The soil water matrix potential averaged over the 7 measurements above a 1 m depth, and over both tensiometer profiles, was chosen as an index of soil water status for use in the analysis and modelling described here.

The tension profiles were usually read at weekly intervals, but were read less frequently during 1983/4 wet season (December 1983 to March 1984), and daily during some of the drier months (September 1983, and July and August 1984). To facilitate the present analysis (and provide a regular input to the evaporation model) daily estimates of the soil water tension index were made by linear interpolation between days on which soil tension readings were made.

3. MODEL SPECIFICATION AND EXPERIMENTAL CALIBRATION

Although the measurements of interception loss were made continuously over a two year period and, apart from occasional experimental malfunctions, data are available for this period, the sampling errors on the throughfall measurement are such that even medium term averages (e.g. monthly values) are unlikely to provide useful estimates of the loss. At the same time, the measurements of dry canopy evaporation made in the course of this study, although extensive, are of limited duration, tend to be biased towards drier months of the year, and, even during intensive experimental campaigns, are only available part of the time. The objective of the present study is to provide monthly values for total forest evaporation for the whole 25 month study period. The philosophy adopted here is therefore to define a physically based, micrometeorological model with parameters specified from the available experimental data, and then to use this to provide monthly values of total evaporation loss.

The required model is descriptive rather than predictive, and, although necessarily providing a plausible, physical description of the forest, it has minimum complexity. Its purpose is:

(i) to synthesize values for short term interception loss (since the experimental results are useful only as long term cumulative totals) and

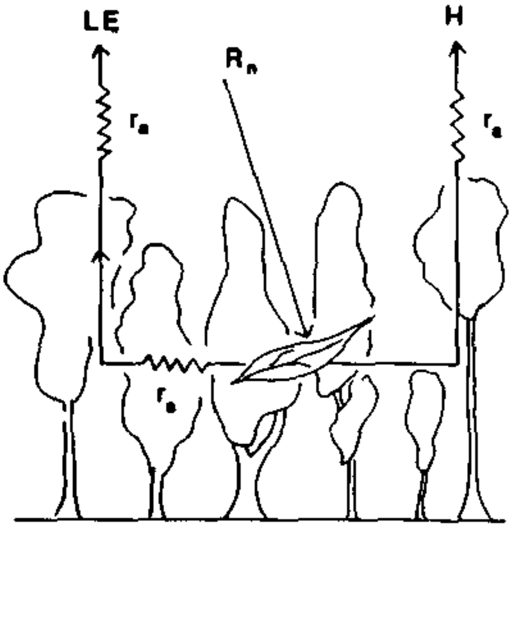
(ii) to interpolate the dry canopy results between periods of direct measurement, and (by including an explicit dependence on measured soil water tension) to compensate for any bias in the sample data towards drier portions of the year.

The description adopted is a combined Penman-Monteith/Rutter model, which has been successfully used in the past to describe evaporation from temperate forests (e.g. Calder, 1977) and is represented diagrammatically in Figure 1. It comprises a 'single source' description of evaporation, see Shuttleworth (1976). In dry canopy conditions the stomatal control of the forest is represented by a bulk surface resistance, r_s. Transfer to a 'screen height' above the forest is assumed to be controlled by an aerodynamic resistance, r_a. In the present case the screen height is coincident with the location of the automatic weather station. The timing of wet canopy evaporation, when $r_s = 0$, is given by the Rutter canopy water balance model which has been extensively described in the literature (e.g. Rutter *et al.*, 1971, 1974; Gash and Morton, 1978; Calder, 1977). The application of this model in this experiment is described in more detail elsewhere (Lloyd *et al.*, 1988), and this description is not repeated here. It is however appropriate in the course of the next section to make some general observations.

(a) Wet Canopy Model Validation

Rainfall at this wet, tropical site is largely convective, and generally falls as short, heavy storms. Interception loss in the form of evaporation from water stored on the forest canopy after a storm is important. The amount of water which can be stored on the forest canopy, the canopy storage, S = 0,74 mm, (Lloyd *et al.*, 1988) is therefore an important parameter. Less important is the fraction of rain which falls directly through gaps in the canopy, the free throughfall parameter, p = 0.08 (Lloyd *et al.*, 1988), since this primarily affects the rate at which the canopy store is filled at the beginning of the storm. Rainfall intensity is such that the canopy generally 'wets up' very quickly. The aerodynamic transfer resistance, r_a, affects the rate of evaporation in both dry and wet conditions. In a later section of this paper we investigate the sensitivity of the model to the parameters S, p and r_a in terms of its ability to describe both the total evaporation loss, and the interception component of that loss.

Figure 2 provides a comparison between the experimentally measured cumulative interception loss and that computed by the Rutter model, as described by Lloyd *et al.*, (1988). The experimental results are presented as the standard error band around the measured cumulative loss. This error increases in absolute terms but decreases in relative terms during the course of the experiment. The cumulative above canopy precipitation is also plotted in



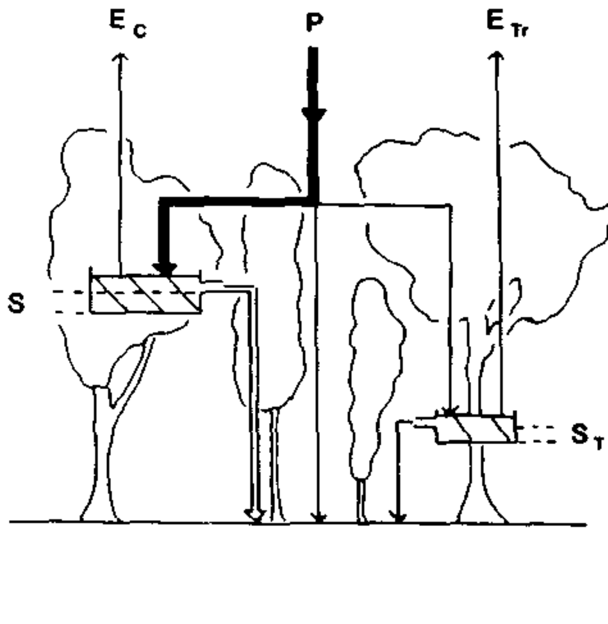


Figure 1 — Elements of the micrometeorological model used to describe and interpolate the experimental data.
(a) The single effective source 'Penman-Monteith' resistance framework. r_a is the aerodynamic transfer resistance, r_s is the surface resistance which is equal to zero for a totally wet canopy, and calibrated by applying the model in reverse in dry canopy conditions (see text).
(b) The 'Rutter' canopy water storage model. The precipitation is routed as direct throughfall, or into stores S_C and S_{Tr} on the canopy and trunks respectively. Evaporation and drainage rates are related to the proportional fill of these stores (see Rutter *et al.*, 1971, 1975).

Figure 2 as a guide perspective. The description provided by the Rutter model is consistent within experimental error with measured interception. In the absence of any additional evidence, it is sufficient to regard the fact that the model estimate of interception loss is only just consistent with measurement at the end of the experiment as an accident of statistics.

(b) Aerodynamic Resistance

A previous paper (Shuttleworth *et al.*, 1984a) presented and discussed the experimentally observed relationship between measured friction velocity and windspeed at this site for a sample of data collected during September 1983. This was presented as a working description of aerodynamic resistance for use in data interpretation and model application. The reader is referred to that work for further details. Here we update and supplement the observational data by including additional results drawn from studies made from July to September 1984 and from March to August 1985.

The data have again been heavily selected to include only daytime conditions with the Hydra well exposed for momentum flux measurement. When the friction velocity, u_*, was less than 0,1 m/s and wind speed, u, less than 0.5 m/s the results were prone to the effects of anemometer stalling, and were rejected. The remaining sample of 1004 hourly measurements are presented in Figure 3. In this figure the results are sorted into 0.5 m/s groups, and are presented together with their standard error. Also shown in the figure is a line of linear regression which has been constrained to pass through the origin and has a gradient of 0.185. The average value of the stability length, L, for all the data presented in Figure 3 is -294 m. Making a first

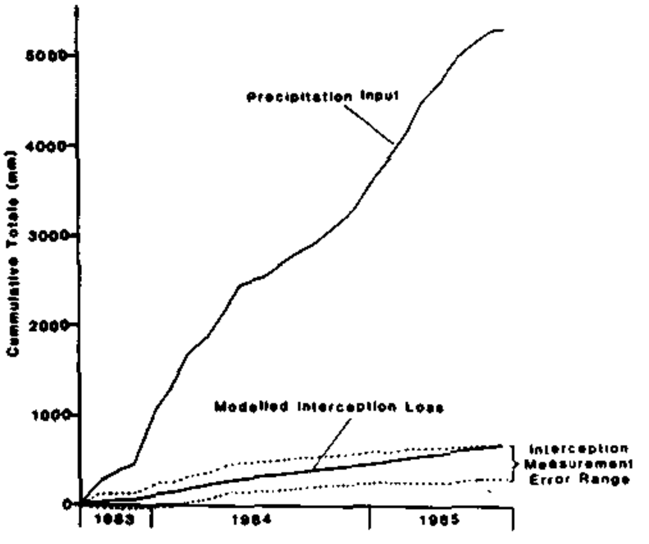


Figure 2 – Cumulative values for precipitation and model calculated interception loss for the experimental period over which interception measurements were made. Also shown is the range within which cumulative interception lies within one standard error of the experimentally measured interception loss.

order correction for the effect of stability, see Shuttleworth (1988), gives a value for r_a in neutral condition with the form:

$$r_a = \frac{f}{u} \qquad (2)$$

where f = 34.2.

This is the expression used to describe aerodynamic resistance in the remainder of this paper. A later section investigates the sensitivity of model calculations to the value of the constant, f, in Equation (2).

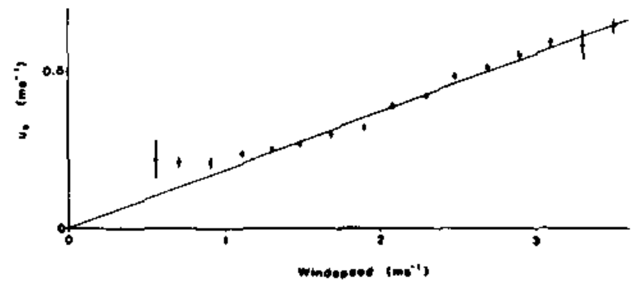


Figure 3 – Mean value and standard error of values of u* measured by the Hydra for selected hours, collected into groups and plotted against the windspeed measured with the AWS. Also shown is the line of linear regression constrained to pass through the origin.

(c) Surface Resistance

For each hour during which measurements of the evaporation flux, λE, and sensible heat flux, H, were available, the effective value of surface resistance was computed using the Penman-Monteith equation arranged as:

$$r_s = \left(\frac{\lambda}{c_p}\,\Delta\,\frac{H}{\lambda E} - 1\right) r_a + \rho\lambda\,\frac{D}{\lambda E} \qquad (3)$$

In this equation Δ is the rate of change of saturated specific humidity with temperature, λ is the latent heat of vaporization of water, c_p is the specific heat of air at constant pressure and ρ is the density of air. The specific humidity deficit, D, is calculated from the wet and dry bulb temperatures measured by the automatic weather station, and r_a is calculated from Equation (2). In practice the surface resistance is more conveniently represented and statistically analysed as its reciprocal, the surface conductance c_s.

Almost 1500 individual values of surface conductance were calculated for daylight hours in dry canopy conditions: measurements one hour prior to, during, and three hours after a rain storm were rejected to enforce this last condition. The results indicate a significant diurnal trend in the derived surface conductance which is shown in Figure 4. In this figure the values shown are the hourly mean and standard error of all the measurements available in the two year study period. Also shown is the simple quadratic curve

$$C_S^D(t) = 12.17 - 0.531\ (t-12) - 0.223\ (t-12)^2$$

$$(\mathrm{mm\ s^{-1}}) \qquad (4)$$

where t is the local time of day in hours (midnight equals zero). This equation provides an adequate description of the mean diurnal trend (correlation coefficient = 0.97).

A diurnal variation in surface conductance of the type illustrated in Figure (4) and described by Equation (4) would result if the stomatal conductance of forest trees, of which c_S is a measure, exhibited a significant dependence on increasing leaf water potential through the day. A parallel paper (Roberts *et al.*, 1988) describes the behaviour of stomatal conductance and its response to environmental variables in greater detail. The modelling analysis which follows is specifically orientated towards producing medium term (monthly) evaporation. We choose to regard the short term day to day variability in surface conductance as a combination of random experimental error, and a quasi-random physiological response to transient weather conditions. We further assume that these are adequately averaged at the monthly time scale.

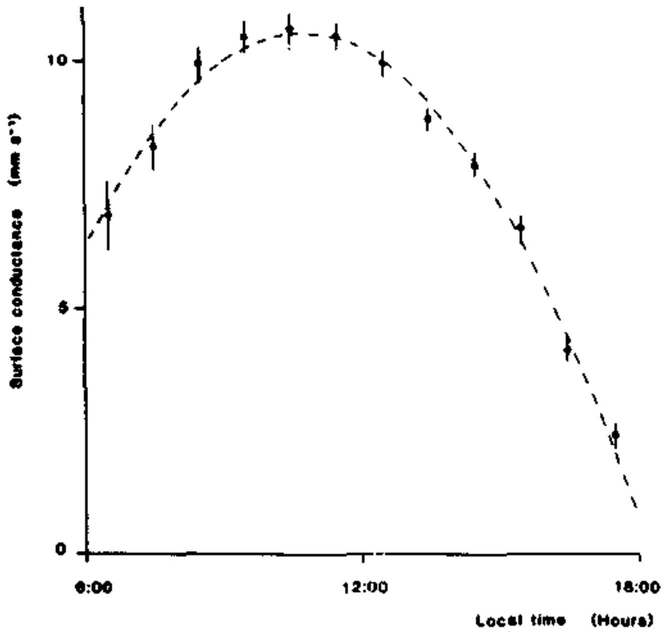


Figure 4 — Mean value and standard error of surface conductance, C_s^D for each daylight hour measured in dry canopy conditions over the whole 25 month experimental period. Also shown is the quadratic curve which adequately describes this diurnal trend, and which is described in the text.

More important in the context of a study of annual trends in evaporation is any systematic, longer term variation in stomatal conductance in response to seasonal changes, particularly those in soil water status. To investigate this, the mean surface conductance was expressed in the form

$$c_S = C_s^T\ (\phi)\ C_s^D(t) \qquad (5)$$

where ϕ is the daily value of the average soil water matrix tension to a depth of one metre obtained by interpolation between regular measurements. Individual values of C_s^T were computed from measured values of c_S using Equations (4) and (5) for each hour of data, and the resulting average values of C_s^T against the average soil water matrix tension, ϕ. Figure 5 illustrates the observed dependence and indicates that for the range of soil moisture tensions which prevailed in the course of the present experiment, and for which a descriptive model is required, the function C_s^T is represented within experimental error by a simple linear relation of the form:

$$C_s^T = 1.16 + 0.0047\ \phi \qquad (6)$$

in which the units used are kPa (Note: the intercept in Equation (6) is not zero because the sample measurements of c_S which were used to give the mean diurnal trend, Equation (4), correspond to an average soil water matrix tension of 34 kPa. The 'saturated soil' extrapolation of the mean diurnal trend is given by multiplying each of the coefficients in Equation (4) by the factor 1.16).

It is important to emphasize that the *apparent* dependence of surface conductance on soil water tension given in Equation (6) is merely an expedient and empirical mechanism to introduce seasonal dependence into the descriptive model used here to interpolate experimental data. In practice this equation is presumably also providing an indirect and first order description of other seasonal

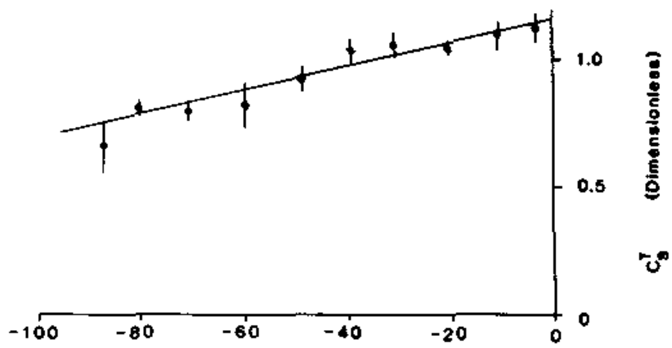


Figure 5 — Mean value and standard error of the function C_s^T for values of the average soil water matrix potential to a depth of 1 m expressed in kPa. Also shown is the line of linear regression described in the text.

component evaporation presented in Table 1 are to the assumptions made in the model used in their calculation.

In dry canopy conditions with forest vegetation there is a strong relationship between transpiration and surface conductance. A systematic error in $C_s^D(t)$, the function used to describe the diurnal variation in surface conductance, will propagate strongly into the calculated transpiration flux. However, the model used to calculate evaporation from a dry canopy is the exact reverse of that used to synthesize the values of surface conductance during calibration. In this sense the calculated transpiration flux is model independent. Any systematic error in this flux calculation would merely reflect a systematic error in the original evaporation measurements used to calibrate the model.

This situation is complicated, however, by the observed relationship between surface conductance and soil water tension. The explicit inclusion of this dependence in the model is designed to compensate for any seasonal bias in the original dry canopy measurements, and to provide a description of the apparent seasonal response to soil water status. In practice it also implicitly incorporates the response to other (unmeasured) seasonal changes such as those in leaf area index, and it is of interest to consider the effect on the annual cycle of evaporation of omitting this feature from the model. Figure 8 shows the fractional change in (a) calculated transpiration and (b) total evaporation caused by such omission. The fact that calibration measurements were biased towards drier portions of the year, when surface conductance is in general lower, means that the average daily trend derived from them leads to an underestimation in transpiration (of the order 6 per cent) in wet portions of the year. The fractional error in total evaporation is less than this (around 2,5 per cent) because this also includes interception loss, which is not affected and is indeed proportionately enhanced at this time of year. On the other hand, transpiration is over-estimated by up to 10 per cent in very dry months with the omission of soil tension dependence, and this largely propogates into total evaporation loss.

TABLE 1

Monthly totals (in mm) of measured rainfall, the water equivalent of net radiation, and (model calculated) total evaporation and interception loss. Values marked † correspond to months when data is extrapolated from a restricted number of measurement days as described in the text.

Time Period	Rainfall	Net Radiation	Total Evaporation	Interception Loss
Sep 1983	195	132	126	24
Oct	197	111	103	23
Nov	73	107	105	13
Dec	484	92	91	35
Jan 1984	325	117	109	36
Feb	386	98	98	50
Mar	218	124	119	42
Apr	257	114	108	34
May	326	125	117	35
Jun	66	133	117	21
Jul	144	138	129	26
Aug	130	133	122	22
Sep †	120	142	130	30
Oct	162	139	124	23
Nov	147	132	116	17
Dec	312	119	104	27
Jan 1985	286	123	104	27
Feb †	279	112	87	29
Mar †	340	116	94	32
Apr †	357	136	116	28
May	261	115	100	32
Jun	155	116	101	25
Jul	130	123	109	28
Aug	90	129	103	16
Sep	52	144	116	8
Total	5492	3070	2748	683

features such as variation in leaf area index. Its effect on calculated evaporation is limited, see Section 4(b).

Equation (5), with C_s^D given by Equations (4) and (6), is the formula used to describe surface conductance during daylight hours in the model calculations of dry canopy evaporation rate which follow. At night the surface conductance is set to an arbitrary low value (0.1 mm/s).

4. CALCULATED EVAPORATION RATES

In the following section the calibrated micrometeorological model just described is used to calculate the total evaporation loss, and its dry canopy and wet canopy components. The sensitivity of these calculations to the values of the forest structure parameters is investigated. Monthly evaporation is compared with precipitation and radiation input, and with potential evaporation estimates. However, to place the 25 month study period in historical perspective, Figure 6 illustrates the annual cycle in above canopy precipitation measured at the experimental site for each month from September 1983 to September 1985. This is compared with the mean and standard deviation of monthly precipitation derived from measurements made over sixteen years, from 1965 to 1980 at a climatological station in a forest clearing 2 km from the experimental site. The rainfall observed in the course of the present study is clearly representative of the annual cycle over this longer time scale: only one month, December 1983, has rainfall which is more than two standard deviations from the monthly mean.

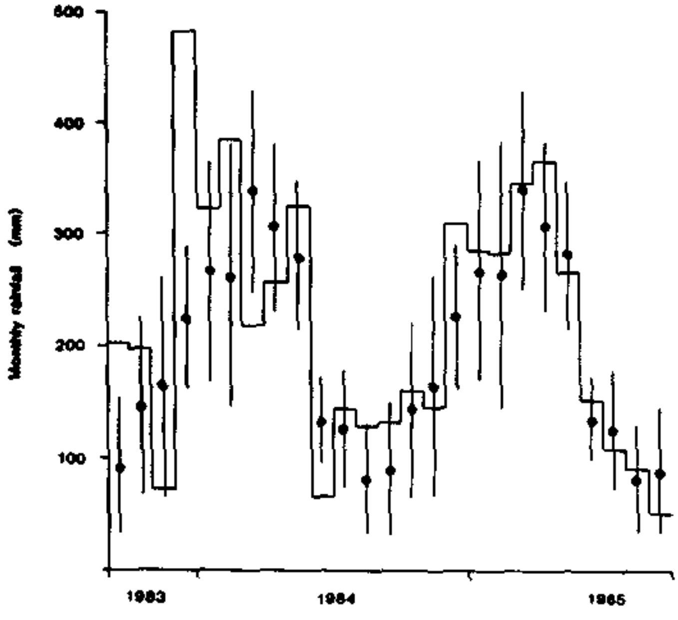


Figure 6 — Monthly precipitation in mm measured at the experimental site for the 25 month experimental period. Also shown as points and error bars is the mean and standard deviation of monthly precipitation measured at a nearby climatological station for the years 1965-1980.

(a) Monthly Values

Table 1 presents the measured monthly values of precipitation and net radiation input (expressed as a water equivalent), together with the total evaporation and interception provided by the model. Figure 7 illustrates the monthly precipitation, calculated total actual evaporation loss and interception component of this loss from September 1983 to September 1985. The values may be prone to systematic error estimated as in the order 5 to 10 per cent, but calculated values for months with significant AWS data loss are considered significantly less reliable than this.

Precipitation exceeds evaporation for 20 of the 25 months studied, and over the whole period 50 per cent of the rainfall falling on the site returned to the atmosphere as evaporation. In particularly dry months evaporation exceeded precipitation by about a factor of two. Total evaporation is reasonably constant throughout the year (110 mm per month), with deviations of up to 20 per cent around this value. The model calculates that the interception process contributed approximately 25 per cent of the total evaporation loss on average, but this changed from over 50 per cent in very wet months to less than 10 per cent in very dry months. Although the *absolute* interception loss (around 30 mm per month) is greatest in wet months, the *fraction* of precipitation lost by direct interception in the forest canopy is less in the wet portions of the year. This fraction, calculated by the model as 12.4 per cent on average (and measured as 9.1 ± 3.5 per cent), changes from as little as 8 per cent in Aprill 1985 to as much as 32 per cent in June 1984.

(b) Model Sensitivity Studies

Before proceeding to further discussion it is convenient to investigate how sensitive the values of total and

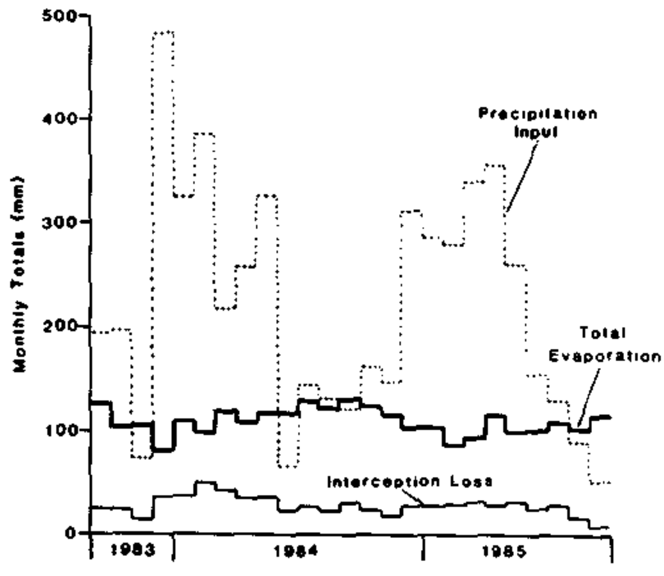


Figure 7 — Monthly values for precipitation, total evaporation and the interception component for the 25 month experimental period.

The values of the canopy storage, S, the free throughfall coefficient, p, and the constant, f, used in Equation (2) all affect the calculated total evaporation loss, and the interception and transpiration components. A change in any of these parameters will affect the calculated interception loss, but such changes also have a reduced and, in general, opposite effect on the calculated transpiration. For example, if the value of S is raised, the canopy is assumed to store a greater amount of water after each storm. The calculated interception loss increases because of this, but in compensation the modelled canopy also takes longer to dry out after the storm. The calculated transpiration, which is in part suppressed by the presence of free water on the canopy, is reduced. Compensation of this type also applies for changes in the parameter p.

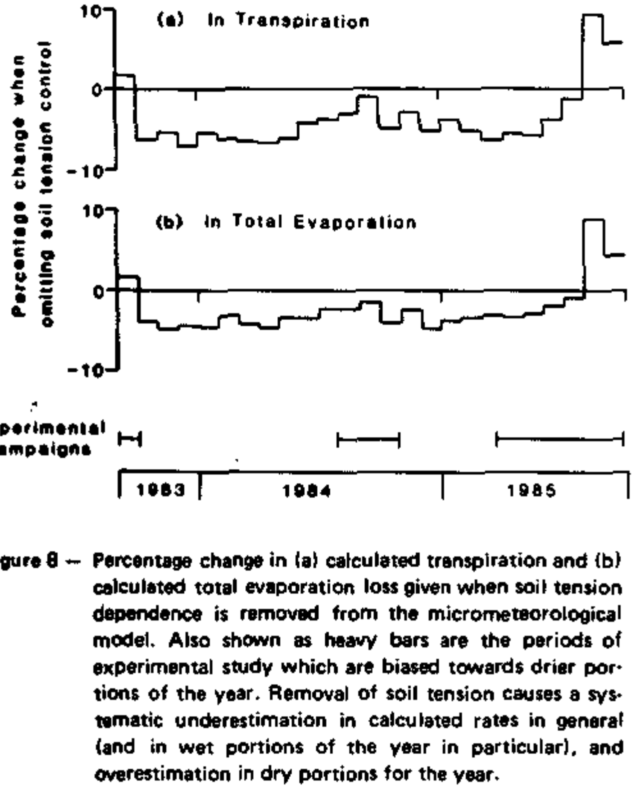


Figure 8 — Percentage change in (a) calculated transpiration and (b) calculated total evaporation loss given when soil tension dependence is removed from the micrometeorological model. Also shown as heavy bars are the periods of experimental study which are biased towards drier portions of the year. Removal of soil tension causes a systematic underestimation in calculated rates in general (and in wet portions of the year in particular), and overestimation in dry portions for the year.

Tables 2(a) and 2(b) give the percentage change in calculated interception and total evaporation when the values of S, p and f are each individually and arbitrarily altered to be 50 per cent less and 50 per cent greater than their preferred values for nine sample months during the experimental period. On averaged calculated interception is altered by 23 per cent and total evaporation by 3 per cent in response to a 50 per cent change in the value of S, but neither shows any significant sensitivity to the value of p. Changes of minus 50 and plus 50 per cent in the value of f generate average complementary changes of plus 19 and minus 8 per cent in calculated interception loss, and plus 3 and minus 2 per cent in total evaporation.

In summary, model calculations of *total* forest evaporation given in Table 1 seem stable at about the five per cent level to rather extreme changes, i.e. omitting the effect of soil water tension from the dry canopy calibration, and changing the more important parameters in the wet canopy description by 50 per cent.

(c) Comparison with Standard Rates and Radiation

The concept of potential evaporation, λE_p, (Penman, 1948) has provided a robust framework for describing medium term, average evaporation in many hydrological and meteorological applications. Moreover, it has generally been found to provide a description of actual evaporation for well-watered, short agricultural crops, which is adequate at the 5 to 10 per cent level for averaging periods in the region of 10 to 30 days. Priestley and Taylor (1972) suggested a simplified definition, with potential evaporation, λE_{PT}, expressed as an easily calculable function of temperature and radiation. For the explicit forms used to calculate λE_p and λE_{PT} see Shuttleworth *et al.*, (1984a).

In light of the considerable literature indexing actual evaporation to calculated potential rates for a broad range of crops and a wide range of soil conditions, it is a constructive exercise to carry out a similar procedure for the monthly total evaporation rates presented in Table 1. In this way the present analysis may provide an enlightened basis for future speculation regarding the consequences of any proposed land use change in the Amazon region.

Figure 9 provides the desired comparison, with estimates of potential rate made with the Penman equation in Figure 9(a), and with the Priestley-Taylor equation in Figure 9(b). In general the agreement between the model-calculated evaporation rate and estimated potential evaporation rate given by the Penman equation is good within estimated experimental errors at about the 5 to 10 per cent level throughout most of the experimental period. The most obvious discrepancies occur during the two wettest months, December 1984 and February 1984, when the Penman equation underestimates evaporation and during September 1985, the driest months towards the end of the second dry season, when the Penman equation overestimates evaporation. It is fairly clear that enhanced evaporation by interception loss is responsible for the first discrepancy. A higher proportion of dry days with evaporation less than potential rate (see Shuttleworth *et al.*, 1984a), exacerbated by a response to increasing soil water tension and a possible reduction in leaf area, is responsible for the second. These two extremes arguably highlight a more general relationship between monthly rainfall and the monthly average ratio of (E/E_p). A linear regression against monthly rainfall P_m, has the form:

$$\frac{E}{E_p} = 0.97 + 0.00035\,P_m$$

TABLE 2

Percentage change in model calculated (a) interception loss and (b) total evaporation in response to changes of −50 per cent and +50 per cent in the preferred values of S, p and f which are defined in the text.

PERCENTAGE CHANGE IN CALCULATED INTERCEPTION

(a) Sample Monthly Period	S Changed by		p Changed by		f Changed by	
	−50%	+50%	−50%	+50%	−50%	+50%
Sep 1983	−27.1	25.0	0.8	−1.3	19.6	−10.0
Dec	−24.0	20.3	0.9	−0.8	23.1	−11.4
Mar 1984	−20.5	18.8	1.0	−1.1	20.5	−8.8
Jun	−23.3	22.4	1.4	−1.4	14.3	−6.2
Sep	−22.0	24.7	1.0	−1.0	22.3	−8.7
Dec	−28.9	24.4	0.8	−0.8	14.8	−7.8
Mar 1985	−22.5	20.6	0.6	−0.9	10.0	−6.6
Jun	−19.6	22.4	1.2	−1.1	18.0	−4.4
Sep	−17.5	23.4	0.5	−1.3	25.0	−11.3

PERCENTAGE CHANGE IN TOTAL EVAPORATION

(b) Sample Monthly Period	S Changed by		p Changed by		f Changed by	
	−50%	+50%	−50%	+50%	−50%	+50%
Sep 1983	−3.3	2.8	0.1	−0.2	2.8	−1.6
Dec	−5.1	5.1	0.2	−0.1	9.8	−4.1
Mar 1984	−4.5	4.0	0.2	−0.3	7.4	−3.3
Jun	−2.5	2.1	0.1	−0.2	1.5	−0.9
Sep	−3.8	2.9	0.2	−0.2	2.5	−1.6
Dec	−3.8	3.0	0.1	−0.2	2.7	−1.5
Mar 1985	−3.4	3.7	0.1	−0.1	3.2	−0.4
Jun	−2.4	3.1	0.2	−0.2	2.8	−0.4
Sep	−1.2	0.4	0.0	−0.1	−3.1	−1.2

with a correlation coefficient of 0.561. Over the whole 25 month period the total evaporation estimated by the Penman equation is 2637 mm, 4.1 per cent less than that calculated by the calibrated micrometeorological model.

The level of agreement in the comparison between monthly evaporation and potential evaporation estimated by the Priestley-Taylor equation is equally satisfactory. The total estimated potential evaporation over the 25 month study period is 2868 mm, 4.4 per cent bigger than that given by the model. The correlation between (E/E_{PT} and rainfall is described by the expression:

$$\frac{E}{E_{PT}} = 0.923 + 0.00018\, P_m$$

with a regression coefficient of 0.30.

In this environment the mean monthly air temperature varies by just 2 to 3°C over the year. The relationship between λE and λE_{PT} shown in Figure 10(b) is therefore in large measure a reflection of the relationship between total evaporation and the available radiant energy in this case. The proportion of energy used for evaporation varies greatly from day to day. On fine days typically 75 to 80 per cent of energy is used for evaporation (Shuttleworth *et al.,* 1984a) but on rainy days evaporating water from the wet canopy routinely absorbs energy in considerable excess of that locally available as radiation. Over the whole study period the evaporation process accounted for 89.5 per cent of the incoming radiant energy measured above the canopy at this site.

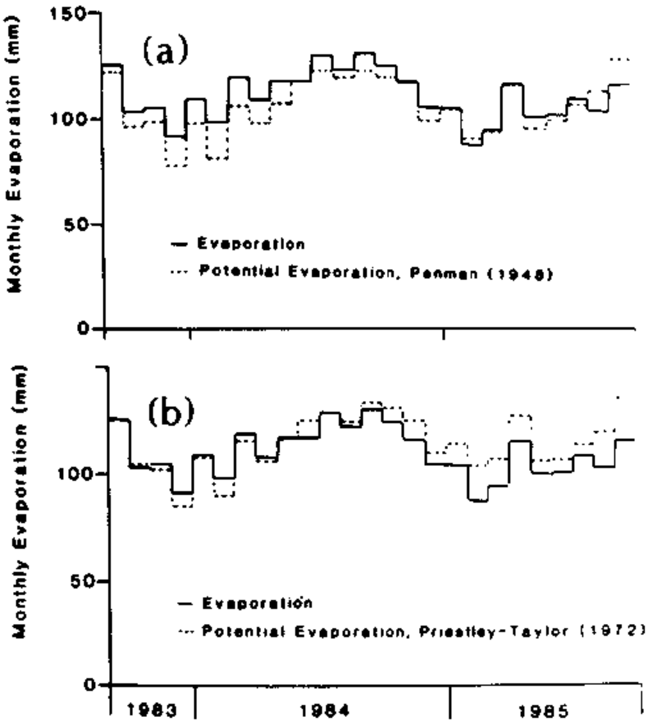


Figure 9 — (a) Monthly values of evaporation, E, and estimates of potential evaporation made with the Penman equation, E_p (see text), in mm for the 25 month experimental period.
(b) Monthly values of evaporation and estimates of potentital evaporation made with the Priestley-Taylor equation, E_{PT} (see text), in mm for the 25 month experimental period.

5. DISCUSSION

In the course of the present paper a simple, descriptive micrometeorological model has been constructed and using measurements of above canopy energy flux and below canopy throughfall made during a comprehensive, 25 month study of forest evaporation at an undisturbed forest site in central Amazonia. A set of routine hourly measurements of meteorological variables above the canopy, and regular measurements of soil water tension made throughout the study period provide the input to the model. The calculated monthly evaporation obtained given by this model agrees well, on average, with estimates of potential evaporation based on the Penman equation and the Priestley-Taylor equation. The total evaporation exceeds potential estimates by about 10 per cent in very wet months and can fall below potential estimates by about 10 per cent in dry months towards the end of the dry season.

The fact that evaporation from tropical forest at this site is close to estimates of potential evaporation estimates is an interesting and important result which is discussed later. However it is important to qualify this apparently simple observation, especially since it seems to contradict published data for forests in temperate climates, see for example Calder (1977). A short-term rate of evaporation from a portion of forest is more strongly dependent on surface control than that from other shorter vegetation types (Shuttleworth and Calder, 1979), and particularly depends on whether the canopy is wet or dry. The time-average behaviour is therefore a balance between less than potential evaporation rate, with a dry canopy, and greater than potential rate with a wet canopy. The near equality between average evaporation and potential evaporation in central Amazonia is therefore, in part, fortuitous, and should only be extrapolated to other portions of this forest with caution. It is, for instance, very probable that Amazonian regions with higher rainfall, where the forest spends a greater proportion of the time wet, will tend to have average evaporation rates greater than potential rate. Perhaps behaviour in the wetter portions of the year at this site may prove typical of behaviour at wetter sites, and *vice versa.* A more reliable extrapolation is possible by applying a canopy water balance model of the type described here using meteorological data measured above the canopy, similar to that given by the AWS in this study.

Notwithstanding the comments of the previous paragraph, it is important to remember that the climate of continental Amazonia is often held to be self-regulating. It is therefore probable that spatial variability of average energy partition in response to changes in average rainfall is less than for temperate, maritime climates. In Amazonia convection is the predominant rain generating mechanism: it is clear that the enhanced evaporation during and immediately after rain at one site is, in part, supported by energy entering the atmosphere as sensible heat at another site where evaporation is depressed by stomatal control. Energy conservation must act through this mechanism to moderate spatial variability if the system is partly enclosed.

In recent years there has been considerable concern regarding the consequences of a large scale and sustained land use change in the Amazon basin, and some of this concern has focussed on the possible consequences on local and global climate. The atmosphere and the land surface beneath it represents an interacting system, and the results and interpretation presented here do not therefore by themselves represent a complete basis for prediction. Rather they represent a description of the existing forest's response in central Amazonia during an arguably typical two year period. Clearly interpreting the consequences of any hypothetical change in land use on climate here and elsewhere is beyond the scope of the present paper. Nonetheless, these results, coupled with existing knowledge of the likely hydrological behaviour of alternative land use practices, could provide important pointers in this predictive process.

In this last context the most important result is the observation, emphazied earlier in this section, that the average evaporation over this arguably typical 25 month period was similar to that given by a Penman estimate of potential evaporation, *computed using meteorological variables measured above the forest canopy.* This sets the average level for the yearly evaporation flux. In addition there is the observation that evaporation from the forest exceeds potential estimates by up to 10 per cent in very wet months and (only) falls below these by about 10 per cent in drier months. This provides the order of magnitude of the present range of variation in evaporation with respect to potential rate.

It is instructive to consider a hypothetical, carefully managed land use change to stable grassland. Present understanding suggests that this will evaporate at the potential rate computed from meteorological variables *measured over the grassland surface* in months when precipitation exceeds this potential rate; but can fall to significantly less than this in the drier months of the year. It is an important point in this context to remember that calculated potential evaporation rates are systematically lower, by about 10 per cent, for grass covered surfaces than for adjacent forested surfaces. This is a direct consequence of the fact that the albedo of tall vegetation is less than that for short surfaces, and more of the incoming solar radiation is adsorbed. *Assuming there were no changes in climate,* the annual average evaporation rate would fall by about 10 per cent in consequence of the albedo change alone. In addition it will fall by up to 10 per cent in the wettest months of the year, when interception plays its most important role, and possibly by as much as 30 per cent in the driest months when the grassland surface can be expected to come under significant stress in consequence of its shallower rooting depth. A poorly managed land use change would exacerbate this difference. Soil compaction and the destruction of surface macropores during deforestation could dramatically reduce infiltration and increase surface runoff, thereby generating soil erosion and nutrient loss, and reducing the resilience and persistence of the ensuing grassland in dry months. The authors await with interest the interpretation of these estimated evaporation changes in models of global climate, and their consequential impact on agroeconomic models of Amazonian development.

Meanwhile the primary conclusions of the present study can be summarized as follows:

(i) The spatial variability in the precipitation throughfall beneath the canopy of Amazonian forest is very high, and can result in a large and systematic experimental error unless adequately sampled. This may well have contributed to the extreme variability in previously published results, and quite possibly towards an upwards bias in reported interception ratio and canopy storage capacity.

(ii) On average at this site, 50 per cent of the incoming rainfall is re-evaporated, about 25 per cent of this through the interception process, and the remainder by transpiration.

(iii) The annual-average actual evaporation calculated from a calibrated model of the hydrological processes agrees within experimental error with potential evaporation estimated from meteorological variables *measured above the forest canopy.*

(iv) The monthly-average evaporation similarly calculated is about 10 per cent higher than potential evaporation in wet months and about 10 per cent lower in dry months. The relative variation in daily rates is much greater, typically 20 to 30 per cent less than potential on fine days, but 20 to 70 per cent greater than potential rate on wet days.

(v) On the basis of these data, a land-use change to stable grassland (or bare soil) is likely to reduce Amazonian annual average evaporation by 10 to 20 per cent.

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REFERENCES

CALDER, I/R., 1977. A model of transpiration and interception loss from a spruce forest in Plynlimon, central Wales. *J. Hydrol.*, **33**, 247-265.

GASH, J.H.C., 1986. A note on estimating the effect of a limited fetch on micrometeorological evaporation measurements. *Boundary Layer Meteorol.*, 35, 409-413.

——— and MORTON, A.J., 1978. An application of the Rutter model to the estimation of the interception loss from Thetford Forest. *J. Hydrol.*, **38**, 49-58.

LLOYD, C.R., SHUTTLEWORTH, W.J., GASH, J.H.C. and TURNER, M., 1984. A microprocessor system for eddy-correlation. *Agric. and For. Meteorol.*, **33**, 67-80.

——— and MARQUES, A. de O., 1988. Spatial variability of throughfall and stemflow measurements in Amazonian Rain Forest. *Agric. and For. Meteorol.*, **42**, 63-73.

——— GASH, J.H.C., SHUTTLEWORTH, W.J. and MARQUES, A. DE O., 1988. The measurements and modelling of rainfall interception by Amazonian rain forest. *Agric. and For. Meteorol.*, in press.

MONTEITH, J.L., 1965. Evaporation and environment. *Symp. Soc. Exp. Biol.*, 19, 205-234.

MOORE, C.J., 1986. Frequency response corrections for eddy correlation systems. *Boundary Layer Meteorol.*, **37**, 17-35.

PENMAN, H.L., 1948. Natural evaporation from open water, bare soil and grass. *Proc. Roy. Soc.*, A, **193**, 120-145.

PRIESTLEY, C.H.B. and TAYLOR, R.J., 1972. On the assessment of surface heat flux and evaporation using large scale parameters. *Mon. Weather Rev.*, **100**, 81-91.

ROBERTS, J., CABRAL, O.M.R. and FERREIRA, L. DE A., 1988. Stomatal and boundary-layer conductances measured in an Amazonian Terra Firma Rain Forest. *J. Applied Ecology,* in press.

RUTTER, A.J., KERSHAW, K.A., ROBINS, P.C. and MORTON, A.J., 1971. A predictive model of rainfall interception in forests. I. Derivation of the model from observations in a plantation of Corsican pine. *Agric. Meteorol.,* 9, 367-384.

——— MORTON, A.J. and ROBINS, P.C., 1975. A predictive model of rainfall interception in forests. II. Generalization of the model and comparison with observations in some coniferous and hardwood stands. *J. Appl. Ecol.,* 12, 367-380.

SHUTTLEWORTH, W.J., 1988. Evaporation from Amazonian rain forest, 1988. *Phil. Tran. R. Soc. (Lond.) B,* 233, in press.

——— GASH, J.H.C., LLOYD, C.R., MOORE, C.J., ROBERTS, J., MARQUES, A. DE O., FISCH, G., SILVA, V. DE P., RIBEIRO, M.N.G., MOLION, L.C.B., DE SA, L.D.A., NOBRE, J.C., CABRAL, O.M.R., PATEL, S.R. and DE MORAES, J.C., 1984a. Eddy correlation measurements of energy partition for Amazonian forest. *Quart. J. R. Met. Soc.,* **110**, 1143-1162.

——— GASH, J.H.C., LLOYD, C.R., MOORE, C.J., ROBERTS, J., MARQUES, A. DE O., FISCH, G., SILVA, V. DE P., RIBEIRO, M.N.G., MOLION, L.C.B., DE SÁ, L.D.A., NOBRE, J.C., CABRAL, O.M.R., PATEL, S.R. and DE MORAES, J.C., 1984b. Observations of radiation exchange above and below Amazonian forest. *Quart. J. R. Met. Soc.,* **110**, 1163-1169.