Eddy correlation measurements of energy partition for Amazonian forest

By W. JAMES SHUTTLEWORTH, JOHN H. C. GASH, COLIN R. LLOYD, CHRISTOPHER J. MOORE and JOHN ROBERTS
Institute of Hydrology, Wallingford, Oxon, U.K.

ARI De O. MARQUES FILHO, GILBERTO FISCH, VICENTE De PAULA SILVA FILHO
and MARIA De NAZARÉ GOES RIBEIRO
Instituto Nacional de Pesquisas da Amazônia, Manaus, Amazonas, Brasil

LUIZ C. B. MOLION, LEONARDO D. De ABREU SÁ and J. CARLOS A. NOBRE
Instituto de Pesquisas Espaciais, São José dos Campos, São Paulo, Brasil

OSVALDO M. R. CABRAL
Centro Nacional de Pesquisa da Seringueira e Dende – EMBRAPA, Manaus, Amazonas, Brasil

SUKARAN R. PATEL
Universidade Federal da Paraíba, Campina Grande, Paraíba, Brasil

and

J. CARVALHO De MORAES
Universidade Federal do Pará, Belém, Pará, Brasil

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SUMMARY

Measurements of energy partition for Amazonian forest made with novel eddy correlation equipment are presented for eight dry days in September 1983. These are interpreted to provide estimates of the aerodynamic and surface resistance for this vegetation type. Daily total evaporation for a transpiring canopy accounts for 70% of the available radiant energy, and is two thirds of conventional estimates of potential evaporation. The results are used to provide an initial calibration of a simple, physically based model of daily evaporation for Amazonian rain forest.

1. INTRODUCTION

Progress in forest micrometeorology over the past fifteen years has been such that the acquisition, and even to some extent the interpretation, of data has become routine. Finance and practicability have dictated that the overwhelming majority of this work has occurred in temperate latitudes (e.g. Black and McNaughton 1971; Stewart and Thom 1973; McNeil and Shuttleworth 1975; Gay and Fritschen 1979). At the same time, there has been increasing awareness that tropical rain forest, and the continental rain forest of the Amazon basin in particular, may have an important role in global climatology. Energy partition measurements for this important vegetation type represent a significant gap in the literature.

This paper describes results from the first field season of a major Anglo–Brazilian collaborative study of the micrometeorology and plant physiology of Amazonian rain forest. Several intensive periods of data collection from micrometeorological instrumentation are planned over the next few years. Data collection during the first field season was limited by the need to first erect a meteorological tower and set up a field site in this difficult environment.

Arguably the most important single result from past studies of forest evaporation is the realization that interception loss, the evaporation of free water from the leaf surface during and after rain, and transpiration require separate description for this vegetation type (e.g. Shuttleworth and Calder 1979). This is a direct consequence of the
different magnitudes of the transfer resistances involved. During transpiration the latent heat flux is subject to a surface resistance, intimately related to stomatal control, which is generally at least an order of magnitude greater than the aerodynamic transfer resistance. When intercepted water is evaporating from a wet canopy this resistance is effectively short circuited and evaporation can alter dramatically.

It remains to be seen how important the interception component is in the context of evaporation from tropical rain forests, but its direct measurement and modelling have proved tractable elsewhere (e.g. Gash et al. 1980; Calder 1977), and its detailed study forms part of the experimental programme of the collaboration described above. On the other hand, the measurement and modelling of transpiration can be realistically studied only using micrometeorological techniques. In this paper we concentrate on this last aspect and present data corresponding to dry canopy conditions. These data are for eight dry days: 6, 7, 9, 10, 17, 18, 25 and 27 September 1983.

2. EXPERIMENTAL SITE AND BACKGROUND METEOROLOGY

The measurements were made using a 45 m scaffolding tower at a site, 2°57'S 59°57'W, selected as representative of the natural vegetation and regional topography in the Ducke Reserve, 25 km from Manaus, Amazonas. The plant density is high, up to 3000 stems ha⁻¹, but less than 10% have girths of 200 mm or more. The forest canopy is of very varied species and typical of undisturbed natural forest. It extends with no obvious sub-storeys to a height of 35 m, with occasional emergent trees reaching 40 m, one upwind and quite close to the experimental tower (within 20 m). Rooting in the yellow laterite soil is shallow, with a dense root mat down to 150 mm.

The topography is gently undulating with valleys several tens of metres deep occurring at about 300 m intervals. The topography of the canopy top is modulated by the differential growth of the vegetation. The tower is sited near the top of a broader than average ridge. The eddy correlation and profile measurements described in this paper assume that the fluxes are one dimensional and an adequate fetch is required for this assumption. The tower has fetches of undisturbed forest extending many kilometres in most directions. The minimum fetch occurs over a narrow range of angles downwind of the site; here the species content of the forest has been modified by human intervention about 1 km from the tower.

The climatological average rainfall pattern for this region exhibits a marked seasonal dependence, with a monthly maximum of 270 mm in March and a minimum of 40 mm in August. The data presented here were collected during September 1983, towards the end of a 'dry season' in which rainfall was anomalously high (150 mm per month). Over the period of data collection the specific humidity remained fairly constant in the range 16 to 19 g kg⁻¹, while specific humidity deficit above the canopy ranged from (2 ± 2) g kg⁻¹ at night to (10 ± 2) g kg⁻¹ in the early afternoon. The average temperature was in the order 27°C with a daily range of 7 to 10°C. Wind speeds were low, around (2 ± 1) m s⁻¹ with a strong easterly bias. Hourly average solar radiation exceeded 900 W m⁻² on rare occasions but peak daily values of 500 to 700 W m⁻² were more common. Net radiation varied between −40 W m⁻² at night and an average daytime maximum around 500 W m⁻².

3. INSTRUMENTATION

The evaporation, sensible heat and momentum fluxes presented in this paper were measured with a 'Hydra', a battery-powered eddy correlation, flux measuring device
developed by the Institute of Hydrology. The instrument was mounted on a pole above the tower at a height of 48.4 m. Measurements of the other meteorological variables used in this analysis were obtained using initially one, and later two, automatic weather stations mounted at the top of the tower at a height of 45 metres. These instruments are visible in Fig. 1, an aerial photograph of the tower. Towards the end of the experiment data from other instrumentation became intermittently available and are used in this paper. In particular for two of the days studied some temperature and humidity gradients were obtained with a Thermometer Interchange System (TIS). This last instrumentation has been described in detail by McNeil and Shuttleworth (1975), and this description is not repeated here.

![Aerial photograph of the scaffolding tower. The Hydra is mounted vertically above it at a height of 48.4 m, the two automatic weather stations are at the top of the tower at a height of 45 m.](image)

(a) *The Hydra*

(i) **Hardware components.** The Hydra system comprises a vertical sonic anemometer (Shuttleworth *et al.* 1982), a single beam, 2.7 μm, infrared absorption hygrometer (Moore 1983), a 50 μm thermocouple thermometer, and two Gill propeller anemometers mounted orthogonally in the horizontal plane. These sensors are interfaced to a battery-powered RCA 1802 CMOS microprocessor system which is integral to the device, and provides on-line calculation of fluxes, variances and mean values. The data are logged on a solid state store (G.K. Instruments) which, in this experiment, was removed and interrogated with a PET microcomputer every few days.

Manufacturers calibrations were assumed for the Gill propellers, but the other sensors were calibrated. The calibration of the infrared hygrometer exhibits a marked temperature dependence (see Moore 1983), and was calibrated against an EG & G Type 992 dew point hygrometer over a wide range of temperatures in a controlled temperature environment chamber. This chamber was also used to calibrate the thermocouple thermometer. The sonic anemometer was calibrated by extensive comparison against a Kaijo-Denki DAT 311 sonic anemometer, with both anemometers operating in field conditions. This empirical calibration was deemed necessary in view of defects in the Hydra anemometer performance, related in part to limited cosine response (Shuttleworth *et al.* 1982), and, possibly, in part to the interaction between the Hydra framework and
the wind field. Because of these effects, the calibration of the anemometer, and in consequence the flux measurements produced by the Hydra, could well be prone to systematic error in the order of 5% due to sonic anemometer calibration alone. Calibration defects in other sensors are believed to be small, in the order of one or two per cent, but could contribute to further potential error. Probably more important is the possibility of flux underestimation due to sensor exposure, sensor separation, and the frequency limitations imposed by sensor time constants and the averaging procedures used to evaluate the eddy flux. In the context of the present paper, there is clearly a need to check the performance of this novel device in this application. Such investigation receives detailed attention in later sections of this paper.

The Hydra system cannot provide reliable measurements during rain, when the hygrometer, sonic anemometer and thermocouple are wet, and the sensors generally take about an hour to dry out after a rain storm. Moreover the on-line software, outlined in the next section, uses running auto-regressive means with long time constants to identify the fluctuating components of the measured meteorological variables. The version of the software used in this experiment did not contain subroutines to provide rapid initialization of these auto-regressive means after a discontinuity in sensor availability, i.e. after initial ‘turn on’ or a heavy rainstorm. For this reason data produced by the software within three hours of such an event are not used in the present analysis. This fact was in part responsible for the decision to restrict the present analysis to completely dry days.

(ii) On-line calculations. The real-time software used in the Hydra microprocessor system is described in detail elsewhere (Lloyd et al. 1984).

In outline, the program takes samples of the fluctuating meteorological variables at a frequency of 10 Hz and computes the evaporation ($E$), sensible heat flux ($H$) and friction velocity ($u_*$) from the expressions

$$E = \langle w \rho_v \rangle - \langle w \rangle \langle \rho_v \rangle$$  \hspace{1cm} (1)

$$H = \rho c_p (\langle w T \rangle - \langle w \rangle \langle T \rangle)$$  \hspace{1cm} (2)

and

$$u_*^2 = \langle w u \rangle - \langle w \rangle \langle u \rangle$$  \hspace{1cm} (3)

where $w$ is the vertical wind velocity, $\rho_v$ absolute humidity, $\rho$ air density, $c_p$ the specific heat of air at constant pressure, $T$ temperature and $u$ the wind velocity in the direction of the mean wind. In these expressions $\langle \cdots \rangle$ symbolizes a moving average calculated for temperature, for example, by the expression

$$\langle T_i \rangle = a \langle T_{i-1} \rangle + (1 - a) T_i$$  \hspace{1cm} (4)

where $T_i$ is the present temperature sample, $\langle T_{i-1} \rangle$ the previous moving average and $\langle T \rangle$ the present resultant value of the moving average. The procedure is a digital analogue of a low pass electrical filter of time constant $\tau$, and the weighting function $a$ is given by the expression

$$a = \exp(-\Delta t/\tau)$$  \hspace{1cm} (5)

where $\Delta t$ is the time interval between inputs. The calculation of the variance of fluctuating variables is analogous to Eqs. (1), (2) and (3), and the variance of temperature, for example, is given by the expression

$$\sigma_T^2 = \langle T^2 \rangle - \langle T \rangle^2.$$  \hspace{1cm} (6)
The infrared absorption hygrometer is of the type first described by Hyson and Hicks (1975), in which humidity fluctuations are given by the ratio of the fluctuating signal to its mean component. In the Hydra this division is performed digitally. The calculation of $u_*$ requires knowledge of $u$, the wind speed in the direction of the mean wind. The program identifies this direction from suitably defined moving averages of the two orthogonal measured components, and provides the required wind vector by resolving in the direction of the mean wind. In addition to performing these calculations, the software monitors outputs generated by the sonic anemometer and hygrometer which indicate their functional status, and logs this as part of the data output.

An important consequence of calculating the flux, variance and mean according to the above prescription is that the resulting values are delayed with respect to real time. This effect is apparent in the data and a correction applied to take account of it is discussed in the next section.

There is some doubt over the most appropriate value of the parameter $\tau$ (and hence $a$) which should be used in all these calculations. It is necessary that it should be long enough to include most of the low frequency contributions in the co-spectra between vertical wind speed and the other fluctuating meteorological variables. At the same time very long time constants may make the calculations susceptible to contamination by instrument-generated low frequency correlations between the measured variables.

Previous experience suggests that time constants of at least 20 minutes are required, but, in view of the uncertainty, tests were carried out to investigate the suitability of different time constants in this environment. On some days two identical microprocessor systems were used in parallel to analyse the same sensor inputs using different values of the parameter $a$, corresponding to time constants of 18-25 and 31-25 minutes. Figure 2 shows a typical result, and illustrates the evaporation and latent heat flux obtained on 27 September 1983. The different time constants give two effects: the daily cycle in the measured flux with the longer time constant is delayed (by about 12 minutes); and is also slightly larger. Both of these effects are apparent in Fig. 2 but apply more generally.

![Figure 2](image)

Figure 2. Measured values of latent and sensible heat obtained from the Hydra with two computer systems operating in tandem using different time constants, $\tau$, in the autoregressive mean calculations. The full line corresponds to $\tau = 31-25$ minutes, the dashed line $\tau = 18-75$ minutes.
The integrated increase in measured flux given by the longer time constant over the four complete days for which comparative data were available was 2%. The longer time constant, 31.25 minutes, was used for all the results given in this paper.

(iii) Off-line corrections. The sonic anemometer calibration has a slight temperature dependence corresponding to the change in the speed of sound with temperature (Shuttleworth et al. 1982), while the calibration of the infrared hygrometer varies in a complex but defined way with both temperature and absolute humidity (Moore 1983). The data produced by the Hydra were passed through a calibration analysis to take account of these effects.

An adjustment has to be made to the measured latent heat to compensate for the temperature contamination of the humidity fluctuations which results from variations in atmospheric and sensor temperature (see Moore 1983). Corrections were also made for the effect described by Webb et al. (1980), who showed that density changes in the atmosphere, due to changes in temperature and humidity, induce small changes in the effect of a given vertical wind velocity. The joint effect of these corrections in the present data is some minor redistribution of measured flux during the day at the $10 \text{ W m}^{-2}$ level. The effect on integrated daily latent heat flux is negligible, typically less than 0.5%.

The measurement of horizontal wind speed made with two orthogonal Gill propeller anemometers is subject to the errors associated with this sensor, and in particular to stalling at low wind speed and defects in cosine response. The on-line software used in the Hydra assumes a constant calibration for these sensors. Retrospective, first-order corrections were made to those turbulence parameters (e.g. $u_*$, $\bar{u}$) which involve a measurement of horizontal wind speed. The correction procedure follows Drinkrow (1972) in expressing the response curve $R(\theta)$ of the Gill anemometer in the form

$$R(\theta) = \cos \theta - 0.085\sin(2|\theta|) + 0.036\sin(4|\theta|)$$  \hspace{1cm} (7)

where $\theta$ is the angle between the wind vector and the anemometer axis. For each hour the standard deviation of vertical wind speed and the mean wind vector were used to estimate the r.m.s. angle between the wind and the horizontal plane, and the two measured mean horizontal components used to provide an initial estimate of the orientation in the horizontal plane. A simple iterative procedure based on Eq. (7) was then used to derive correction factors for the two horizontal axes and related turbulence parameters.

It is recognized that the Hydra instrumentation provides a less than perfect measurement of $u_*$ as a result of defects in sensor performance and shortcomings in the corrections made for these defects. It is not possible to make simple corrections for anemometer stalling and, in the light and variable wind conditions prevalent during this experiment, this could well result in significant error, probably a reduction in all wind-speed-related measurements. Previous experience with this device (Wallace et al. 1984) suggests systematic underestimation in $u_*$, possibly as high as 30%, though more probably less than this. Moreover, the measurement of $u_*$ using eddy correlation techniques is in any case sensitive to errors associated with the distortion of mean wind flow around supporting structures and the instruments themselves. Estimates of the error in $u_*$ for the Kansas data (Businger et al. 1971) range between 7 and 33%, varying with the model used and the assumed effective size of tower obstructions (Wieringa 1980; Wyngaard et al. 1981; and Wieringa 1981). The results based on measurements of $u_*$ presented below exhibit significant experimental 'noise', and in addition could well be subject to systematic errors, possibly as large as 30%. It is perhaps fortunate that in the context of the present paper, and indeed in the broader context of energy partition from transpiring forests, large errors in the description of aerodynamic transfer have little effect.
Earlier mention was made of the fact that the application of autoregressive means in the real-time analysis carried out in the Hydra software has the effect of introducing a phase delay into the measurements it produces. The effect is in many respects analogous to the application of an electrical voltage to a simple RC filter, although the analogy is not perfect for flux and variance calculations because there can be some time dependence in the second term of Eqs. (1), (2), (3), and in particular of Eq. (6). We develop this analogy and make a first-order correction for the most important frequency component, that corresponding to the daily cycle. It is easily shown that, at low frequencies, the output of a simple low pass filter is subject to a phase delay with respect to the input which is equivalent to a time slip equal to the time constant, \( \tau \). A first-order correction was made to take account of this in the Hydra output. Accordingly, the value of (say) evaporation, \( E_i \), relevant to the \( i \)th hour in a particular day was calculated from the expression

\[
E_i = E_i(60 - \tau)/60 + E_{i+1}/60
\]  

where \( E_i \) and \( E_{i+1} \) are the latent heat fluxes calculated on-line for hours \( i \) and \( i + 1 \) respectively, and \( \tau \) is the time constant in minutes. Application of this procedure introduces an element of smoothing into the data.

(b) Automatic weather stations (AWS)

The experiment draws data from initially one and, after 8 September, two automatic weather stations (Didcot Instruments). These stations were used to provide the hourly average measurements of temperature, wet bulb depression, net radiation and wind speed used in the analysis. Independent measurements of these meteorological variables were intermittently available from other micrometeorological instrumentation and were used for intercomparison and verification when this occurred. In particular, the automatic weather stations were non-standard in that they included aspirated psychrometers. A comparison between the psychrometers and the accurate quartz crystal psychrometers used in the TIS, confirmed that the AWS psychrometers were adequately aspirated. The measurements of temperature and wet bulb depression provided by the AWS are, however, prone to experimental noise in the order 0-2°C.

The anemometers used in the AWS have metal cups of sturdy design and are prone to stalling errors at low wind speed (less than 0.5 m s\(^{-1}\)), and overspeeding at all speeds. In part such overspeeding is due to the fact that wind speed is \( a \ priori \) greater than the wind vector. Bernstein (1967) indicates that the size of this last effect is estimated by the expression

\[
s = \bar{u}(1 + \sigma_u^2/2\bar{u}^2)
\]  

where \( s \) is the wind speed, \( \bar{u} \) the mean wind speed vector and \( \sigma_u^2 \) the variance of the transverse wind speed. Preliminary estimates of this contribution to overspeeding, based on measurements of \( \sigma_u^2 \) obtained with the Hydra in daytime conditions, suggest an increase in the order of 10 to 15% at wind speeds around 2 m s\(^{-1}\). Mechanical overrun will further contribute to the overspeeding error. Despite these several shortcomings, the AWS wind speed is used throughout this analysis. It is the only measurement consistently available throughout the experiment, and is in any case typical of the type of wind speed measurement commonly available in climatological data, with which future energy partition models will necessarily be used. Its use does, however, further contribute to experimental noise and possible systematic uncertainties in the wind-speed-based results presented here.
A comparison was made between the two net radiation measurements provided by the AWS systems and independent measurements made with two Funk net radiometers which were available for part of the experimental period. Integrated daily values of net radiation agree to within systematic errors in the order 5%, but hourly values exhibit scatter greater than this, typically 10 to 15%. The hourly average value provided by the two stations is based on twelve individual measurements made at five-minute intervals and this sampling procedure is in significant part responsible for the experimental noise observed in the data. To minimize the effect, the mean values of net radiation, wind speed, temperature and wet bulb depression from the two weather stations are used in this analysis whenever both were available. Only one station was operational on 6 and 7 September and the results for these two days are perhaps less reliable because of this.

4. Results

(a) Integrated daily flux

Assuming horizontal homogeneity, the energy budget of the forest vegetation can be written as:

\[ A = \lambda E + H \]

\[ = R_n - G - S - P \]

where \( A \) is the available energy per unit area; \( \lambda E \) is the latent heat flux out of the forest; \( H \) is the sensible heat flux out of the forest; \( R_n \) is the net all-wavelength radiation; \( G \) is the energy flux into the ground; \( S \) is the equivalent energy flux into storage; \( P \) is the net rate of energy absorption per unit area by the combined processes of photosynthesis and respiration, estimated to be of the order of a few per cent of \( R_n \) and neglected here.

The presence of a storage term in this energy budget adds considerable complexity to the measurement and modelling of the daily cycle in forest energy partition. Early attempts to model this component simply (e.g. Stewart and Thom 1973) proved incapable of describing sustained nighttime radiation (Thompson 1979). More recent, sophisticated computer models (e.g. Goudriaan 1979) suggest that storage can be a very significant contribution to the energy budget, both in terms of magnitude and duration. The size, complexity and density of tropical forest is such that energy storage is likely to be a particularly important component, but difficult to calculate.

By adopting an eddy correlation measurement technique we successfully avoid the need to make the calculation, but the presence of the term complicates validation of the measurement to some extent. Accordingly we choose to assume that, to first order, the storage term, \( S \), and indeed the soil heat flux, \( G \), in Eq. (11) are approximately zero over an integrated daily cycle. Within this assumption, the cumulative net radiation over a complete day should equal the cumulative sum of the eddy correlation measurements of latent and sensible heat, i.e. \( \Sigma R_n = \Sigma A = \Sigma (\lambda E + H) \). It should be remembered that this check is not completely definitive in that it cannot rule out the possibility of a fortuitous numerical cancellation between daytime and nighttime errors.

In Fig. 3 we present the comparison of cumulative net radiation and the cumulative sum of outgoing energy flux for the eight complete days considered in this analysis. The agreement is satisfactory at about a 5% level, but integrated over all eight days indicates a flux loss in the eddy correlation measurement of 6-3 MJ m\(^{-2}\) in a total radiant input of 96-5 MJ m\(^{-2}\). Two days, 6 and 10 September, are significantly worse than the other six and no definite reason is advanced for this, although wind directions on 10 September
and the early morning of 6 September, when most of the shortfall occurs, were such as to leave the Hydra sensors poorly exposed.

Figure 3. Comparison of the cumulative net radiation and the cumulative sum of latent and sensible heat measured by the Hydra for the eight fine days studied in this analysis. The 1:1 line is shown dashed.

(b) Energy fluxes

Figure 4 illustrates the daily energy exchange cycle for the eight fine days considered in this analysis. In most respects the behaviour is similar from day to day. There is a fairly constant partition of the available energy during daylight hours as latent and sensible heat, often some sustained evaporation into the early evening, and outgoing radiation in the order 30 to 40 W m$^{-2}$ as the forest cools during the night. The suppression
of turbulent interaction between the forest and the atmosphere at night, typified by the low energy fluxes, is a feature of the data and applies to all the turbulence variables. In fact the detailed behaviour of measured turbulent fluxes at night is difficult to interpret, and it is suspected that hourly measurements produced by the Hydra are subject to apparently haphazard fluctuations in the order 10 W m$^{-2}$ which are of instrumental origin, perhaps related to correlations in the long term drift of the component sensors. No doubt such noise also occurs during the day but is less apparent: presumably the effect on integrated flux measurements is small.

Table 1 presents the integrated daily flux of radiant energy, latent and sensible heat for the eight fine days together with the 'evaporative fraction' defined in two ways:

$$\alpha = \frac{\Sigma \lambda E}{\Sigma (\lambda E + H)}$$ (12)

and

$$\alpha' = \frac{\Sigma \lambda E}{\Sigma R_p}$$ (13)

The value of this parameter is reasonably similar for the eight days studied, the value
TABLE 1. INTEGRATED DAILY ENERGY FLUXES FOR THE EIGHT FINE DAYS CONSIDERED IN THE ANALYSIS, TOGETHER WITH THE EVAPORATIVE FRACTION DEFINED BY Eqs. (12) AND (13).

<table>
<thead>
<tr>
<th>Date (Sept. 1983)</th>
<th>Integrated daily energy flux (MJ m⁻²)</th>
<th>Evaporative fraction</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Radiation</td>
<td>Latent heat</td>
</tr>
<tr>
<td>6</td>
<td>12.58</td>
<td>7.59</td>
</tr>
<tr>
<td>7</td>
<td>14.12</td>
<td>10.32</td>
</tr>
<tr>
<td>9</td>
<td>10.94</td>
<td>8.13</td>
</tr>
<tr>
<td>10</td>
<td>6.67</td>
<td>4.28</td>
</tr>
<tr>
<td>17</td>
<td>14.89</td>
<td>10.20</td>
</tr>
<tr>
<td>18</td>
<td>10.99</td>
<td>9.18</td>
</tr>
<tr>
<td>25</td>
<td>14.09</td>
<td>9.05</td>
</tr>
<tr>
<td>27</td>
<td>12.27</td>
<td>8.64</td>
</tr>
<tr>
<td>Mean</td>
<td>12.06</td>
<td>8.42</td>
</tr>
</tbody>
</table>

of α' is generally lower than α because of the integrated eddy flux loss mentioned in section 4(a). It is also more variable from day to day since it includes the additional variability in this loss factor. However, the implication of Table 1 is clear: the daily evaporative fraction for transpiring tropical forest for these data is around 0.7, corresponding to a Bowen ratio, β (= H/LE), of 0.43. We return to this important result later.

Earlier mention was made of the fact that on two days, 25 and 27 September, a Thermometer Interchange System (TIS) of the type described by McNeil and Shuttleworth (1975) was operational for significant portions of the day. It is possible to use this to provide some confirmation of the above result. The system provided two differential measurements of temperature and humidity between the height ranges 3.5-6 to 39.3 m and 40.9 to 44.6 m. Under the assumption that the fluxes of sensible and latent heat obey a one-dimensional diffusion equation, and that the eddy diffusivity for heat and water vapour are equal, the ratio of the differential temperature and humidity measurements provides a measure of the Bowen ratio in steady state conditions (see McNeil and Shuttleworth for details). The evaporative fraction is directly calculable from this ratio through the expression α = (1 + β)⁻¹. Hourly measurements made in this way tend to be noisy because the differential gradients are small.

Figure 5 illustrates the daily trend in α calculated for the two portions of the TIS and compares this with that calculated from the eddy correlation measurements for daylight hours when relevant measurements were available. Although these results are subject to considerable experimental noise, they are in broad agreement, with the TIS results perhaps tending to favour an even lower evaporative fraction.

(c) AERODYNAMIC RESISTANCE ESTIMATES

The aerodynamic resistance for momentum transfer, r_A, is given by the expression

\[ r_A = u/(u_*)^2 \]  \hspace{1cm} (14)

where \( u \) is the measured wind speed and \( u_* \) is the friction velocity. In neutral conditions (Thom 1975) it is usual to write

\[ u_*/u = k \ln \{(z - d)/z_0\} \]  \hspace{1cm} (15)

where \( k \) is von Kármán's constant (assumed equal to 0.41), \( z \) is the measurement height of the wind speed \( u \), and \( z_0 \) and \( d \) are the roughness length and zero plane displacement of the vegetation.

Our purpose in this section is to provide a working description of \( r_A \) for use in
Figure 5. The variation in evaporative fraction, \( \alpha = \frac{E}{(H + LE)} \), for daylight hours on 25 and 27 September as measured by the Hydra (---) and the upper (---) and lower (-----) portions of the Thermometer Interchange System.

...deriving surface resistance and in later discussion. For this purpose we reference the aerodynamic transfer resistance to the wind speed at 45 m measured by the automatic weather stations for the reasons mentioned in section 3(b). Measurements of \( u_* \) at 48.5 m are drawn from the Hydra and we assume no divergence in momentum flux between the measurement heights of these two instruments.

There is reason to believe that aerodynamic resistance to the transfer of energy flux differs from that for momentum in non-neutral conditions (e.g. Stewart and Thom 1973), and all aerodynamic resistances are subject to stability corrections. In practice this complexity is of very little numerical importance in modelling evaporation from forests for reasons which become clear in the following sections. Consequently such models generally assume simple formulations corresponding to neutral stability (e.g. Gash et al. 1980), and some assume a constant value for this resistance (e.g. Gash and Stewart 1977). In accordance with this philosophy we present data for daytime conditions, when...
the vast majority of flux transfer occurs, regardless of atmospheric stability. The data are, however, heavily selected to correspond to wind directions in which the two Gill propeller anemometers are reasonably well exposed and where their mutual interference is minimized. Only mean wind directions which are within 45° of the bisecting angle between the two propellers are considered. In fact the large variability in wind direction in this environment is such that, even with this selection, some interference cannot be ruled out in the hourly average values considered. The data are further selected to exclude hours when they are unreliable by virtue of their timing relative to initial turn-on of the Hydra system, or sonic anemometer loss during rain.

Measured values of $u_s$ are plotted against AWS wind speed in Fig. 6(a), together with a line of linear regression constrained to pass through the origin. The gradient of this line is $0.175$ which, from Eq. (15), corresponds to a value of $(z - d)/z_0 = 10.4$. In

![Figure 6](image-url)

Figure 6. (a) Values of $u_s$ measured with the Hydra as a function of $u$ measured with the automatic weather station. The dashed line is a line of linear regression constrained to pass through the origin. (b) Values of aerodynamic resistance, $r_A$, deduced from measurements of $u_s$, made with the Hydra, and $u$, made with the automatic weather station, expressed as a function of $u$. The dashed line is the curve $r_A = 337/u$ used as a working description in this analysis.
previous studies of momentum transfer to forest canopies, attempts have been made to refer observed values of \( d \) and \( z_0 \) to canopy height \( h \), but the results are extremely variable. The values given in Table V of Jarvis et al. (1976), when interpreted for this experiment by setting \( h = 35 \text{ m} \) and \( z = 1.29h \), give estimates of \( (z - d)/z_0 \) in the range 6 to 20, with a mean value of 10.1.

Assuming the value 10.4 for \( (z - d)/z_0 \) and combining Eqs. (14) and (15) gives the function

\[
 r_A(u) = 33/u
\]  
(16)

which we adopt as a working description of these data for the purpose of this analysis. Figure 6(b) compares this function with values of \( r_A \) calculated from the individual measurements using Eq. (14). The results are subject to considerable scatter particularly at low wind speeds when problems associated with propeller and cup anemometer stalling, directional variability and fractional overspeeding errors are most important. Errors here are in the order of 100%; the dependence at higher wind speed is slightly better defined but scatter is still in the order of 50%. We claim no better accuracy than this for Eq. (16). Although most of the scatter in Figs. 6(a) and (b) is probably of instrumental origin, large experimental noise in the measurement of the aerodynamic properties of forests is in fact an extremely common, though rarely discussed, phenomenon. It is probable that the absence of steady state conditions and the (relative) proximity of measurement systems to the surface make some contribution.

\( d \) Surface resistance

On rearranging the Penman–Monteith equation (Monteith 1965), the surface resistance, \( r_s \), is given by

\[
 r_s = \{(\Delta' c_p)/(\beta - 1)\} r_A + \rho \lambda q_w(T) - q / \lambda E
\]  
(17)

where \( \Delta' \) is the slope of the saturated specific humidity curve at mean air temperature; \( \rho \) is the density of air; \( \lambda \) is the latent heat of vaporization of water; \( c_p \) is the specific heat of air at constant pressure; \( q_w(T) \) is the saturated specific humidity at \( T \), the air temperature at height \( z \); \( q \) is the specific humidity at height \( z \); and other quantities are defined earlier in the text.

Values of surface resistance were calculated from Eq. (17) for each hour using Eq. (16) to estimate \( r_A \). In this way a value of \( r_A \) is consistently available from the AWS wind speed measurement. The alternative, direct measurements of \( r_A \) using Eq. (14) has restricted availability as a result of wind direction constraints on the \( u_\ast \) measurement. For the eight days considered the surface resistance follows a similar daily trend and the eight-day median value is presented in Fig. 7. The resistance falls rapidly with daybreak and then stays fairly constant around 130 \( \text{s.m}^{-1} \) until early afternoon, after which time it rises progressively, increasing by almost an order of magnitude by early evening.

It is worth noting at this stage that the coefficient of \( r_A \) in Eq. (17) is small. For Bowen ratios around 0.43 and mean temperatures of 28°C, a 50% change in a typical value of \( r_A \) (18 \( \text{s.m}^{-1} \)) only causes a 3% change in a typical value of \( r_s \) (130 \( \text{s.m}^{-1} \)).

The daily trend in surface resistance shown in Fig. 7 is in qualitative agreement with some separate, direct measurements of stomatal resistance made in parallel with these data. The physiological cause of this behaviour is not clear at this stage. There could well be some response to vapour pressure deficits, since these are higher in the afternoon, but the behaviour is moresuggestive of a response to a daily increase in leaf water potential. The rapid decrease at dawn and increase at dusk could well reflect a dependence on radiation level.
5. DISCUSSION

The data presented in section 4 provide an initial specification of the aerodynamic and surface controls on energy partition for Amazonian forest. Assuming this site is representative, they are arguably typical of natural forest in mid-continent Amazonia in conditions when soil moisture is freely available. The results may be of interim use in hydrological and meteorological energy partition models of dry canopy evaporation in conjunction with short-term meteorological variables, measured (or model specified) near the forest canopy. Their usefulness in this role is, however, presently curtailed by the limited time period and soil moisture conditions they describe, and, equally importantly, by the present lack of an experimental calibration of model parameters in an equivalent short-term model of wet canopy (interception) loss. Extending the descriptive range of the dry canopy specification presented here, and providing a calibration of detailed wet canopy models remain important objectives of the collaborative study described in section 1. We therefore postpone further discussion of this potentially important application of these data.

All models of energy partition are ultimately empirical, their complexity, physical realism and descriptive accuracy are conditioned by the data requirement involved in their use (Shuttleworth 1979, 1983; Stewart 1983). The past absence of any measurements against which more complex but physically realistic models could be calibrated in the Amazon basin, has necessarily required the speculative use of energy partition models with a high empirical content. Moreover there is, in this region, a shortage of short-term, above-canopy meteorological data with which complex models could in any case be operated—a shortage which is unlikely to be rectified in the near future. In this situation, there is a clear need to interpret the present data in the framework of existing, empirical energy partition models, and if necessary to modify these. We concentrate on this aspect in the remainder of this section.

Figure 7. The median value of surface resistance, \( r_s \), for the eight fine days considered in this analysis for daylight hours.
(a) Comparison with estimation formulae

In the phraseology of classical hydrology, the experimental situation in which the data presented in this paper were gathered represents 'evaporation from a well-watered crop'. The meteorologist and hydrologist schooled in the concept of potential evaporation might therefore be surprised by the result (given in section 4(b)), that the fraction of incoming radiant energy used in evaporation for a transpiring forest canopy is around 0.7. To the forest hydrometeorologist this result is entirely plausible, and not unexpected (cf. Shuttleworth and Calder 1979). It reflects the fact that transpiring forest canopies exert significant surface control even when water is freely available in the soil.

To emphasize this point we present, in Table 2, values for some of the more commonly used estimation equations, together with the water equivalent of total incoming radiation and the measured values of actual evaporation. If \( Q_N \) is the integrated radiation input, \( D \) the daily average specific humidity deficit, and \( u \) the daily average wind speed (m s\(^{-1}\)); then, expressed in SI units and referenced to specific humidity (rather than vapour pressure), these estimation equations take the following form:

Penman evaporation (Penman 1948)

\[
\lambda E_P = \left( \Delta' Q_N + \rho C_p D f(u) \right) / (\Delta' + c_p/\lambda)
\]  

(18)

where

\[
f(u) = 0.004(1 + 0.54u);
\]  

(19)

Thom–Oliver evaporation (Thom and Oliver 1977), with values of the constants suggested for general use,

\[
\lambda E_{TO} = \left( \Delta' Q_N + \rho C_p 2.5 D f(u) \right) / (\Delta' + 2.4 c_p/\lambda);
\]  

(20)

Priestley–Taylor evaporation (Priestley and Taylor 1972),

\[
\lambda E_{PT} = 1.26 \Delta' Q_N / (\Delta' + c_p/\lambda);
\]  

(21)

equilibrium evaporation, the steady state, asymptotic limit approached by a closed atmosphere over a large expanse of vegetation with fixed surface resistance (McNaughton and Jarvis 1983),

\[
\lambda E = \Delta' Q_N / (\Delta' + c_p/\lambda).
\]  

(22)
Inspection of Table 2 reveals that all the estimates of potential evaporation \( \lambda E_p \), \( \lambda E_{TO} \), \( \lambda E_{PT} \), and even \( \lambda E_E \) exceed actual evaporation in dry canopy conditions, and the first two exceed the available radiation input. The potential evaporation estimates, which agree to within 5%, exceed actual evaporation by a factor 1.5. Equilibrium evaporation is a much closer approximation to measured transpiration.

(b) Towards a practical estimate of daily evaporation

Given the results presented in section 4, i.e. high measured values of surface resistance and comparatively low values of aerodynamic resistance, the conclusion drawn from work in temperate forest (cf. Shuttleworth and Calder 1979), that transpiration and the evaporation of intercepted rainfall must be estimated separately, applies equally to Amazonian rain forest. Gash (1978) demonstrated that the approach advocated by Thom and Oliver (1977) is in essence the Penman–Monteith equation (Monteith 1965) written in a form suitable for daily application. We follow this philosophy and write, as a working model of daily evaporation for forest surfaces, the equation

\[
\lambda E_F = \frac{\Delta' Q_N + \rho c_p D/r_A}{\Delta' + (c_{p/2}/\lambda)(1 + r_{sd}/r_A)} + I(1 - c)
\]  

(23)

where \( I \) is a model calculation (or measurement) of interception loss; \( r_{sd} \) is the effective daily value of dry canopy surface resistance in well watered conditions; and \( c \) compensates for including transpiration in wet conditions, and is given by

\[
c = \frac{\Delta' + c_{p/2}/\lambda}{\Delta' + (c_{p/2}/\lambda)(1 + r_{sd}/r_A)}.
\]  

(24)

In their original formulation Thom and Oliver recommend the equation

\[
r_A = 53 \ln^2[(z - d)/z_0]f(u)
\]  

(25)

as an estimate of aerodynamic resistance. In the present context we prefer to combine Eqs. (14) and (15) and adopt

\[
r_A = \ln^2[(z - d)/z_0]/k^2u
\]  

(26)

as a working description of the aerodynamic resistance, with \((z-d)/z_0 = 10.4\). The values of \( r_A \) given by these two expressions differ significantly at low wind speeds but the consequences of this difference in estimates of transpiration in Eq. (23) is small, typically in the order 2%.

The first term in Eq. (23) describes transpiration and the present data provide an initial calibration of the empirical constant \( r_{sd} \). Taking mean values of the measured variables, measured over the eight days considered in this analysis, and using Eq. (23), to calculate the value of \( r_{sd} \) which gives the evaporative fraction 0.7 suggested by Table 1, we deduce that the effective daily value of dry canopy resistance for this forest is 215 s m\(^{-1}\). The fact that the effective daily value of dry canopy resistance is significantly greater than the measured midday values presented in section 4(d) serves as a reminder that \( r_{sd} \) is an empirical parameter linked to daily average values. Evaporation, which largely occurs during the day, is implicitly assumed to occur more slowly over twenty-four hours, and the effective resistance is increased.

The correction factor \((1 - c)\) in Eq. (23) compensates for the fact that some transpiration is calculated even under wet conditions (cf. Gash 1978). Using the effective dry canopy resistance \( r_{sd} = 215 \text{ s m}^{-1} \) in Eq. (24) gives a value of \((1 - c) = 0.71\). This is less than that typical of temperate forests, e.g. 0.93 (Shuttleworth and Calder 1979), and
this in part reflects the fact that the rates of evaporation in dry and wet canopy conditions are more comparable for tropical forests since mean temperatures are higher.

The second term in Eq. (23) describes the evaporation from the wet forest canopy and implies the separate existence of a direct measurement or sub-model of the interception loss, $I$. Gash (1979) created an analytic model of interception based on physical principles similar to those used in the numerical model of Rutter et al. (1971). In so doing he demonstrated that there is at least some physical justification for expressing the interception loss from a forest in terms of a regression on precipitation input. The regression coefficients depend on mean rainfall rate and mean wet canopy evaporation rate, and on the canopy storage capacity. For temperate forests the interception component can make an important and possibly dominant contribution to total evaporation. Preliminary estimates (see later) suggest that interception is of less relative importance in this environment and in this period contributes only about 30% of total evaporation. In this way the errors involved in rather a coarse model of $I$, in Eq. (23), have a proportionately reduced percentage error in estimates of $\lambda E_F$.

In the present study parallel measurements of interception loss were made during September and early October 1983 at a site adjacent to the meteorological tower. Measurements of the rainfall amount reaching the ground were made with 16 rain gauges which were relocated randomly along a linear transect after each rainstorm. The total interception loss was 32-7 mm in a total precipitation, $P$, of 193-25 mm (16-9%). Using $I = 0.169 P$ as a working model of interception loss during September 1983, and taking meteorological data from the automatic weather stations, Eq. (23) predicts a mean daily evaporation of 3.73 mm for this period. This can be compared with the mean daily estimates $\lambda E_F = 4.27$ mm, $\lambda E_{PT} = 4.15$ mm, $\lambda E_{TO} = 4.30$ mm and $\lambda E_E = 3.29$ mm, given by Eqs. (18), (20), (21) and (22) applied over the equivalent period. Over these, atypically wet, days in September 1983, an estimated 48% of the precipitation falling on the site returned to the atmosphere by evaporation.

The measurements of transpiration given in section 5(a), and the estimates of total evaporation given in this section, together typify the general behaviour of forest evaporation at a catchment scale (cf. Shuttleworth and Calder 1979). When the canopy is dry, evaporation proceeds at rates which are significantly less than those suggested by estimation formulae developed and calibrated over short vegetation ($\lambda E_F$, $\lambda E_{PT}$ and $\lambda E_{TO}$), rates which are more comparable to equilibrium evaporation ($\lambda E_E$). However, during periods when rainfall is common, as in the present data, the interception loss tends to compensate (or even over-compensate) for this. The best understanding of forest micrometeorology currently available, calibrated with the dry canopy data presented here, and merged with preliminary estimates of interception loss, suggests that evaporation from Amazonian rain forest will:

(i) approach, and possibly exceed, accepted estimates of potential evaporation in wet months (rainfall 270 mm per month);  
(ii) fall to 70% of potential rates, or less, in dry months (rainfall 30 mm per month).

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