The forest contribution to the hydrological budget in Tropical West Africa

B. A. MONTENY (1), and A. CASENAVE (2)

(1) *ORSTOM, Laboratoire de Bioclimatologie INRA,* 78850 *Thiverval-Grignon, France*
(2) *ORSTOM, Laboratoire d'Hydrologie, B.P.* 375, *Lomé, Togo*

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ABSTRACT. Micrometeorological data from a rubber tree plantation in the Ivory Coast were used to define a simple, empirical model of its evapotranspiration, expressed in terms of equilibrium evaporation, available soil water, precipitation and incident solar energy. This model is assumed representative of the regional evaporation. When incorporated into a hydrological model, it is found to give a satisfactory description of the catchment water balance covered with undisturbed natural forest some distance away. On the basis of this agreement, the average energy balance given by the model is assumed to be broadly representative and used to argue that the water vapour content of the southwesterly trade wind does not decrease when passing over the large-scale forest, in spite of the heavy rainfall.

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INTRODUCTION

It is possible that large-scale modification of vegetation cover in the Tropical regions may have effects on regional climates (Aubreville, 1949 ; Bernard, 1953 ; Henderson-Sellers and Gornitz, 1984 ; Kandel and Courel, 1984 ; Gornitz, 1985). Recent improvements in the modelling of large scale climate dynamics (Walker and Rowntree, 1977 ; Shukla and Mintz, 1982 ; Eagleson, 1982 ; Manabe, 1982 ; Rind, 1984 ; Dickinson, 1983, 1986 ; Hunt, 1985 ; Sellers *et al.*, 1986 ; Wilson *et al.*, 1987 ; Druyan, 1987) show the importance of the earth surface characteristics on atmospheric conditions (fig. 1). As yet, many of these simulations have not been compared to equivalent data.

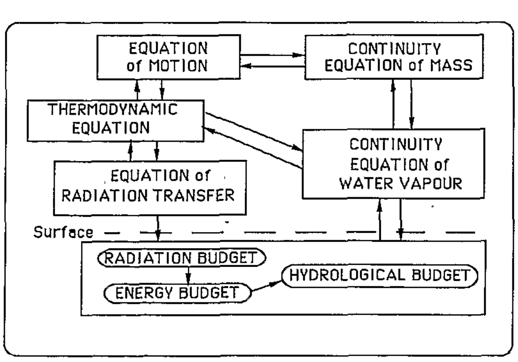

Figure 1
Schematic diagram of a General Circulation Model of the atmosphere showing the incorporated processes of surface hydrology (Manabe, 1982).

Hydrological studies in the Tropical forest ecosystems have been conducted by ORSTOM for more than three decades (Roche, 1982 ; Casenave *et al.*, 1982 ; Dubreuil, 1985, 1986). Instrumented catchments provided water input-output budgets to determine average evaporation loss as opposed to short-term evapotranspiration estimates. However, until recently, little quantitative knowledge about Tropical forest evapotranspiration and interception losses were available (Huttel, 1975 ; Lloyd *et al.*, 1988b ; Shuttleworth, 1988). It is generally agreed that forests, in relation to other land uses, consume more water on a yearly basis.

By using water budget methods associated with the energy budget, a better understanding of the importance of the quantitative water vapour exchange processes can be obtained (Pinker *et al.*, 1980 ; Shuttleworth *et al.*, 1984a ; Monteny *et al.*, 1985 ; Calder *et al.*, 1986 ; Shuttleworth, 1988).

The first purpose of this study is to model the evapotranspiration rate from known plant physiological activities, climatic and soil properties. This has already been inferred from previous studies on *Hevea* plantations using the energy budget (Monteny *et al.*, 1985 ; Monteny, 1987).

The second purpose is to model a complete hydrological catchment budget, covered completely by Tropical forest. The structure of the model is based on statistical and physical equations which define each component of the hydrological and energy balance. P 160

METHODS

Energy balance technique

Micrometeorological measurements allow the determination of mean water vapour transfer from the vegetated surface to the atmosphere. The aerial evapotranspiration rate has been measured by the energy budget-Bowen ratio method (Monteny, 1985) given as :

$$Rn = \lambda E + H + G + P \pm \Delta S \quad (\mathrm{W\,m^{-2}\ or\ MJ\,m^{-2}\,d^{-1}}) \quad (1)$$

where Rn : net radiation ; λE : latent heat flux density ; H : sensible heat flux density ; G : soil heat flux density ; P : energy absorbed for photosynthesis ; ΔS : change in energy stored in the air and the biomass between the levels of measurement of G and Rn.

Generally, most of the radiant energy absorbed by the forest canopy is converted into sensible and latent heat flux densities on a daily basis. Rn is positive when there is an energy gain by the canopy ; E and H are positive when there is an energy loss by the surface. Assuming that the fluxes are constant with height above the canopy, the introduction of the Bowen ratio (ratio of the sensible heat flux to the latent heat flux : β) in a simplified form of eq. (1) gives the instantaneous evapotranspiration rate of the surface :

$$E \approx (Rn - G)/\lambda \cdot (1 + \beta)\,. \quad (2)$$

Direct determination of evapotranspiration (micrometeorological techniques) is based on point measurements taken in the atmosphere and soil at selected sites. A 22 m high open frame tower equipped at its top with a platform of 2 m^2 was erected in this plot. From this, a 11 m high mast was mounted on which the sensors are fixed. The height of 33 m is necessary to record the relatively small gradients of temperature, dew point, CO_2 concentration and wind speed. (See for more detail, Monteny, 1987.)

Data analysis provides a description of the plant control of water loss (leaf area and stomatal-root resistances) which is also incorporated into the evapotranspiration rate models.

Hydrological measurements

Rainfall into a landscape initiates changes in the soil-vegetation-water continuum. Little work has been so far undertaken on the quantification of the hydrological processes in the humid forest regions of the Tropical zone. Water budget studies provide baseline data on which analyses of climatic characteristics depend.

Generally, watershed data provide water input-output budgets, and the net watershed loss due to evapotranspiration and interception rates is calculated by difference.

The hydrological budget of a catchment area is given by :

$$P = Q + R + Pi + E \pm \Delta sw \ (\text{mm per unit time}) \quad (3)$$

where P : precipitation (mean for the catchment) ; Q : drainage to water table which runs over as base flow ; R : overland flow ; $Q + R$ = runoff ; Pi : precipitation intercepted by the canopy ; E : actual evapotranspiration rate ; Δsw : variation of the volumetric soil moisture in the root zone ;

Most of the hydrological balance components are measured on a daily time scale. The variation in solar energy Rg and rain P inputs induces changes in soil moisture storage which affect evapotranspiration. Whereas the monthly water budget can be informative, shorter budget evaluations (10 days) give a more detailed understanding of the hydrological processes.

Precipitation is the only water input in the study catchment area. A network of fifteen recording rain gauges gives acceptable accuracy for the estimated precipitation volume over the experimental catchment area (38 km^2) (Casenave *et al.*, 1980). The individual gauges supply the data from which the isohyetal map can be drawn. They describe the spatial rainfall variability and give the average amount of rainfall over the catchment area. Because of the weakness of the rain gauge network in the African forest zone and in order to minimize any error by extrapolating measured precipitation to a larger scale, these mean values are compared with those measured at a nearby meteorological station taken as a reference site.

The stream discharge is measured at the outlet catchment stream equipped with a calibrated V-notch weir and a recording stage gauge monitors continuously the water height of the river. Rainfall of sufficient intensity and duration exceeding the soil surface infiltration capacity induces overland flow after the soil water holding capacity has been recharged. Streamflow Q and intermittent overland flow (after rain), R, are deduced from the graph (Casenave *et al.*, 1980, 1984 ; Collinet and Valentin, 1979). Streamflow and overland flow are converted to millimetre depth, giving unit area runoff values.

The changes in soil water storage (Δsw) in the root zone of the forest vegetation is predominantly seasonal. In southern Ivory Coast, the maximum storage occurs during May-June until late July and from mid-September to November, depending mainly upon the Meteorological Equator (M.E.) motion which induces rainfall. The minimum soil water storage is measured generally during February-March, at the end of the dry season. The difference between these two values of the volumetric soil moisture represents the maximum available moisture *swc* stored in the root zone, equivalent of 220 mm for 3.5 m depth. The ratio of actual available soil moisture *swd* to his maximum value represents the soil moisture wetness.

Synoptic situation

The Ivory Coast tropical rain forest, situated at the border of the Gulf of Guinea, is influenced by the oceanic effects. Seasonal changes in climate can be considerable due to the movement of the Meteorological Equator (M.E.). The motion of the M.E. over the forest zone determines the occurrence of dry and rainy seasons (Wauthy, 1983 ; Leroux, 1983 ; Collinet *et al.*, 1984). The tropical forest is subjected to 1-3 months of dry season with a period of a few weeks when the northern wind is blowing intermittently (the Harmattan). This continental air drastically reduces the humidity, temperature and visibility. The wet season is associated with the tropical rain belt with moisture laden southwesterly winds. The rain characteristics are associated with the structure of the M.E. (Leroux, 1988) :

- in front of the M.E., the InterTropical Front (ITF) structure separates two distinct air masses in temperature and humidity. This structure, with high precipitated water potential, gives frequent heavy rainstorms of short duration (March-May and November-December) ;
- behind the ITF, the InterTropical Convergence Zone (ITC) structure corresponds to the tropical rain belt and is responsible for continuous and abundant rainfall (i.e. rainy season).

Micrometeorological measurements were undertaken in a *Hevea* plantation (70 km^2) at Dabou, near Abidjan (5° 19′ N-4° 13′ W). Watershed measurements were conducted on an undisturbed forest (38 km^2) at Taï, in the south western region (5° 45′ N-7° 23 W).

RESULTS AND DISCUSSION

The experiment identified and measured the dominant processes, thus allowing satisfactory predictions to be made (Monteny, 1985, 1987).

Radiation budget

In the Tropics, net radiation measurements at the canopy interface are not usually available even though they are of primary interest. Following some previous measurements (Monteny *et al.*, 1981), net radiation can however be derived from the total solar radiation *Rg* or sunshine duration which are more readily available. The following relations were obtained between net radiation and incident solar radiation where *Rn* and *Rg* are expressed in W m^{-2} (table 1).

Table 1

Forest albedo and relationship between net radiation Rn and global solar radiation Rg for different tropical forests.

	Albedo (%)	$Rn = \mathbf{a}Rg + \mathbf{b}$ (W m^{-2}) a	b	r	n
Rubber forest (humid air)	14	0.72	− 0.8	0.98	282
Rubber forest (dry air)	14	0.78	− 50	0.98	113
Oil palm forest (humid air)	13	0.71	− 12	0.98	135
Amazonian forest (Shuttleworth *et al.*, 1984*b*)	12	0.86	− 35	—	—
Thaïland forest (Pinker *et al.*, 1980)	12	0.87	− 25	0.99	—
Puerto Rico forest (Odum, 1970)	12	0.72	—	—	—

Rubber, oil palm and Tropical rain forests have higher regression factors (**a**) than all other canopies in this area, mainly due to the low reflectivity and weak thermal fluxes. The long wave radiation balance has little influence on the net radiation flux in relation to the high atmospheric water vapour concentration (2.8 kPa). This is not the case when dry continental air is blowing over the forest region as can be seen from table 1.

Compared with other results over forests, the factor **a** for oil palm and rubber forests is lower but it is associated with a weak long wave radiation budget.

Water budget

Rainfall intercepted by the canopy is considered to evaporate directly. The amount of intercepted water is calculated as the difference between gross precipitation (measured at the top of the canopy) and the through-fall reaching the ground. Because only rainfall totals are readily available, simple relationship is derived by considering their efficiency for prediction.

Through-fall was measured from 39 linear rain gauges of 0,05 m^2, arranged in 3 lines of 110 m each (direction : N-E, S-W and S-N) on the forest floor close to the gross rainfall gauge site. Since it was possible to take readings twice a day (at 7 h and during the day hours 1 h after the rainfall), a considerable amount of single periods of continuous rain exists. The calculated standard deviation of the throughfall represents 24 % of the gross precipitation (Cardon, 1979).

Intercepted rainwater by a dry forest canopy and the daily individual gross precipitation gives the following regression equation :

$$Pi = 0.773 \cdot \ln\ (1 + P) + 0.138 \quad (\mathrm{mm\ d^{-1}}) \quad r_2 = 0.86\,. \quad (4)$$

The maximum canopy (leaf and branches) storage capacity is evaluated at 2.8-3.0 mm (Cardon, 1979). The intercepted quantity depends on the precipitation characteristics and on the vertical leaf distribution, leaf area index (LAI) varying from 6 to 8 (Alexandre, 1981). The variability of the intercepted amount is enhanced by the canopy movement due to the wind during rainstorms. The evaporation resulting from this canopy water storage is assumed to occur at the Penman potential rate which is 25-30 % higher than the actual rate for well watered canopy. This is due to the surface resistance, r_v, which decreases from 80-100 s m^{-1} for dry canopy without soil moisture limitation to nearly zero with intercepted water.

Stemflow, expressed as a depth of rainfall over the canopy, is very small and neglected. But if expressed as a volume of water, stemflow can represent a considerable input to the soil since it is concentrated around the base of the tree, depending upon its structure.

Stem water is collected by an adhesive plastic collar put on 16 adjacent trees covering an area of 300 m^2 (Huttel, 1975). The volumes of the containers were recorded during two years (1970 and 1971) and the following expression was obtained :

$$\text{stemflow}: \quad 0.00033 \cdot P^{1.68} \quad (\mathrm{mm\ d^{-1}}) \quad r_2 : 0.92\,.$$

The percentage of gross rainfall which reached the soil surface as stemflow represents 0.83 % of the annual precipitation (1800 mm), compared with 1.8 % for the Amazonian forest (Lloyd and Marques, 1988*a*).

Priestley and Taylor (1972) proposed a formula to calculate the regional evapotranspiration rate. It is based on large scale data rather than on micrometeorological results : the quantity of energy required for evaporation comes predominantly from net radiation Rn, and the effect of vapour pressure deficit is introduced by a coefficient C in the equation depending on the surface wetness :

$$E = C \cdot E_0 = C\,(\Delta/\Delta + \gamma) \cdot (Rn - G)\,. \quad (5a)$$

The equilibrium evaporation rate

$$E_0 = (\Delta/\Delta + \gamma) \cdot Rn$$

(Davies and Allen, 1973) is more representative of the climatic demand in this humid atmosphere. The ratio between the measured value of the actual evapotranspiration rate of the rainforest (measured by micrometeorological techniques) and E_0 gives the coefficient C, also called crop coefficient (Katerji and Perrier, 1983). Generally, in daytime conditions, C is taken as 1.26 for well watered crops in temperate regions (Priestley and Taylor, 1972) corresponding to a value of $\beta = 0.05$. The soil heat flux G is generally neglected because it represents only 1 to 2 % of Rn in the case of rain forests.

The value of C results from different processes which control the water vapour transfer from the leaf surface to the atmosphere (Jarvis *et al.*, 1976 ; Perrier, 1980) : aerodynamic r_a and climatic r_c resistances and mean canopy resistance r_v which depend on leaf physiological activities through the growth stages :

$$C = [1 + \{\gamma/(\Delta + \gamma)\}\,r_c/r_a]/[1 + \{\gamma/(\Delta + \gamma)\}\,r_v/r_a]. \quad (5b)$$

From the micrometeorological measurements, figure 2 gives the diurnal variation of C for daytime hours, in relation to plant phenology and soil conditions :

04.28.82 : young leaves (2 months old) without soil moisture limitation

04.15.81 : young leaves (2 months old) with a soil water deficit

12.22.81 : old leaves (10 months old) and well watered.

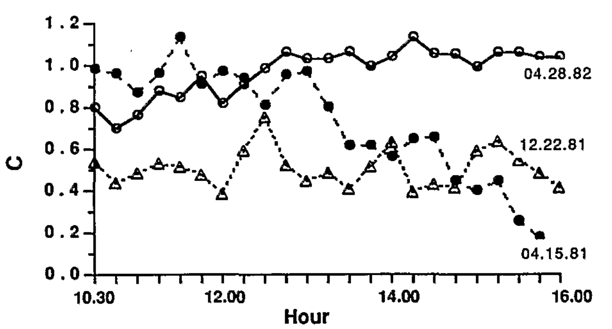

Figure 2
Diurnal variation of the coefficient C for different plant and soil conditions.

It shows that :

- a well watered canopy with young leaves has the maximum transpiration rate compared with equilibrium evaporation E_0, $(0.9 < C < 1.1)$;
- the soil water deficit affects the actual evapotranspiration rate of a young and well developed canopy surface as noted by the reduction of the coefficient C to a mean value of 0.55 ;
- leaf ageing reduces C as shown by 12.22.81.

Figure 3 illustrates the relationship between C and the calculated canopy resistance values.

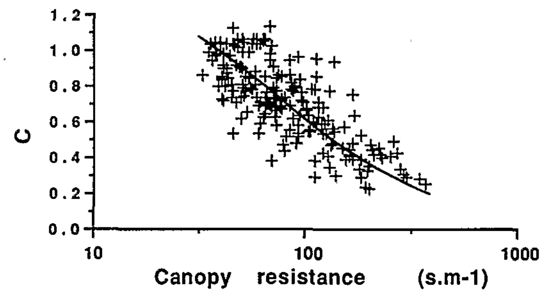

Figure 3
Relationship between the coefficient C and the forest canopy resistance (dry surface) for different soil water contents.

The reduction of the evapotranspiration rate is mainly related to an increase of the canopy resistance r_v (eq. (5*b*)), the climatic resistance changing only slightly from one day to another. C varies from 0.3 to 1.1 depending on leaf physiological activities in relation to the soil moisture availability. Higher values of C could be measured during the rainy season due to the evaporation of intercepted water. Without soil water limitations, C varies between 0.80 and 1.1 for dry canopy conditions depending on leaf surface properties and atmospheric demand. Actual evapotranspiration rate under dry surface conditions is generally near the equilibrium evaporation rate E_0.

Because of small variation in the atmospheric water vapour deficit in this climatic zone, C is more characteristic of the stomatal regulation (plant physiological properties) in relation to water availability in the soil volume explored by the root system. Soil moisture depletion thus affects canopy behaviour and offers a water component relatively easy to evaluate.

Weekly soil moisture content, rain quantities and the fraction intercepted by the canopy were measured during the year 1983 to evaluate the actual evapotranspiration rates. The general response of the forest is given, using C, in figure 4.

Soil moisture depletion affects the relative evaporation rate only when nearly 40 % of the total soil moisture *swc* in the root zone has been extracted, confirming the relationship in figure 3. The point distribution in the graph is mainly due to changes in the climatic parameters (variability of incoming solar energy and precipitation). The fact that the canopy behaviour depends on soil moisture content contributes to an expression of forest evapotranspiration rate in the Tropical humid zone.

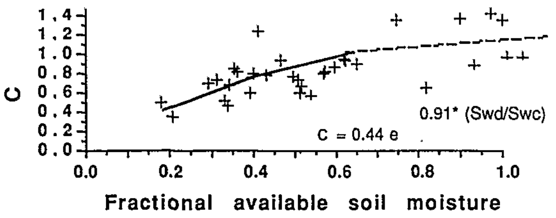

Figure 4
Forest relative evapotranspiration rate in relation to the fraction of depleted soil moisture by the root system (depth : 3.5 m*).*

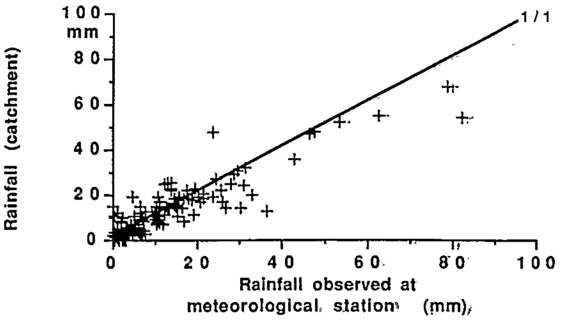

Figure 5
Relationship between daily precipitation values at the meteorological station and the mean rainfall over the catchment area (Taï, August 1979 *to July* 1980*).*

For a well developed forest (LAI $\geqslant 4$), the evapotranspiration rates were evaluated from the relationship presented in figure 4 :

$$E/E_0 = C = [0.44 \cdot e^{0.91 \cdot (swd/swc)}] \qquad (6a)$$

which gives :

$$E = [0.44 \cdot e^{0.91 \cdot (swd/swc)}] \cdot (\Delta/\Delta + \gamma) \cdot Rn \,. \qquad (6b)$$

The influence of soil water depletion on canopy transpiration is incorporated in the model as an average value of the soil moisture availability over the whole catchment. Because the transpiration rate in humid regions is mainly due to radiative energy, the evapotranspiration rate is based on the radiation budget as in the Priestley and Taylor (1972) formula. Meteorological information is scarce for the forest regions and the Penman-Monteith combination method (Monteith, 1965) must be used carefully at the catchment scale for estimating water vapour transfer to the atmosphere (McNaughton and Jarvis, 1983 ; Mawdsley and Ali, 1985).

Dunin and Aston (1985) found the same kind of relation with an Eucalyptus forest as in figure 4. They show the effect of the leaf area index evolution and the variability of the climatic demand on coefficient *C*. In the case of the Tropical rain forests, the leaf area index does not change greatly during the year, leaf fall or growth not being simultaneous for all species. *C* is thus not influenced in the same way as for temperate deciduous forests (Singh and Szeicz, 1980 ; Sharma, 1984).

Our regional evaporation model is consistent with the concept that soil moisture depletion exerts an action on the canopy resistance which controls the water use.

Water catchment balance studies provide an indirect measurement of the total aerial evapotranspiration rate which must be compared with the computed evapotranspiration rate. By subtracting catchment losses (stream, Q, and overland flows, R) from average catchment rainfall, P, the total amount of aerial evapotranspiration rate and the changes in soil water storage are estimated $(E + Pi + \Delta sw)$.

Input of rainwater into surface boundaries of the catchment must be known with a certain accuracy. When using climatic data for modelling catchment hydrological budget, the validity of observed meteorological values and the average catchment values must be compared. This is particularly important concerning water input (precipitation) and evapotranspiration rate. Figure 5 illustrates the relationship between mean daily catchment rainfall data with the values observed at the meteorological station from August 1979 to July 1980 so that only one gauge recording measurement could be representative.

While there is some daily variation, a fairly good correlation exists which allows to use the precipitation data of the meteorological station in order to determine rain input in the hydrological catchment model.

Streamflow (base flow + overland flow) is one of the most common runoff surface processes. Overland flow occurs when the hydraulic conductivity of the forest soil surface is saturated. The floor surface depends on certain soil characteristics such as the multichannel macropore system and the soil horizon moisture.

Daily precipitation *versus* overland flow of more than 1 mm rainfall equivalent at the weir discharge (38000 m^3) are plotted in figure 6. Data equivalent to less than 1 mm were discarded.

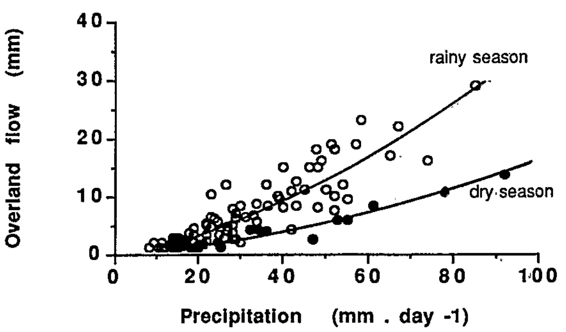

Figure 6
Relationship between daily rainfall and overland flow for discharge flow at the weir, higher than the equivalent of 1 mm *during dry and rainy seasons.*

A distinction is made between dry and humid climatic conditions, giving two curves in relation to the soil water status which affects an overland flow volume. Under dry conditions, soil surface of a forested catchment plays a major role in the rapid absorption of stormflow due to the empty macropore system. The corresponding equation is :

$$\text{overlandflow} : R = 0.85 \cdot (e^{0.003 \cdot P}) \quad r_2 = 0.81 \,. \qquad (7a)$$

But when precipitation or successive rainfalls of sufficient intensity and duration exceed the soil surface infiltration capacity, overland flow takes place after saturation. This gives the following equation :

$$R = 1.012 \cdot (e^{0.043 \cdot P}) \quad r_2 = 0.72 . \qquad (7b)$$

When the soil surface macropores are near saturation, as during the rainy season, precipitation discharge induces more overland flow. The previous soil moisture conditions and the rainfall intensity/duration are of critical importance on the water speed infiltration (Collinet, 1983).

On the basis that the catchment does not collect water from outside the topographic surface boundaries, the forest water losses and the changes in the soil water storage can be estimated using the hydrological budget. The seasonal evolution of the forested catchment water balance components is presented in figure 7.

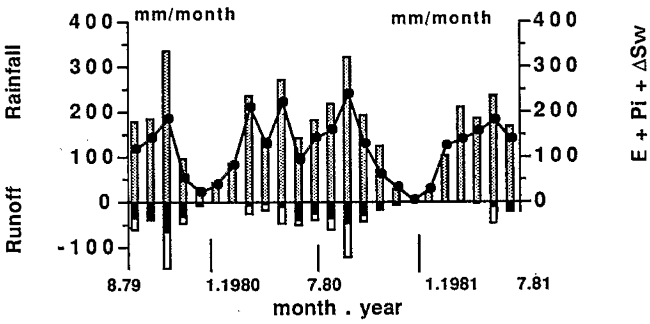

Figure 7
Evolution of the water balance components of a forested catchment in the humid tropical region at Taï, Ivory Coast ; for runoff, baseflow equals black bars and overland flow white ; total evaporation rate + soil water storage variation : (•).

Precipitation in the tropical rain belt (ITC structure) of the M.E. is characterized by high intensity associated with large droplets : October 1979 ; March to May and September 1980 and May 1981 are typical. Due to the saturated surface soil, the infiltration rate is reduced and the rain water flows on the forest floor. Streamflow or runoff is very low during the dry season (from January until March) and the total losses $E + Pi + \Delta sw$ follow the same course.

Without soil moisture measurements, catchment water balance is not precise enough to evaluate accurately the forest evapotranspiration rate during certain periods of the dry and the rainy seasons due to the variations in soil moisture storage Δsw. On a yearly basis, the cumulative evaporation from August 1979 to July 1980 represents 1442 mm for 1936 mm of rainfall and 1452 mm for 1806 mm the following year, or 66 and 59 % of the precipitation amount respectively (see table 2).

In order to know accurately the water vapour transfer from the forest to the atmosphere, the evapotranspiration rate needs to be evaluated using the biophysical formula which takes into account both climatic demand and biological reactions.

Table 2
Annual values of the water balance components of the catchment (mm year^{-1}).

P	Overland flow	Base flow	$E + Pi \pm \Delta s$	$E + Pi$ estimated	E/Rn	E/P
08.79-07.80 1936	226	266	1442	1279	0.77	0.66
08.80-07.81 1806	172	182	1452	1059	0.69	0.59

A daily water budget model is developped that satisfactorily simulates evapotranspiration and interception losses by the forest, as well as overland flows.

HYDROLOGICAL CATCHMENT MODEL

Water vapour balance of the atmosphere over a region depends on :

- the regional transfer of water vapour from the soil-canopy to the atmosphere ;
- the condensation-precipitation of this atmospheric water vapour to the soil-vegetation surface ;
- the budget between incoming-outgoing water vapour across the air volume boundaries above the region.

Only the first two points will be considered by the model to show the importance of the West African forest zone in maintaining climatic stability in this region.

The model presented here is used to simulate the forested (or cultivated) watershed behaviour and to obtain estimates of the long-term evapotranspiration losses from the catchment surface. It also determines the range of soil-moisture deficits which could occur during the dry season. But on a regional basis, difficulties arise in relation to the data needed to define the flows and storage levels for detailed modelling. Our hypothesis is that methods which are proven on a small catchment (38 km^2) are also appropriate for describing processes over larger watershed areas.

The daily transpiration rate, E, is calculated using (eq. (6b)). The fraction of water intercepted on the canopy (eq. (4)) evaporates at a potential rate, transpiration is assumed to be zero. The evaporation of intercepted rain is 25-30 % higher than the equilibrium evapotranspiration rate E_0 on a daily basis. So during rainy days the total represents the actual evapotranspiration rate. The catchment model has only a single soil storage compartment retaining a fraction of rain water. The excess water drains towards the watertable which is represented as base flow.

Some specific characteristics must be known such as the mean depth of the root zone of forest trees, the evolution of the leaf area index, soil hydraulic conductivity, water storage capacity. The structure of the model is shown in a schematic diagram (fig. 8). The different components of the model are as follows :

- interception is related to the amount of daily precipitation (eq. (4)).

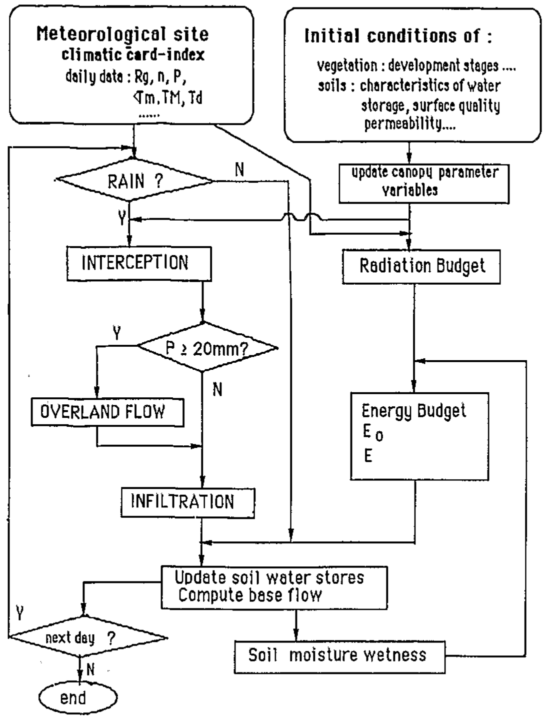

Figure 8

Schematic diagram of the hydrological catchment model.

• evapotranspiration is related to equilibrium evaporation (a function of the net radiation) which is controlled by soil moisture wetness in the root zone (eq. (*6b*)) ;

• overland flow depends on the amount of daily precipitation (eqs. (*7a*) or (*7b*)) ; rainfall of sufficient intensity and duration when exceeding the soil surface infiltration capacity induces overland flows ;

• soil water status of the catchment determines the volume of drainage flow generated after the rainfall.

From the data of the nearby meteorological station and the different equations based on average values from a representative regional catchment, the hydrological components of a watershed have been deduced from August 1979 to July 1981. One of the most important terms, the evapotranspiration, will be discussed here in more detail. Figure 9 shows results on the catchment water behaviour for 1979-80.

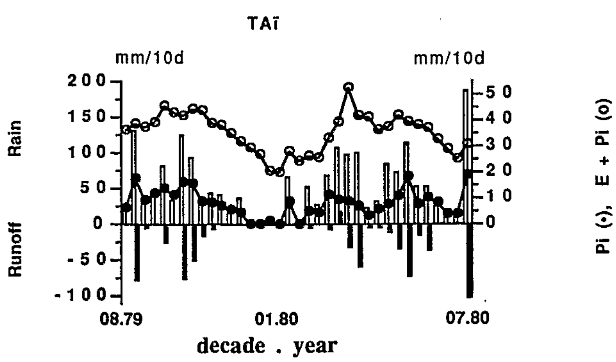

Figure 9

Seasonal water balance based on a ten day period : observed rainfall and estimates of total runoff (white and black bars respectively), intercepted rain evaporation (•) and total evapotranspiration rates (O) of a forested catchment, Taï, Ivory Coast.

During the rainy season, the forest evapotranspiration rate is higher than the equilibrium evaporation due not only to transpiration, the major source of water loss, but also to a significant amount of intercepted rain which evaporates directly from the forest canopy.

During the dry season, the evapotranspiration rate depends on the available solar radiation. But, after a few weeks, the soil moisture wetness (*swd/swc*) affects the canopy resistance and hence the transpiration rate. The transpiration decreases while the equilibrium evaporation increases.

The comparison between the estimated evapotranspiration loss and the results obtained from the hydrological balance (fig. 10) shows some discrepancy in relation to the inertia of the catchment flows.

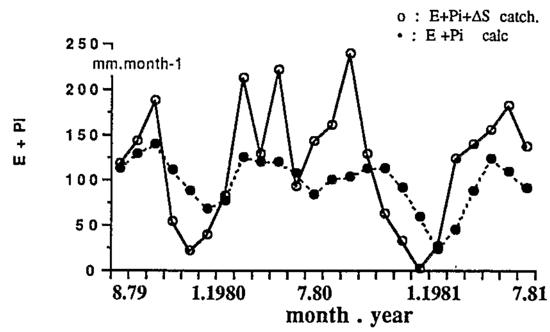

Figure 10

Comparison between measured ($E + Pi + \Delta s =$ O) and estimated total evapotranspiration rates ($E + Pi =$ •) of the undisturbed forested catchment during 2 years.

The water loss by the forested catchment, estimated by the hydrological equation, depends on the frequency of moderate-heavy rainfall in this rain forest environment. The comparison between these losses and the estimated aerial evapotranspiration gives some idea about what happens over a period when the amount of water is stored or depleted in the catchment soil (fig. 10), particularly during rainless periods.

Storage is a combination of soil moisture changes and water table fluctuations. The changes in soil moisture in the root zone are predominantly seasonal, depending on the rainfall distribution. The mean available soil water for the tree roots from the ground surface to a depth of 3.5 m is 220 mm. Evapotranspiration generally depletes the soil moisture during the dry season, reaching the wilting point towards the end of the season. At this time, the catchment baseflow is very low. Changes in the watershed storage term are generally neglected on an yearly basis, so the hydrological budget simulation results agree better with the experimental work (Morton, 1983 ; Holmes, 1984).

Nearly all the net radiation input is transferred as latent heat to the atmosphere during the wet season [$(E + Pi)/Rn = 0.95$], due to the rainfall frequency and the high interception rate of the forest canopy (fig. 11).

With the reduction of soil moisture availability, the available net radiation, if evapotranspiration rate decreases, is transferred as sensible heat to the air

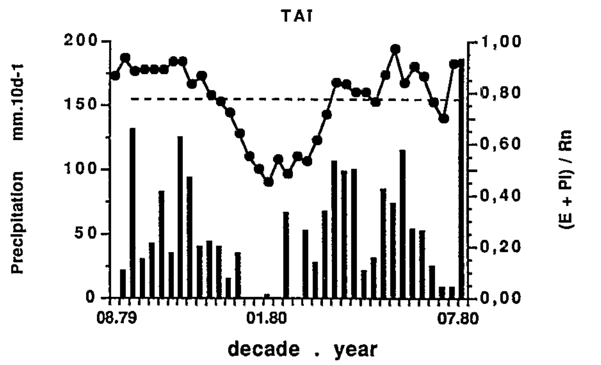

Figure 11
Evolution of the relative energy transfer as water vapour to the atmosphere (•) *in relation to the rainfall distribution, at Taï, Ivory Coast.*

masse (eq. (1)) above the catchment $[(E+Pi)/Rn=0.50]$. This induces an increase in the atmospheric temperature and therefore an increase of the water vapour pressure deficit.

The West African Tropical forest injects the equivalent of 60 to 75 % of the annual precipitation as water vapour into the atmosphere depending on rainfall frequency and seasonal distribution (table 2). It confirms the results obtained by Huttel (1975) at Banco, a location in the Ivory Coast forest zone. This represents the equivalent of 70 to 80 % of the yearly net radiant energy. Calder *et al.* (1986) found 100 % in the West Java forest and Shuttleworth (1988) found that it amounted to 90 % in the Amazonian forest. The variation of the transfer amount is due mainly to the precipitation distribution at the different forested locations.

The regional recycling of rainfall water has been calculated for different locations from the coast to the northern border of the Sudanese forest zone (fig. 12).

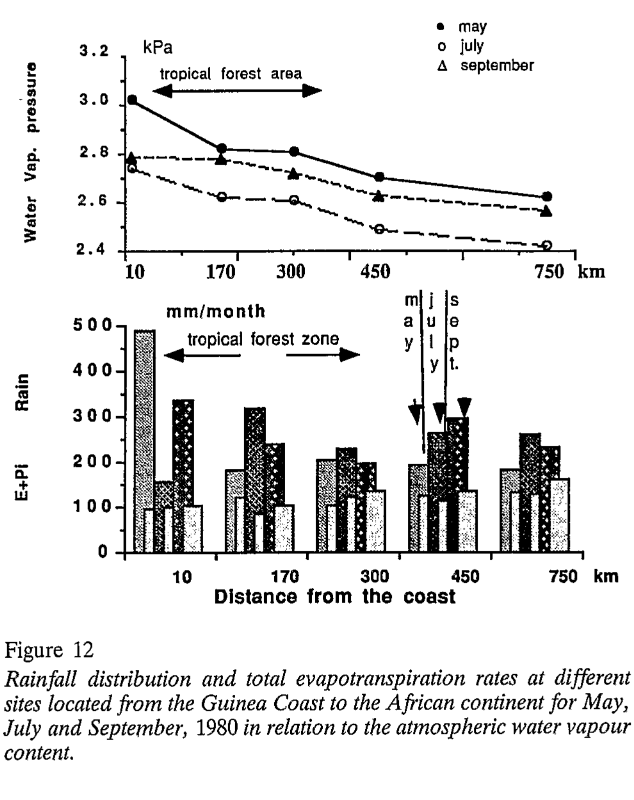

Figure 12
Rainfall distribution and total evapotranspiration rates at different sites located from the Guinea Coast to the African continent for May, July and September, 1980 *in relation to the atmospheric water vapour content.*

The forested land acts as a water source for the air mass layer of the ITF within the Meteorological Equator. The depletion of the atmospheric water vapour content by precipitation is reduced by the evapotranspiration rates, with the mean monthly vapour pressure decreasing slowly (fig. 12).

The Tropical rain forest accounts for a large turnover of the precipitated water back to the atmosphere during the shift of the Meteorological Equator. From April to November, the importance of the latent heat transfer from forested surface to the air mass (monthly evapotranspiration rates between 85 and 150 mm) mostly depends on the available solar energy. Especially in July and August, the ITC structure over the forest area consists mainly of different cloud stratifications which absorb and reflect a large amount of solar energy : only 25 to 35 % of extraterrestrial radiation is transmitted to the forest canopy (Monteny *et al.*, 1981 ; Lhomme and Monteny, 1982). This climatic parameter is the main limiting factor which could restrict the atmospheric moisture supply by the vegetated land. Strong upwelling during the period of June-July in the oceanic equatorial zone (Bakun, 1978 ; Merle, 1983 ; Hisard *et al.*, 1986 ; Lamb *et al.*, 1986) contributes to the condensation of this atmospheric water vapour forming large scale cloudiness (stratocumulus, altostratus) (Monteny, 1986*a*, *b*).

The large forested area induces steady state in the atmospheric characteristics : water vapour content and temperature which are determined by the available energy and the surface resistance. These characteristics lead to an equilibrium value for the evapotranspiration rate (Perrier, 1982). The forest zone tends to maintain itself in a moist state : water vapour pressure over the forested area from the coast to 400 km inside the continent does not decrease despite variations in monthly rainfall. Water vapour transport between the surface and 850 mb (equivalent to 3000 m high) over West Africa shows a progressive penetration of moisture from the coast to the north (Cadet and Nnoli, 1987). The climate of the continental Amazonian forest has been demonstrated by Salati and Vose (1984) to be partially self regulating : the turnover of water vapour between air mass and forest is rapid, maintaining a moist and cloudy overlying atmosphere.

Without soil water limitation, the forest actual evaporation is at the equilibrium rate, but during oceanic upwelling periods, the sea surface-atmosphere interaction affects the air vapour pressure and induces important modifications in the global atmospheric circulation (GAC), affecting some climatic parameters in this region as rain distribution. Their impacts on GAC have been simulated by Walker and Rowntree (1977).

CONCLUSIONS

Estimated forest evapotranspiration rates, obtained from soil and micrometeorological relationships, give an insight into the dynamics of water movement through catchment areas. Soil water content in the root zone is one factor which affects the actual evapotranspiration in the humid Tropical zone. The evaluation of the different processes according to the soil-climatic conditions allows remodelling of the micrometeorological dynamics of the forest zone. The comparison of the evaporative water losses obtained using the hydrological budget or estimated with the hydrological catchment model shows some differences due to the soil water depletion and to the inertia of the catchment flow with time.

Forest vegetation in the West African Tropical region transfers the equivalent of 75 to 80 % on an average of the yearly net radiation as latent heat. It varies from 50 % for the drier month of the year to 95 % during the rainy season. This re-evaporation represents 60 to 75 % of measured rainfall.

The forest area plays the same role as a large energy converter. It acts as an important water vapour source for the above air mass : the mean monthly water vapour pressure decreases only slowly from the southern coast to the northern border of the forest region (400 km inside the continent) despite the importance of the precipitation amount. The main limiting factor is the available solar energy. Strong upwelling during June-July in the oceanic equatorial zone reduces the atmospheric instability and contributes to an important cloud cover, intercepting a large amount of solar energy.

The hydrological catchment model presented here could be easily applied to link general circulation models with regional watershed behaviour. The climatic impact assessment could be derived from any surface modification due to human activities.

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