Amazonia's Hydrologic Cycle and the Role of Atmospheric Recycling in Assessing Deforestation Effects

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ABSTRACT

Atmospheric and hydrologic phases of the water cycle in Amazonia are modeled coherently by subdividing the basin (which extends from about 47.5° to 77.5° west longitude) into six adjacent regions each 5° wide. Due to long-distance travel of the air masses, the recycling of vapor evaporated from upwind land surfaces is important. For example, it is found that 88% of the water precipitating on the westernmost region falls at least a second time from the air. The modeling of the longitudinal profiles of precipitable water in the atmosphere as well as exchangeable moisture in the soil is supplemented by the modeling of the complementary longitudinal profiles of annual mean surface and air temperature. In a climatological experiment, the forest cover in central Amazonia (i.e., in the two regions between about 57.5° and 67.5°W) is assumed to be reduced to one-half of the original value, thereby increasing the evaporative flushing rate of soil moisture in the two partly deforested regions. Computations show that due to recycling the precipitation increases at all longitudes between 57.5° and 77.5°W; however, the primary augmentation of evaporation outweighs the precipitation increase. Consequently, total runoff from the entire basin decreases from the original 1075 mm year⁻¹ to 1018 mm year⁻¹, but the atmospheric throughflow (i.e., precipitable water moving westward across the continental divide) increases by 3%, which could benefit the "dry side" of the Andes cordillera. Soil moisture in the Amazonas basin is found to decrease at all longitudes eastward of the midpoint of the partly deforested area.

1. Dimensions of the system and external conditions

The Amazonas basin has an estimated area of 6.3 Mm²; approximately 4.8 Mm² are in Brazilian territory, the remaining area is shared by Bolivia, Colombia, Ecuador and Peru. The basin is bounded to the west by the Andes Cordillera with elevations of up to 6000 m; to the north by the Guyanas Highlands (with elevations of up to 3000 m) which divide the Amazonas from the Orinoco basin; to the south by the 1200 m high Centro-Oeste Plateau which divides the Amazonas from the Parana River basin; to the east it opens to the Atlantic Ocean. About 75% of the basin is covered with natural forests. Present plans call for deforestation and eventual cultivation of nearly 50% of the Brazilian area.

The main channel of the Amazonas runs almost parallel to the equator, with the eastward streamflow being opposite to the easterly winds prevailing over equatorial South America during the year. Tropospheric streamline fields suggest that the Northern Hemispheric tradewinds are the main source of moisture influx into the basin (see Molion, 1975). Because of these external conditions the

Amazonas system can be looked upon as the prototype of a hydrologic cycle: Precipitation falls out of the moist air as it moves from the continental coastline to the continental divide and the water is then returned as riverflow to the ocean.

What portion of the precipitation in a continental region is formed from advected (external) moisture and what portion from regional evaporation, i.e., from recycled (internal) moisture, has been extensively discussed in the literature [see, for example, Budyko (1974) or Yi-Fu Tuan (1968) or Sellers (1965)]. Hydrologists deemphasize the role of recycling [see, for example, UNESCO (1971)]. Statements can be found in the literature expressing the concept that evaporation from the land is unimportant for continental precipitation and hence such activities as land drainage and reservoir construction have little effect on the precipitation regime of an area (Linsley, 1951). As a first indication that this may not be true for the hydrologic cycle of Amazonia's dimensions, we refer to the streamline maps with superimposed lines of precipitable water constructed by Molion (1975). Based on the average aerological conditions as presented by Newell *et al.* (1972), these charts show that moisture increases as the air travels over land for more than 2000 km.

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Using the methods of climatology (Lettau, 1969; Lettau and Lettau, 1975) to model coherently both the atmospheric and hydrologic phases of the water cycle, this report will show that recycling of vapor is important in Amazonia and has consequences for the assessment of deforestation effects.

2. Atmospheric import

We consider the normal annual means of the following area-averages representative of defined regions of the river basin:

P	precipitation (mm year ⁻¹)
w	precipitable water (mm), i.e., water equivalent of atmospheric total vapor above the soil/air interface
E	evaporation (mm year ⁻¹)
$x - x_1$	distance (km) from the coastline (where $x = x_1$)
U	effective velocity (km year ⁻¹) of air mass motion from coast to continental divide, with $\partial U/\partial x$ = divergence (year ⁻¹) of U
N	effective runoff (mm year ⁻¹) with $N/P = C^* =$ runoff ratio
m	exchangeable soil moisture (mm), i.e., equivalent column of water below the air/soil interface.

In the conventional one-dimensional model of the hydrologic cycle P , E , w , U , N and m are functions of the independent variable x . The two averaged balance equations simultaneously valid for annual means of total moisture above and below the air/soil interface are

$$P - E + \partial(Uw)/\partial x = 0$$

and

$$P - E - N = 0. \quad (1)$$

For the assumed steady-state climatic conditions the direct mathematical consequence of coherence between atmospheric and hydrologic phase is that $\partial(Uw)/\partial x = -N$. In physical terms this necessitates that at any distance from the coast surface and sub-surface runoff (N) must be equal to the atmospheric advection $[-\partial(Uw)/\partial x]$ coming from the opposite direction. This fact shall be used for closure in Section 3.

According to Molion (1975), the defining equation for U is

$$U = w^{-1} \int_{p_0}^{p'} -qu dp/g \quad \text{with} \quad w = \int_{p_0}^{p'} -q dp/g, \quad (2)$$

where the conventional defining equation for w is used, with p the atmospheric pressure (p_0 the surface pressure and p' a constant ≤ 300 mb), g gravity, q specific humidity of the air, and $u = u(p)$ the wind component in the x direction. The inflow or import of precipitable water across the coastline (where

$x = x_1$) is $(Uw)_1 = U_1 w_1$. Above the land regions farther downwind E is the input into the atmosphere (i.e., the forcing) and P the depleting moisture flux. Therefore, we solve the first equation in (1) for E ,

$$E = P + w(\partial U/\partial x) + U(\partial w/\partial x). \quad (3)$$

Because the fluxes (E, P) along x are coupled by U and $\partial U/\partial x$ these two variables shall be referred to as "inter-regional" parameters.

In analogy to the evapoclimatonomical treatment of the hydrologic phase (Lettau, 1969; Lettau and Lettau, 1975) two regional parameters (p_E and r_p) are introduced to divide the depleting process into two parts, i.e.,

$$P = p_E E + r_p w. \quad (4)$$

For any given time increment (here, the average year) the dimensionless p_E serves to express by $p_E E$ the fraction of regional evaporation returned to the regional air/soil interface while the dimensional r_p (year⁻¹) is the flushing frequency of atmospheric moisture. Conventionally, $1/r_p$ is taken as a measure of residence time (in days, or fractions of one year) of water molecules in the average tropospheric column. The regional parameter r_p interacts with the inter-regional parameter $\partial U/\partial x$. The sum $(r_p + \partial U/\partial x)$ when multiplied by dx/U (i.e., the time increment during which the air mass travels from x_1 to $x_1 + dx$) equals the dimensionless trajectory increment $d\xi$; on integration,

$$\xi = \int_{x_1}^x (r_p + \partial U/\partial x) U^{-1} dx. \quad (5)$$

For a visualization of the physical meaning of ξ , reference is made to the discussion of recycling of water in Section 3. To establish the basis for this discussion we define three auxiliary functions of x , denoted by r , W and X . The first is the effective atmospheric flushing frequency r :

$$r = r_p + \partial U/\partial x \text{ (year}^{-1}\text{)}; \quad (6a)$$

the second is W , a reformulation of the effective forcing input, converted to physical units of precipitable water, with the aid of (6a),

$$W = (1 - p_E)E/r \text{ (mm)}; \quad (6b)$$

and the third is X , the nondimensional mathematical expression of atmospheric retention which measures the capacity of the tropospheric column to retain the imported precipitable water; with the aid of (5)

$$X = \exp(-\xi). \quad (6c)$$

Upon expressing P in (3) with the aid of (4) and separating the terms involving E and w with the aid of (5), (6a) and (6b) the balance equation (3) is

reformulated as follows:

$$W(w) = w + \partial w / \partial \xi, \quad (7)$$

i.e., a linear first-order differential equation of the form characteristic of all climatological algorithms. This equation for the effective forcing W as a function of w is rigorously solved by integration with respect to ξ yielding the "climatological transform," i.e., the atmospheric moisture response as a function of W ,

$$w(W) = w_1 X + X \int_{x_1}^x X^{-1} W d\xi, \quad (8)$$

where X is defined in (6c). Of the two constituents on the right-hand side of (8) the first one expresses the gradual depletion of the initial value w_1 as the air mass moves inland while the second expresses replenishment of precipitable water by the evaporative flux of soil moisture in a weighted form as defined in (6b). Note that ξ can become negative for $r < 0$, i.e., for sufficiently strong convergence ($\partial U / \partial x < 0$) of atmospheric transport. The initial input remains unchanged if $r = 0$ but in this special case E must be equal to zero. The second term of (8) expresses the effects of recycling as shall be discussed in Section 3.

In practice, neither inputs nor parameters are prescribed as mathematical functions of the independent variable x . Instead, they are given as discrete step-values that are representative for regional sections, i.e., for finite trajectory increments from x_i to x_{i+1} . Let $j = (i + 1 + i)/2 = i + 1/2$; then the version of the climatological transform (8) most useful for practical computations is based on the regional nondimensional parameter X_j , the "at-

mospheric retentivity" derived from (6c):

$$w_{i+1} = w_i X_j + (1 - X_j) W_j,$$

where

$$X_j = \exp(-\xi_{i+1} + \xi_i). \quad (9a)$$

Most conveniently the regional value of the effective velocity of large-scale air motion U_j is measured by using $x_{i+1} - x_i$ (km) as unit length increment. Then, based on (6b), the atmospheric retentivity X_j of the region from x_i to x_{i+1} is computed using the previously defined regional parameters r_j and U_j :

$$\log_e X_j = -r_j / U_j. \quad (9b)$$

Following Molion (1975), the Amazonas basin is divided into six regions of 5° longitudinal width. Since the geographical latitude is close to zero, $x_{i+1} - x_i$ is taken as 556 km with $i = 1$ at the coastline and $i + 1 = 7$ at the continental divide; consequently j varies from 1 to 6. In Table 1, representative values of E , p_E , r_p , U and $\partial U / \partial x$ are based on Molion's analysis, only slightly smoothed to achieve interregional consistency. Derived variables are also listed, and the computed outputs w_i , w_j , P , N and $C^* = N/P$ are summarized. Note that w_j is obtained with the aid of the balance equation (7) in finite difference form, i.e.,

$$w_j = W_j + (w_{i+1} - w_i) / \log_e X_j. \quad (9c)$$

The initial value is set at $w_1 = 42.0$ mm according to the average precipitable water content of the air near the Atlantic coastline. The documentation of computed w , P and N values in Table 1 shows that w increases inland to a maximum of 45.5 mm near $62.5^\circ W$, whereas precipitation decreases to a weak minimum near $65^\circ W$ followed by an increase

TABLE 1. Evapoclimatology of Amazonia, atmospheric phase. Annual means of (1) regional input (E) and parameters (p_E and r_p); (2) inter-regional parameters (U_j and $\partial U / \partial x$); (3) derived parameters (W_j and X_j); and (4) computed outputs (w_i , w_j , P , N , C^*) for six 5° wide regions centered at indicated longitudes.

	West longitude (deg)					
	50	55	60	65	70	75
(1) E (mm year $^{-1}$)	1440	1368	1284	1116	1200	1264
p_E (-)	0.2407	0.3008	0.3558	0.3781	0.5132	0.7142
r_p (year $^{-1}$)	45.46	41.04	38.09	38.48	40.35	43.93
(2) U_j (556 km year $^{-1}$)	430	392	370	350	328	304
$\partial U / \partial x$ (year $^{-1}$)	-41.14	-29.10	-22.22	-18.31	-22.14	-28.34
(3) W_j (mm)	253.00	80.11	52.10	34.41	32.08	23.17
X_j (-)	0.990	0.970	0.958	0.944	0.946	0.950
(4) w_i (mm)	42.00	44.11	45.19	45.48	44.86	44.17
w_j (mm)	43.06	44.65	45.34	45.17	44.51	43.64
P (mm)	2304	2244	2184	2160	2412	2820
N (mm)	864	876	900	1044	1212	1556
C^* (-)	0.375	0.390	0.412	0.483	0.502	0.552

Note: Whereas w_j is valid at the indicated longitude, w_i [computed from Eq. (9a)] is valid at longitudes 2.5° less than indicated; w_{i+1} for $i = 6$ (which refers to w at $77.5^\circ W$) is 43.12 mm.

TABLE 2. Four nondimensional characteristics related to recycling of atmospheric vapor in six 5° wide regions of Amazonia: $P(r_p w)^{-1}$ = (regional precipitation)/(flushing of regional precipitable water); α = (regional precipitable water)/(precipitable water of oceanic origin); $\gamma = \alpha P(r_p w)^{-1}$ = (regional precipitation)/(flushing of precipitable water of oceanic origin); β = Budyko's water exchange coefficient [Eq. (11c)].

	West longitude (deg)					
	50	55	60	65	70	75
$P(r_p w)^{-1}$	1.177	1.225	1.265	1.243	1.343	1.471
α	1.010	1.041	1.087	1.151	1.217	1.281
γ	1.189	1.275	1.375	1.431	1.634	1.884
β	1.039	1.078	1.116	1.152	1.193	1.239

towards the mountainous source region. The runoff ratio shows a continuous increase with longitude.

The computed results are derived from parameter values which are consistent with Molion's analysis of aerological conditions based on Newell *et al.* (1972). The effective velocity U_j of atmospheric precipitable water transport decreases steadily from the coastline toward the continental divide. The resulting convergence ($\partial U/\partial x < 0$) is expressed in the same units (year^{-1}) as r_p . A regional value of $r_p = 36.5 \text{ year}^{-1}$, for example, means that the average "residence-time" of a water molecule in the air amounts to 10 days. Correspondingly, for the velocity units used in Table 1, an U_j value larger (or smaller) than 365 means that the average moisture-carrying tropospheric column traverses the regions of 5° width (556 km at the equator) in less (or more) than one full day. According to (7) the sign of the difference $W - w$ determines whether w increases or decreases along the effective trajectory (5). Since $31.5 \text{ year}^{-1} = 10^{-6} \text{ s}^{-1}$, the magnitude of convergences ($\partial U/\partial x < 0$) in Table 1 implies average updrafts amounting to about 5 mm s^{-1} near 5 km above the ground, for example.

3. Recycling of water

Climatonomical modeling produces directly the recycled part $p_E E$ of the regional evaporation. Thus, precipitation derived from advected precipitable water (i.e., vapor formed by evaporation anywhere upwind of the given region) is $P - p_E E = r_p w$, according to (4). Values of $P(r_p w)^{-1}$ computed from Table 1 are summarized in Table 2.

In the climatonomical transform (8) the term $w_i X$ gives exactly the downwind decay of the initial input w_1 . Of interest is the ratio (α_j) of precipitable water present in the j th region (from x_i to x_{i+1}) to that of direct oceanic origin. To compute this ratio w_i in (9a) is expressed by w_{i-1} and this in turn, by w_{i-2} , etc., until w_1 is reached. Then, the equation for w_{i+1} shows on the right-hand side the multiple product $w_1 X_1 X_2 \cdots X_j$ and a remainder which de-

pends exclusively on recycled water. Since the X_j are directly obtained with the aid of (9b),

$$\log_e \alpha_j = - \sum_{j=1}^j \log_e X_j. \quad (10a)$$

With (5) and (6c) $\alpha = \log_e \xi$ which permits visualization of ξ in physical terms. The exact expression for the ratio of regional precipitation to flushing of precipitable water of direct oceanic origin thus is

$$\gamma = \alpha P(r_p w)^{-1}. \quad (10b)$$

The distribution of the characteristic ratios α and γ along the Amazonas system are summarized in Table 2. Recycling of atmospheric vapor is quantitatively evidenced by the downwind increase of $\gamma - 1$.

Budyko (1974, pp. 240–242) introduced the "water-exchange coefficient" β in his analysis of the principal factors affecting water vapor transfer above an area of downwind length L ,

$$\beta = 1 + EL/(2wu). \quad (11a)$$

Note that each of the 5° wide regions in Table 1 contributes 556 km to the length L and that the listed U_j data are equivalent to u over increments of 556 km.

Computation of β for the six regions with the aid of data from Table 1 requires cumulative length-averages such as

$$\bar{E}_j = j^{-1} \sum_{j=1}^j E_j, \quad (11b)$$

and the corresponding cumulative length-averages for $(\bar{U}_w)_j$. Then

$$\beta_j = 1 + j\bar{E}_j/[2(\bar{U}_w)_j]. \quad (11c)$$

All four nondimensional characteristics related to recycling of atmospheric vapor over the Amazonas system are summarized in Table 2.

For the large L values of Amazonia the coefficient β exceeds westward of 60° longitude the yearly average of 1.12 obtained for eastern Europe and Russia by Budyko (1974, p. 242). In comparison with $\gamma - 1$, however, the actual role of recycling appears to be grossly underestimated by $\beta - 1$. The following explanation is offered. Budyko has argued that strong mixing in the troposphere occurs and henceforth he has assumed that the relationship between the sums of precipitation formed from local versus external water vapor is equal to that between the corresponding quantities of vapor molecules present in the atmosphere. The data in Table 2 do not support this assumption. In Amazonia the coefficient β is sufficiently in agreement with the coefficient α except for the regions east of 65°W where w increases along the inland trajectory.

Thus, β is at best a measure of regional precipitable water to that of oceanic origin.

The physical reason for the invalidity of Budyko's assumptions is "fast-recycling". This refers to local showers developing predominantly in the course of the diurnal insolation cycle and yielding rain before all cloud water is mixed laterally or vertically with the total precipitable water in the average tropospheric column above the region. According to Molion (Figs. 4.1 and 4.2) fast-recycling contributes about 30% to the total precipitation in Amazonia, which is consistent with the average $(P/\Gamma_p w)_6 = 1.287$ in Table 2. At 75°W , $\gamma = 1.884$ means that 88.4% of the total amount of rain falls at least a second time as the result of recycling of vapor evaporated from the land regions.

4. Surface and subsurface export on water in conjunction with the atmospheric phase of the hydrologic cycle

Exchangeable soil moisture (m) corresponds to atmospheric precipitable water (w). To establish perfect analogy to the climatonomical treatment of the atmospheric phase, hydrologic runoff (N) is expressed as $N = \partial(Vm)/\partial x$, where V is the virtual velocity of surface and subsurface water transport from source to sink corresponding to the speed of effective atmospheric motion from sink to river source.

To visualize the meaning of the velocity component V it is helpful to compare its magnitude with that of U after converting from the practical units used in Tables 1 and 3 to a more conventional velocity denomination, noting that $556 \text{ km year}^{-1} = 0.0176 \text{ m s}^{-1}$. If we let the six-region averages be denoted by the overbar, it follows from Table 1 that

$\bar{U} = 5.5 \text{ m s}^{-1}$, while Table 3 yields $\bar{V} = 0.14 \text{ m s}^{-1}$. The order of magnitude of the former is that of the easterly component of tropospheric winds and of the latter that of westerly component of average water flow in the river channels of the Amazonas system. Naturally, the actual movement of soil moisture m over the basin's huge surface area of $A = 6.3 \times 10^6 \text{ km}^2$ is highly complex. It includes downward percolation of $P - E$, seepage and overland runoff to creeks and rivers, and only ultimately the organized streamflow in tributaries and the main channel eastward to the ocean. By its definition, only the effective horizontal divergence $\partial(Vm)/\partial x$ of the dominating oceanward component of water movement is directly given by the hydrologic runoff N . This is confirmed by a comparison of magnitudes; namely, according to Table 1, $\bar{N} = \bar{P} - \bar{E} = 1075.3 \text{ mm year}^{-1} = 0.0341 \times 10^{-6} \text{ m s}^{-1}$, less than one millionth of $\bar{V}$.

Although the divergence of effective soil moisture flow is small it accumulates over the entire basin to the enormous total discharge into the Atlantic Ocean of $A\bar{N} = 215\,000 \text{ m}^3 \text{ s}^{-1}$ when the above data are used. The divergence terminates to contribute to Vm at the mouth of the river. Here, at $x = x_1$, the effective cross sectional area (a) of the Amazonas outlet has the order of magnitude of 1 km^2 , i.e., an effective width of about 10 km combined with an effective depth of about 0.1 km. The V value at the estuary has a direct physical meaning insofar as it becomes equal to $\bar{N}A/a$. With the estimated a value and the above $\bar{N}A$ it follows that $V = 0.215 \text{ m s}^{-1}$, in fair agreement with the V value of 0.24 m s^{-1} which derives from Table 3 for 50°W . From another viewpoint, the physical meaning of mV for any x value is discussed in connection with Eq. (15b).

TABLE 3. Evapoclimatology of Amazonia, hydrologic phase. Annual means of (1) regional input (P) and parameters (e_p and r_E); (2) inter-regional parameters (V and $\partial V/\partial x$); (3) derived parameters (M and Y); and (4) computed outputs (m_i , m_j , E , Δt) for six 5° wide regions centered at indicated longitudes.

	West longitude (deg)					
	75	70	65	60	55	50
(1) P (mm year $^{-1}$)	2820	2412	2160	2184	2244	2304
e_p (-)	0.4052	0.4258	0.4118	0.4724	0.5005	0.4756
r_E (year $^{-1}$)	0.200	0.286	0.384	0.440	0.440	0.636
(2) V (556 km year $^{-1}$)	2.913	5.126	7.304	9.203	10.783	12.959
$\partial V/\partial x$ (year $^{-1}$)	2.510	2.113	1.972	1.848	1.885	1.945
(3) M_j (mm)	619.0	577.3	539.2	503.7	482.0	468.1
Y_j (-)	0.3944	0.6262	0.7243	0.7799	0.8060	0.8194
(4) m_i (mm)	600.0	611.5	598.7	582.3	565.0	548.9
m_j (mm)	606.6	604.6	590.1	573.3	556.7	541.4
E (mm year $^{-1}$)	1264	1200	1116	1284	1368	1440
Δt (months)	4.43	5.00	5.09	5.24	5.16	4.65

Note: Whereas m_j is valid at the indicated longitudes, m_i [computed from Eq. (15a)] is valid at longitudes 2.5° larger than indicated; m_{i+1} for $i = 6$ (which here refers to m at 47.5°W) is 534.3 mm.

Annual averages as in (1) satisfy the balance

$$P = E + \partial(Vm)/\partial x = E + m(\partial V/\partial x) + V\partial m/\partial x. \quad (12a)$$

Exchangeable soil moisture (m) is replenished by precipitation and depleted by evaporation. As in Section 2, the depleting process is divided into two parts with the aid of two regional parameters (e_p and r_E), i.e.,

$$E = e_p P + r_E m, \quad (12b)$$

where e_p (dimensionless) measures the fraction of local precipitation returned to the atmosphere, and r_E (year^{-1}) is the evaporative flushing frequency of moisture in the average soil column of the region. The local parameter r_E and the two inter-regional parameters V and $\partial V/\partial x$ serve to derive three auxiliary quantities, in analogy to Eqs. (6a)–(6c): first, the effective soil flushing frequency (r)

$$r = r_E + \partial V/\partial x [\text{year}^{-1}], \quad (13a)$$

second, the soil moisture equivalent of the effective forcing input (M)

$$M = (1 - e_p)P/r [\text{mm}], \quad (13b)$$

and third, the nondimensional moisture retentivity (Y) of the soil

$$Y = \exp(-r/V). \quad (13c)$$

Since $V = 0$ at the continental divide and increases monotonically downstream, the arrangement of the six 5° wide sections is reversed from that of the previous section. The divergence ($\partial V/\partial x$) shows a weak minimum between 60° and 55°W . Computations of m_i and m_j proceed in perfect analogy to the w_i and w_j computations in the atmospheric phase. Therefore, writing down the equivalent of (7) and (8) may be omitted and only the hydrologic counterpart of (9a and b) may be given:

$$m_{i+1} = m_i Y_j + (1 - Y_j) M_j, \quad (14a)$$

$$m_j = M_j + (m_{i+1} - m_i)/\log_e Y_j. \quad (14b)$$

Results are shown in Table 3, where P is input and E is output, contrary to Table 1 where E is input and P is output.

Regional and inter-regional parameters (e_p , r_E , V , $\partial V/\partial x$) are again based on Molion's analysis. However, two important new considerations are necessary. First, we found that the evaporative flushing frequency of soil moisture varies with the fraction f of forest cover of the regions, i.e.,

$$r_E = (1.25 - f)/(0.55 + f) [\text{year}^{-1}], \quad (15a)$$

yielding, for example, $r_E = 0.714$ for $f = 0.5$ and $r_E = 0.161$ for $f = 1.0$. According to Molion, $f = 0.55, 0.70, 0.70, 0.75, 0.85$ and 0.95 for the six 5° wide regions from 50° to 75°W . Second, exact

coherence between the hydrologic and atmospheric phase is achieved at any $x - x_1$ between source and sink by fulfilling the coupling balance requirement that

$$Uw - Vm = \text{constant}. \quad (15b)$$

For the Amazonas system the value of the constant in (15b) was found to be 11 500 in the units used for U or V , and w or m in Tables 1 and 3, equivalent to a H_2O transport of 191 kg s^{-1} per crosswind length unit of 1 m. Because $V = 0$ at the continental divide this constant is the atmospheric "through flow". Since m is significantly larger than w , the values of V in Table 3 must be given with higher numerical accuracy than those of U in Table 1.

The initial value of soil moisture (at the continental divide) was set at $m_1 = 600 \text{ mm}$. Computations with the aid of the derived parameters (M_j and Y_j) yield m_i and m_j , also E_j from source to sink regions. The soil moisture reaches a maximum near 72.5°W , continuously decreasing further eastward to the coastline value of 534.3 mm. These m values are very similar to those derived for wet years in the Philippine Islands (Lettau and Baradas, 1973).

Effective residence-time of moisture in the soil [$\Delta t = 12/(r_E + \partial V/\partial x)$, see Table 3] is of the order of 5.0 months, substantially exceeding the regional residence-time (about 10 days) of vapor in the troposphere. These Δt values agree within error tolerance with those derived by Molion who analyzed each of the six regions separately using month-to-month variations of mean monthly P , E and N values. In the Philippine Islands study Δt varied between 3 and 5 months as a function of m (Lettau and Baradas, 1973).

In the climatonomical theory of the hydrologic cycle the full coherence between atmospheric and hydrologic phases is used for mathematical closure. The algorithm is such that the two functions $P(x)$ and $E(x)$ are the same regardless whether $P(x)$ is input and $E(x)$ computed (as for x increasing downstream in the hydrologic phase), or $E(x)$ is input and $P(x)$ computed (as for x increasing upstream in the atmospheric phase).

In a case study such as this, one may accept the parameter set but find fault with one, several or all forcing inputs. If this is so, one should correct or improve the inputs and recompute new outputs with the aid of the climatonomical transforms for both phases. If one accepts the forcing inputs but finds fault with the parameters, then these should be corrected and all outputs be recomputed. However, all modifications of inputs or parameters must be applied coherently. Meteorologists and hydrologists may have different opinions on what represents an improvement. The criterion of what is the best approach can be based on the annual mean surface temperatures and energy budgets of the regions.

Climatonomical methods for such final closure will be discussed in the next section.

5. Climatonomical closure by computation of annual mean surface and air temperatures

Consider the normal annual means of area-averages of the following variables representative of defined regions of the river basin:

- G shortwave flux (W m^{-2}) of global radiation from sun and sky intercepted by the atmosphere's lower boundary
- a^* surface albedo ($0 < a^* < 1$) so that $(1 - a^*)G$ is the surface-absorbed global radiation
- T temperature ($^{\circ}\text{C}$) of the active surface
- ϵ^* surface emissivity ($0 < \epsilon^* < 1$) for longwave flux of radiation
- σ Stefan-Boltzmann constant, used here as $\sigma^{-1/4} = 64.8^{\circ}\text{K} (\text{W m}^{-2})^{-1/4}$ so that longwave radiation LWU (W m^{-2} , upwelling from the surface) $= \epsilon^*[(T + 273.15)/64.8]^4$
- A^* Angström ratio (flux of effective longwave radiation)/(LWU) so that $(1 - A^*)$ LWU is returned to the surface by atmospheric counter-radiation
- B^* Bowen ratio (flux of sensible heat into air)/(flux of latent heat into air)
- λ latent heat of vaporization, used here as $0.0771 (\text{W m}^{-2})/(\text{mm year}^{-1})$ (a quasi-constant for tropical temperature levels)
- θ air temperature ($^{\circ}\text{C}$) at standard height (1.8–2.0 m above ground)
- h_q coefficient for relating θ to T , used here as $h_q = (T - \theta)/B^*\lambda E$, where $B^*\lambda E$ is the convective flux of sensible heat to the air

Note: The asterisks are used to denote ratios of annual means, which may not be equal to annual means of ratios.

In tropical climates the annual range of surface temperature is sufficiently small so that averages of $(T + 273.15)^4$ differ insignificantly from the fourth power of the averaged surface temperature. On continental areas storage of sensible heat in the soil has a zero annual mean. Thus, the annual energy budget at the soil/air interface is determined solely by absorbed global radiation, effectively emitted longwave radiation, and atmospheric convective fluxes of both latent and sensible heat:

$$(1 - a^*)G - \sigma\epsilon^*A^*(T + 273.15)^4 - (1 + B^*)\lambda E = 0. \quad (16)$$

The sum of the first two terms is the net radiation (R). Recalling from the soil moisture budget that $E = (1 - C^*)P$, the equations for computing the surface temperature and, subsequently, air temperature at climate-station level are

$$T = -273.15 + 64.80 \times \{[G(1 - a^*) - 0.0771P(1 + B^*) \times (1 - C^*)]/A^*\epsilon^*\}^{1/4} \quad (17a)$$

$$\theta = T - 0.0771PB^*(1 - C^*)h_q. \quad (17b)$$

The two forcing inputs [solar energy G (W m^{-2}) and mass P of H_2O (mm year^{-1})] are qualified by the five dimensionless climatic parameters ($a^*, A^*, B^*, C^*, \epsilon^*$) and the two physical parameters (the universal constant $\sigma = 64.80^{\circ}\text{K} (\text{W m}^{-2})^{-1/4}$ and the slightly temperature-dependent λ approximated here by $0.0771 \text{W m}^{-2} \text{ year mm}^{-1}$).

Forcing inputs and parameter sets for the six 5° wide regions of Amazonia are given in Table 4 together with outputs. Computed air temperatures agree within $\pm 1/2^{\circ}\text{C}$ with annual area averages estimated from mean isotherm charts for January and July published by Ratisbona (1976, pp. 234, 235).

TABLE 4. Thermoclimatology of Amazonia. Annual means of (1) regional inputs of energy (global radiation G and energy equivalent of precipitation λP); (2) regional parameters ($a^*, A^*, B^*, C^*, \epsilon^*, h_q$); and (3) computed outputs (surface temperature T , air temperature θ and dryness ratio D^*) for six 5° wide regions centered at indicated longitudes.

	West longitude (deg)					
	50	55	60	65	70	75
(1) G (W m^{-2})	214.7	219.3	217.7	221.0	215.7	211.5
λP (W m^{-2})	177.6	173.0	168.4	166.5	186.0	217.4
(2) a^* (—)	0.1450	0.1300	0.1300	0.1250	0.1150	0.1050
A^* (—)	0.0981	0.1014	0.1055	0.1141	0.1097	0.1065
B^* (—)	0.2793	0.3460	0.4343	0.6512	0.5351	0.4579
C^* (—)	0.3750	0.3904	0.4121	0.4833	0.5025	0.5518
ϵ^* (—)	0.9692	0.9728	0.9728	0.9740	0.9764	0.9788
h_q ($^{\circ}\text{C W}^{-1} \text{ m}^2$)	0.0450	0.0332	0.0332	0.0300	0.0245	0.0200
(3) T ($^{\circ}\text{C}$)	27.6	27.2	27.2	27.3	26.4	25.9
θ ($^{\circ}\text{C}$)	26.2	26.0	25.8	25.6	25.2	25.0
D^* (—)	0.800	0.821	0.843	0.853	0.763	0.653

Partial closure is achieved by relating empirically three of the five climatic parameters to regional forest cover f :

$$a^* = 0.200 - 0.100f, \quad (18a)$$

$$\epsilon^* = 0.956 + 0.024f, \quad (18b)$$

$$h_q = 0.030(1.75 - f)/(0.25 + f). \quad (18c)$$

The physical justification for these specialized expressions is the relative uniformity of the tropical rainforest structure and the known effects of forests on albedo and emissivity [see, for example, Sellers (1965)]. Relationship (18c) approximates increased aerodynamic surface roughness for denser forest cover.

The Ångström ratio is known to vary with cloud cover, vapor pressure and atmospheric turbidity, the Bowen ratio with surface temperature and soil moisture. Evaporation likewise varies with cloud cover, atmospheric humidity and soil moisture. Therefore, it is not surprising that for the relatively moderate range of surface temperatures the values of A^* and B^* resulting from Molion's analysis turned out to be negatively correlated to E as determined from the moisture budget. A tentative closure is achieved by the following two specialized formulas which are consistent with limiting conditions for both ratios. On the one hand, for a hot and humid climate with unlimited water supply, evaporation is large which makes A^* small and B^* finite but negative. On the other hand, for a hot and dry climate, A^* is relatively large and B^* approaches infinity. With reference values E_A and E_B [mm year⁻¹ (as E in Tables 1 or 3 for Amazonia)]:

$$A^* = A_0 \exp(-E/E_A);$$

$$\text{specifically, } A^* = 0.1915 \exp(-E/2153), \quad (19a)$$

$$B^* = B_\infty + E_B/E;$$

$$\text{specifically, } B^* = -1.0 + 1842/E. \quad (19b)$$

The computation of realistic surface and air temperatures for the six regions with the aid of (17a)–(19b) required slight adjustments (within probable error limits) of the G values suggested by Molion (1975). Results are listed in Table 4.

The sensitivity of temperature response to parameter variation can be established and tested by modifying one parameter after the other. For example, a change of ± 0.0010 in the Ångström ratio produces a response of $\pm 0.7^\circ\text{C}$ in surface temperature for the average conditions in Amazonia. The same effect in T (all other conditions remaining unchanged) corresponds to ± 9 mm year⁻¹ in P , or $\pm 0.5\text{ W m}^{-2}$ in G . With due consideration of the uncertainties in absolute T computation, the results summarized in Table 4 show that (15a) and the

highly specialized parameterization [(18a)–(19b)] are satisfactorily reproducing the observed variations from region to region. We conclude that the same parameterization can be used for the relative assessment of thermal deforestation effects in a single region (see Section 6 and Table 6).

The ratio C^* in Table 4 is derived from N/P in Table 1. Use of (19b) in the universally valid relationship for continental climates (Lettau, 1969),

$$(1 + B^*)(1 - C^*) = D^*, \quad (20)$$

yields the "radiational index of dryness" D^* (Budyko, 1974; in reference to work published nearly two decades earlier). D^* values for Amazonia are included in Table 4. Note that the specialized formula (19b) results in net radiation being independent of annual mean evaporation and vice versa.

6. Demonstration of recycling effects by climatonomical experimentation simulating partial deforestation

It appears possible that in the future a considerable part of the Amazon forest will be cleared. To demonstrate inter-regional-coupling and to clarify the role of recycling, let us assume that the forest cover is reduced only in the two centrally located regions of the six considered in Tables 1, 3 and 4. For the numerical assessment the two respective f fractions will be arbitrarily reduced to half of their original values. Directly affected by this change are soil moisture-flushing frequency, surface albedo and emissivity, also the h_q coefficient [see Eqs. (15a) and (18a)–(18c)]. Other parameters of the moisture budget will remain unchanged, but the resulting modification in E will affect the Ångström as well as Bowen ratios according to (19a) and (19b).

Constancy of inter-regional parameters implies that neither velocity nor its divergence of both the airflow in the lower atmosphere and the waterflow in soil and river channels are modified. However, the through-flow constant (15b) will have a different value. There is no physical cause for modifying the initial value of precipitable water above the Atlantic coastline. At the Continental Divide the initial value of exchangeable soil moisture is changed tentatively in proportion to the computed atmospheric precipitable water.

Using Eqs. (14a,b) and (12b), computations begin with P from Table 3 and the modified r'_E , yielding new E values from 65° through 50°W longitude. These new E values are used in Eqs. (9a,c) and (4) to generate a set of new P values from 50° through 75°W longitude. The computations are repeated with the new P values to obtain newer E values for

TABLE 5. Numerical modeling of evapoclimatonomical effects resulting from forest cover reduction in central Amazonia. (1) input: fractions of forest cover f' and flushing frequency of soil moisture r'_E (also differences against original values of f and r_E); (2) output: computed new equilibrium values P' , E' , $C^{*'}$, w' and m' (also differences against original values from Tables 1 and 3).

	West longitude (deg)					
	50	55	60	65	70	75
(1) f' (-)	0.55	0.70	0.35	0.375	0.85	0.95
$f' - f$	0	0	-0.35	-0.375	0	0
r'_E (-)	0.636	0.440	1.000	0.946	0.286	0.200
$r'_E - r_E$	0	0	+0.560	+0.562	0	0
(2) P' (mm year ⁻¹)	2299	2237	2322	2350	2478	2879
$P' - P$	-5	-7	+138	+190	+66	+59
E' (mm year ⁻¹)	1422	1351	1646	1526	1233	1298
$E' - E$	-18	-17	+362	+410	+33	+34
$C^{*'} (-)$	0.381	0.396	0.291	0.351	0.502	0.551
$C^{*'} - C^*$	+0.006	+0.006	-0.121	-0.132	-0.008	-0.007
w' (mm)	43.04	44.61	45.59	46.07	45.74	44.84
$w' - w$	-0.01	-0.04	+0.25	+0.90	+1.23	+1.20
m' (mm)	516.5	526.4	548.9	590.0	620.9	623.0
$m' - m$	-24.9	-30.3	-24.4	-0.1	+16.3	+16.4

all six regions. In the manner of a true iteration these computational steps are repeated until the P used to generate E agrees with the P generated by the E . The final series of computed regional data (P' , E' , w' , m') are shown in Table 5. A total of ten iterative computations were necessary to produce convergence within tolerable error limits. For convenience, the differences against the original data are included.

The major effect of the higher flushing rate at the two central regions is the increase in evaporation, amounting to 28% at 60°W and 37% at 65°W, and still 3% farther west. The importance of recycling is documented by the fact that the precipitable water increase culminates farther downwind than in the control case. This explains high precipitation in the two westernmost regions. Here, soil moisture is also high, but reduced in all regions east of 65°W. Interestingly, consistent with the significant lowering of the runoff ratios in the two central regions, the through-flow constant (15b) is $11\,816 \pm 2$ which is 3% higher than the original value of 11 500. This increase means that west of the Continental Divide, above the Altoplano, the troposphere would receive 3% more moisture from the Atlantic than at present. The total runoff $\bar{N}$ from the Amazonas basin equals $1017.8 \text{ mm year}^{-1}$ in comparison with $1075.3 \text{ mm year}^{-1}$ when r_E is not increased in the two central sections. According to Miller (1977, p. 510) an independent recent estimate of this "yield" of the Amazonas system is $1060 \text{ mm year}^{-1}$, which agrees within 1.4% with $\bar{N}$ according to Tables 1 and 3.

The effects of partial deforestation on Amazonia's hydrologic cycle, summarized in Table 5, were derived by modifying no parameter other than the

evaporative flushing frequency r_E which is directly related to f by (15a). This f dependency of r_E was originally established for simulating the longitudinal profiles of precipitation and evaporation under present conditions. The effects of partial deforestation on surface and air temperatures, summarized in Table 6, resulted from correspondingly established f dependencies [Eqs. (18a)–(18c)] of albedo, emissivity and height-reduction coefficient. In addition, the E dependencies of the Ångström ratio and the Bowen ratio [Eqs. (19a), (19b)] were used in combination with the modified evaporation generated by (15a) as presented in Table 5. Like (15a) the forms (18a,b,c) were originally designed and employed to simulate the present longitudinal profiles of surface and air temperatures. Consequently, the results summarized in Tables 1, 3 and 4 represent the control run for the numerical simulation of climatic effects of deforestation shown on Tables 5 and 6.

Under present conditions forest cover in the study basin increases systematically from $f = 0.55$ around 50°W to $f = 0.95$ around 75°W, while the longitudinal profiles of temperature as well as of evaporation show highest values around 50°W and a general decrease toward the west. It should be emphasized that the physical reason for the documented increase of precipitable water for about the first 2000 km of inland trajectory (see Section 1) was the relatively high evaporation rate as well as the high temperature in the most easterly regions. With this background of model calibration in mind, it is not surprising that the numerical simulation of deforestation in the two central sections (see Tables 5 and 6) produces relative warming and an increase of evaporation which is carried over to the most

TABLE 6. Results of thermoclimatonomical modeling of forest cover reduction in central Amazonia: (1) regional (direct) effects related to f dependency of parameters a^* , ϵ^* and h_q according to (18a,b,c) and on absorbed insolation $F = (1 - a^*)G$ assuming $G' = G$ from Table 4; (2) interregional effects due to coupling by recycling of precipitable water [yielding $\lambda E'$ (W m^{-2}) from Table 5] related to E dependency of A^* and B^* according to (19a,b); (3) modified temperature of surface and air according to (17a,b). Primes denote modified values; the differences against original values are also listed.

	West longitude (deg)					
	50	55	60	65	70	75
(1) $a^{*'}_i$	0.1450	0.1300	0.1650	0.1625	0.1150	0.1050
$\bar{a}^{*'} - a^*$	0.0000	0.0000	+0.0350	+0.0375	0.0000	0.0000
$\epsilon^{*'}_i$	0.9692	0.9728	0.9644	0.9650	0.9764	0.9788
$\bar{\epsilon}^{*'} - \epsilon^*$	0.0000	0.0000	-0.0084	-0.0090	0.0000	0.0000
h_q^i	0.0450	0.0332	0.0700	0.0660	0.0245	0.0200
$\bar{h}_q - h_q$	0.0000	0.0000	+0.0368	+0.0360	0.0000	0.0000
$F' = (1 - a^{*'})G$	186.1	187.5	181.8	185.1	190.9	189.3
$F' - F$	0.0	0.0	-7.6	-8.3	0.0	0.0
(2) $\lambda E'$	109.6	104.2	126.9	117.6	95.1	100.1
$\lambda E' - \lambda E$	-1.4	-1.3	+27.9	+31.6	+2.6	+2.7
$A^{*'}_i$	0.0990	0.1022	0.0892	0.0943	0.1080	0.1048
$\bar{A}^{*'} - A^*$	+0.0009	+0.0008	-0.0163	-0.0198	-0.0017	-0.0017
$B^{*'}_i$	0.2956	0.3628	0.1190	0.2075	0.4932	0.4186
$\bar{B}^{*'} - B^*$	0.0163	0.0168	-0.3153	-0.4437	-0.0419	-0.0393
(3) T' ($^{\circ}\text{C}$)	26.9	26.6	27.3	29.1	27.5	27.1
$T' - T$	-0.7	-0.6	+0.1	+1.8	+1.1	+1.2
θ' ($^{\circ}\text{C}$)	25.4	25.3	26.2	27.5	26.4	26.3
$\theta' - \theta$	-0.8	-0.7	+0.4	+1.9	+1.2	+1.3

westerly regions by inter-regional coupling and recycling processes. With unchanged global radiation the higher albedo accompanying forest cover reduction causes a decrease of solar radiation absorption, but this energy loss is overcompensated by a decrease in the Ångström ratio. In summary, the acceleration of soil moisture flushing rates r_E by partial removal of the thick tropical forest cover causes an increase in temperature and evaporation and, subsequently, of precipitable water or moisture in the air. This augments the greenhouse effect and overrules the reduced absorption of solar energy. Probably, the resulting warming of soil and air serves to maintain higher evaporation because precipitation is also increased due to effective recycling of water.

7. Discussion

Recycling, processing and transforming of water by any ecosystem at the earth-air interface are intricately interwoven (Miller, 1977). When modeling these phenomena, the results must be viewed only as indicators of possible climatic responses and never as absolute truth. A more or less crude approximation of inherently complicated and interwoven processes is unavoidable. The authors of this study are atmospheric scientists and their treatment of soil and plant aspects must appear oversimplified to specialists in these fields. But any

numerical value presented in Tables 1–6 is derived rigorously with the aid of an algorithm developed from basic conservation equations. All steps of the development are outlined, and the complete set of relationships for parameter evaluations is given. Although the required forward integrations of the climatonomical transform can readily be programmed for machine operation, the numerical results summarized in Tables 1–6 were obtained with the aid of a hand computer. With all basic information fully documented (including initial values for the integrations) anyone can check our computations or may derive new results if other input values or parameter choices appear preferable. For example, improved drainage conditions on fields after deforestation can be modeled by enhancement of $\partial V/\partial x$ in (13a).

Limiting the model treatment to a definite large-scale portion of the earth's surface, specifically to a uniform river basin in a uniform climate, has advantages with respect to testing of parameterizations as well as computed results. Similar tests are difficult when the climatic response to tropical rainforest removal is investigated with the aid of worldwide atmospheric models. An example is the numerical experiment by Potter *et al.* (1975).

Model characteristics and main results for the tropics derived by Potter *et al.* are summarized in Table 7. Note that the computed changes in precipitation, precipitable water and temperature are all

TABLE 7. Characteristics and main results of a two-dimensional (zonal) atmospheric model (Potter *et al.*, 1975) in comparison with the present study (climatology) concerning climatic responses to albedo increase in the tropical rain forest zone.

	Potter <i>et al.</i> (1975)	Climatology
Area coverage	Entire globe, 510 Mm ²	Amazonia, 6.3 Mm ²
Independent variables	Height and geographic latitude	Distance from Atlantic Coast
Horizontal resolution	10° latitude	5° longitude
Albedo modification	0.07 changed to 0.25	0.13 changed to 0.16
Area of albedo change	About 15 Mm ² of land between 5°N and 5°S latitude	About 2 Mm ² between 57°W and 68°W longitude
Atmospheric tropical circulation	Weakened Hadley cell as model output	Regional tradewinds assumed unchanged
Precipitation change	-230 mm year ⁻¹ between 5°N and 5°S latitude	+75 mm year ⁻¹ average for Amazonia
Precipitable water	-0.74 mm globally	+0.59 mm over Amazonia
Air temperature near the surface	-0.4°C between 5°N and 5°S latitude	+0.55°C in Amazonia

negative, i.e., the opposite of what results from climatology for the assumptions outlined in Section 6. Note also that in the global model the assumed albedo change is rather extreme since 0.25 is typical of deserts.

Without the machine program it is not possible to establish for the global model the climatic response for a more modest albedo change, since Potter *et al.* provide only verbal descriptions of algorithm and parameterization of the hydrological processes.

Although the set of climatological equations is complete, the empirical parameterization [Eqs. (15a), (18a)–(18c)] is very likely valid only for the restricted forest cover range ($0.55 \leq f \leq 0.95$) of the control run. Furthermore, the assumption that incoming shortwave radiation (G) remains the same in spite of significant effects on precipitable water and precipitation can hardly be justified for more radical modifications. According to (17a) a decrease in G combined with an increase in albedo may not be compensated anymore by a reduction of the Ångström ratio. Lower temperature combined with less solar energy available will increase the soil moisture retention time. Ultimately, with further depletion of forest cover and albedo increase the climatology modeling may approach the conditions predicted by the global model.

Experiences in temperate climates [as, for example, summarized by Baumgartner (1970)] suggest that when forests are changed to field crops the evaporation decreases and soil moisture recycling appears to be slowed down. However, none of the previous studies have been conducted in climates with energy inputs of tropical intensity, more than 2000 mm of annual precipitation, and on horizontal scales of several square megameters, large enough

to cause significant contributions by recycling atmospheric water.

It may be concluded that a modest rate of deforestation in Amazonia can cause a slightly warmer and moister climate with drier soil, while after a more radical deforestation the climate may turn cooler and generally drier than at present. For more detailed modeling it appears necessary to have available improved basic information on the physical factors of the present climate, especially multiannual means from solar and terrestrial radiation recordings, and measurements of atmospheric transports and exchanges of water vapor from a maintained network of aerological stations.

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