POSSIBLE CLIMATIC IMPACTS OF TROPICAL DEFORESTATION

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Abstract. Large-scale conversion of tropical forests into pastures or annual crops will likely lead to changes in the local microclimate of those regions. Larger diurnal fluctuations of surface temperature and humidity deficit, increased surface runoff during rainy periods and decreased runoff during the dry season, and decreased soil moistrue are to be expected.

It is likely that evapotranspiration will be reduced because of less available radiative energy at the canopy level since grass presents a higher albedo than forests, also because of the reduced availability of soil moisture at the rooting zone primarily during the dry season. Recent results from general circulation model (GCM) simulations of Amazonian deforestation seem to suggest that the equilibrium climate for a grassy vegetation in Amazonia would be one in which regional precipitation would be significantly reduced.

Global climate changes probably will occur if there is a marked change in rainfall patterns in tropical forest regions as a result of deforestation. Besides that, biomass burning of tropical forests is likely adding CO_2 into the atmosphere, thus contributing to the enhanced greenhouse warming.

1. Introduction

The conversion of tropical rainforests into a different vegetation, most notably pastures or annual crops, inevitably entails major changes in the ecosystem. There appears to be a consensus that this type of conversion changes the flora, the aquatic and land fauna, and the physico-chemical and biological characteristics of the soil and surface waters. There also appears to be agreement that qualitative and quantitative changes are caused in the biogeochemical cycles.

Yet, when analyzing the climatic impacts involved or associated with the conversion of tropical rainforests into pastures, the question becomes controversial. This is due to several factors, including in particular the difficulty of quantifying the components of the energy and water balances in the undisturbed and disturbed ecosystems, and the difficulty of developing climate models at the regional level that will permit reliable predictions based on the changes in land-use patterns to be introduced. Moreover, it has been generally accepted that the flora is a consequence of the climatic conditions in conjunction with the characteristics of soil and geomorphology. Therefore, it has been assumed that conversion of one type of

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vegetation cover to another should not induce large-scale climatic changes. However, studies are producing evidence that for some ecosystems the present dynamic equilibrium of the atmosphere depends on the underlying vegetation and the present climate is the consequence of the interaction between the biosphere and the atmosphere (Salati, 1985; Shukla *et al.*, 1990).

While the discussions on the subject proceed and the efforts to improve the measurements and develop more suitable models continue, the occupation by settlers, ranchers, and logging and mining operations press steadily deeper into the forest or into what remains thereof throughout the world (Setzer *et al.*, 1988; Malingreau and Tucker, 1988; Repetto, 1988). Setzer *et al.* (1988) have shown that a sizeable area is burnt during the dry season in Amazonia every year ('queimadas') to give rise to settler's agricultural plots and grazing land in large cattle ranches. Their analysis was based on the satellite counting of satellite-detected fires. A large proportion of the area cleared was not primary forest but either secondary growth or crop residuals. A recent Landsat-based estimate of the total area of primary forest that has been cleared in Brazilian Amazonia up to 1989 (Fearnside *et al.*, 1990) puts this figure close to 400 000 km^2 (or about 10.5% of the area covered by forests), and the annual increase of deforested areas from 1988 to 1989 was about 27 000 km^2. This annual rate may indicate a decline in Brazilian Amazonian deforestation since previous estimates for the annual rate of deforestation were of the order of 35 000 km^2 or higher (Fearnside, 1989). At those deforestation rates most of the forests will be gone in less than 100 years.

The forest clearings are not contiguous but spread over large areas, mostly along the region's roads. Given the yet somewhat small percentage of total deforestation and the fact that it is scattered over a large area, one would not expect large changes in the basin-scale hydrological cycle to have been already detected. Indeed, a recent observational analysis of historical series of precipitation and streamflow in the Amazon basin (Rocha *et al.*, 1989) has not shown any trend in the basin's hydrological cycle that could be attributed to changes in the vegetation cover.

2. Amazon Forest Micrometeorology

For subsequent comparisons, the micrometeorological data collected at the Ducke Forest Reserve, near Manaus, in Brazilian Amazonia, will be briefly reviewed. Research was carried out in a joint program by Brazilian and British researchers from 1983 through 1985. The data analysis has been published in Shuttleworth *et al.* (1984a, 1984b, 1985, and Shuttleworth, 1988) and summarized by Molion (1987).

The research area is covered by dense forest with trees measuring 35 m in average, with branches sometimes reaching 40 m. The micrometeorological observation tower is 45 m high. The solar energy reaching the ground was only 1.2% of that reaching the tree tops. The average albedo for this forest was 12%, varying with the solar zenith angle.

By measuring, during the dry season, vertical temperature and humidity profiles at the canopy level, the radiation budget above the canopy, and the fluxes of sensible and latent heating above the canopy, it was possible to conclude that for fine days 75% of net radiation goes into evaporating water and the remaining 25% is used to heat the air. For a daily average of 4.96 mm water equivalent of net radiation, 3.70 mm was used for evapotranspiration. Continuous measurements of throughfall, steamflow and rainfall at the top of the canopy by Lloyd and Marques (1988), during a period of more than 2 years, led to the estimate that interception loss (rainwater which is intercepted by leaves, stems, and trunks and directly re-evaporated into the air) corresponds to about 10% of the measured rainfall at the top of the canopy. This estimate is somewhat smaller than previous estimates (Franken and Leopoldo, 1984) that interception loss accounts for up to 25% of rainfall. Shuttleworth (1988) points out that 'the spatial variability of precipitation throughfall beneath the canopy of Amazonian rainforest is very high and can result in large and systematic errors unless adequately sampled. This may well have contributed to the extreme variability of previous published results, and quite possibly towards an upward bias in reported interception ratio and canopy storage capacity'. Lloyd and Marques (1988) reported a value of 0.74 mm for the interception storage capacity for the experimental site forest at Ducke Forest Reserve.

Regarding the parameters connected with the wind structure in the surface layer, the following values were obtained: zero-plane displacement (d) was 25.3 ± 0.6m, the roughness length (Z_o) was 5.0 ± 0.4 m, and the friction velocity (u_*) was 0.79 ± 0.13 m/s.

Shuttleworth (1988) developed an empirical model to estimate evaporation for the period September 1983 to September 1985. He used hourly-average meteorological measurements taken above the canopy and soil water tension averaged to a depth of 1 m. Based on this model he summarizes the 2 years and 1 month of micrometeorological data collection as follows: '... over the whole study period, approximately 10% of the rainfall was intercepted by the forest canopy, and this accounted for 20 to 25% of the evaporation. The remainder occurs as transpiration from the trees. Over the same period, about half the incoming precipitation is returned to the atmosphere as evaporation, a process which requires 90% of the energy input. These proportions exhibit some seasonal behaviour in response to the large seasonal variation in rainfall. The average evaporation over two years was within 5% of potential evaporation. Monthly average evaporation exceeds potential estimates by 10% during the wet months and fall below such estimates by at least this proportion in dry months.' Note that for fine days about 75% of the energy input is used for evapotranspiration; in contrast, when all days are taken into consideration, including wet days when evaporation of intercepted water is important, this amount increases to 90%.

For established pastures with grasses 0.6 m high the values for albedo range from 19% when the grass is green to 25% during the dry season. These values are about twice as large as typical albedoes for the rainforest. It means that the absorbed solar energy will likely be less over the areas converted to grass. In reality very little is known about the modifications of the surface microclimate for a grassy vegetation cover compared to the better studied microclimate of the rainforest surface. That has implications for the validation of GCM simulations of tropical deforestation (see section 5 for a review of recent simulations) since most parameters that describe the impoverished grassy vegetation of Amazonian pastures have yet to be measured. To address this gap in knowledge a new joint Brazilian-British micrometeorological experiment will be conducted for a 4-year period, starting in September 1990 in a cleared area in Amazonia, to study in detail the micrometeorological and soil changes associated with converting rainforest into pasture.

3. Water Balance in Small Watersheds

A number of studies have been made near Manaus, Amazon, to measure the various components of the water balance in a watershed. Continuous observations were made for more than two years and are summarized in Franken and Leopoldo (1984). The studies were conducted in two small watersheds.

a. *'Bacia Modelo' Watershed.* Located 80 km north of Manaus, this watershed comprises an area of 23.5 km^2 and is covered mainly by dense primary forest, with tree tops reaching 40 m in average, on a heavy yellow latosol. Precipitation, interception loss, and streamflow of *igarapés* (small forest streams) were measured from 2 February 1980 through 10 February 1981. Total precipitation during that period was 2089 mm; the interception loss was estimated at 534 mm, or 25.6% of precipitation, and the measured total runoff was 541 mm, or 25.9% of precipitation. From these measurements it was possible to estimate that the transpiration by plants was equal to 1014 mm. And the total evapotranspiration was 1548 mm, corresponding to 74.1% of precipitation.

b. *'Barro Branco' Watershed.* It is situated 26 km northeast of Manaus, at the Brazilian Institute for Amazonian Research's Ducke Forest Reserve. Ninety-five percent of the catchment area is covered by dense primary forest, 3% by managed forest, and 2% has been clear cut. The catchment area measures 1.3 km^2 and the soils are mainly yellow latosol. The water balance was measured during two different periods: from 29 September 1976 to 22 September 1977 and from July 1981 to 30 June 1982. The annual average values for the two periods studied shows that precipitation was 2293 mm, interception loss was 429 mm, or 18.7% of precipitation, runoff corresponds to 635 mm, or 27.7% and the transpiration was 1229 mm, or 53.6% of precipitation. Therefore, evapotranspiration was 1658 mm, corresponding to 72.3% of precipitation.

In the two areas studied, the measurements for the water balance components were similar and the evapotranspiration averaged 4.3 mm/day. Ribeiro *et al.* (1979), using monthly mean meteorological data for 1965–73, calculated potential evapotranspiration at 1536 mm/yr and 1075 mm/yr for real evapotranspiration. These results agree with those obtained by Franken and Leopoldo (1984). But, in view of the micrometeorological measurements reviewed in the previous section, the measurements of interception loss in the two watersheds appear to be overestimated. That would explain in part why evapotranspiration in these calculations is about 20% higher than in the calculation of Shuttleworth (1988).

4. Water Balance of the Amazon Basin

The water balance of the Amazon basin is difficult to determine due to the lack of basic data systematically collected over time and space. However, using existing data through successive approximations, it was possible to quantify the fluxes involved.

Figure 1 is a schematic representation of the various fluxes involved in the water balance for the Amazon Basin. The fluxes were estimated according to the methods described below.

4.1. *Aerological Data*

The daily radiosounding (upper air) data from the stations shown in Figure 2 for the years 1972–75 was analyzed. Calculations based on these data produced horizontal and vertical wind and humidity structures, precipitable water vapor, and the atmospheric horizontal moisture flux. For more details on these calculations see Salati *et al.* (1984a, 1984b); Marques *et al.* (1979a, 1979b, 1980a, 1980b); and Kagano (1979).

From these calculations it was possible to conclude that:

a. On the Atlantic coast and in the center of Amazonia winds are predominantly from the east at low levels. Figure 3 shows that most water vapor reaches the region coming from the Atlantic Ocean.

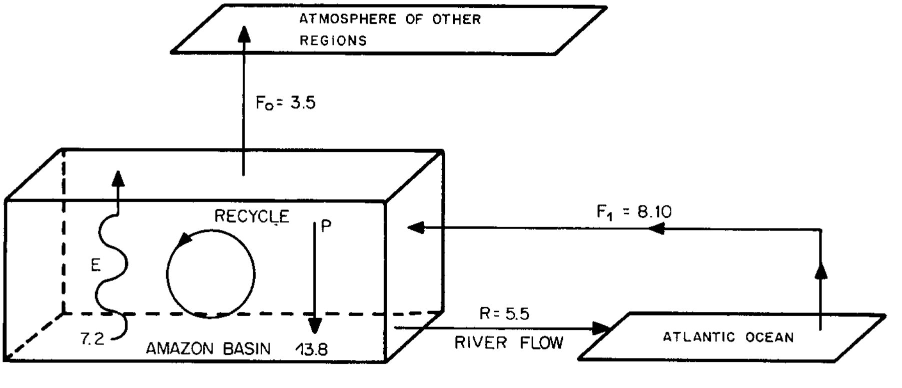

Fig. 1. Schematic depiction of the water cycle in the Amazon Basin. E is evapotranspiration; P is precipitation; F_i represents the amount of water vapor entering the region; F_o represents the amount of water vapor that leaves the region; and R is the flow of the Amazon River into the ocean. The fluxes are in units of 10^{12} m^3/yr.

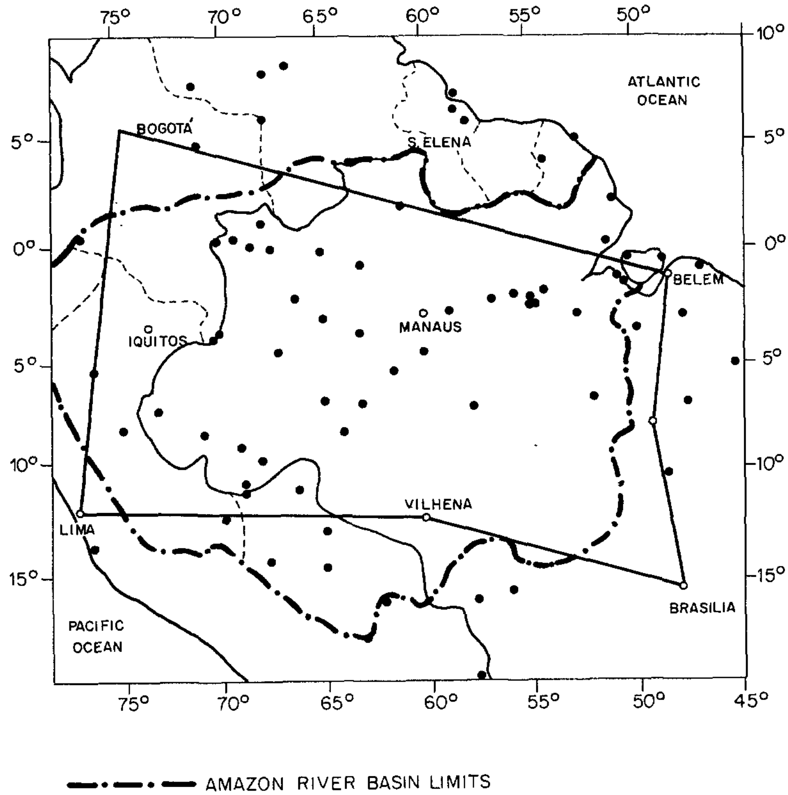

Fig. 2. Network of radiosounding stations used by Marques *et al.* (1980): Bogota, Belem, Carolina, Brasilia, Vilhena, Lima, Manaus.

b. Water vapor fluxes decrease from east to west across the basin.

c. Precipitable water vapor in the region averages 35 mm*, with a seasonal variation of 10 mm. Therefore, the average water vapor stored in the atmosphere above the Amazon basin is of the order of 0.2×10^{12} t. The greenhouse absorption of outgoing longwave radiation by this significant mass of water vapor is largely what accounts for the remarkable isothermal behavior observed in the region (low fluctuation of surface temperature between day and night).

d. Comparison of the seasonal cycle of the basinwide, vertically integrated atmospheric moisture divergence and the Amazon streamflow at Obidos (500 km from the mouth), show that the latter lags the former by approximately 3 months. Therefore, 3 months can be taken as a first-order estimate of the residence time for the water in the Amazon hydrological system (Marques *et al.*, 1980a).

* In the calculations of this average value of precipitable water vapor (PWV) data for Bogota and Lima was included. Because Bogota is on the Andes and Lima on an arid region, typical values of PWV for these two locations are substantially lower than a typical PWV for lowland Amazonia. Therefore the average PWV for the Amazon basin should be higher than 35 mm.

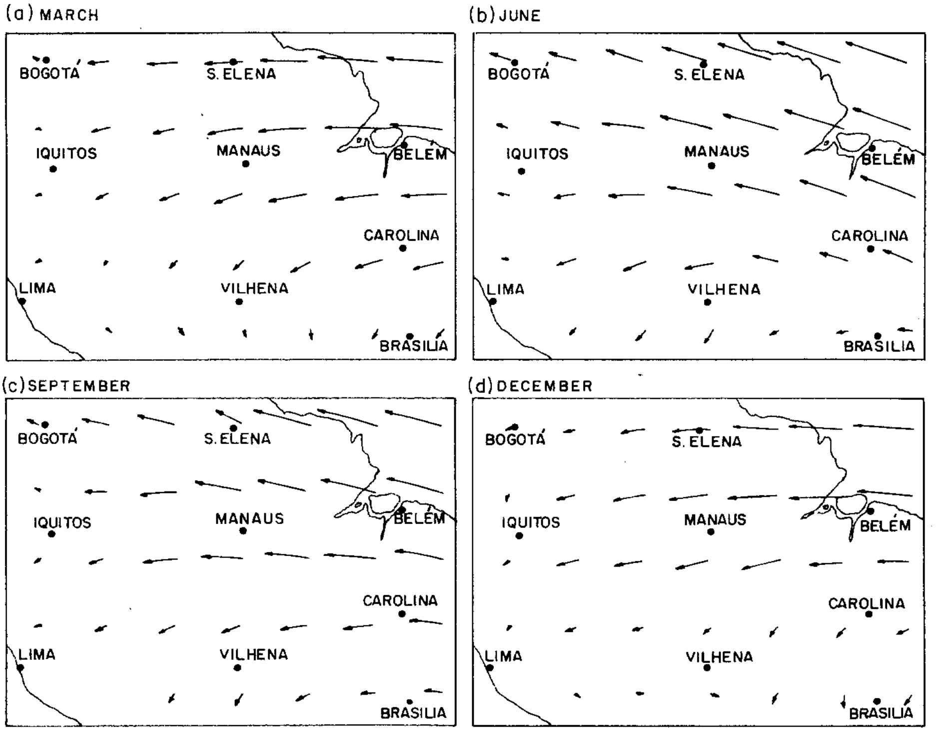

Fig. 3. Vector field values total moisture transport $\Omega = \Omega_\lambda + \Omega_\varphi$. Mean monthly values for the 1972–1975 period obtained for the 5° lat. × 5° long. grid squares. Vector length of 1 cm corresponds to a flux of 2000 g cm^{-1} s^{-1}.

e. Recently, during the Global Atmospheric Experiment-Amazon Boundary Layer Experiment (ABLE-2B), conducted in Amazonia during April–May 1987, upper air data from 6 aerological stations (Figure 4) was collected 4 times a day (00, 06, 12, and 18 UT) during a 1-month period (13 April–13 May 1987). Based on this data, water-balance estimates have been made for the area covered by the aerological stations (approximately 2.2 million km^2). Average precipitation in that area, collected in over 150 rain gauges and for that 1-month period, was 290 ± 20 mm; the calculated convergence of atmospheric water-vapor transport (total water-vapor transport entering the volume comprising the area encircled by the lines in Figure 4 and the depth of the atmospheric column up to 20 km minus the total water-vapor transport exiting the volume) was 127 ± 12 mm. The variation of water vapor in that volume (storage term) during that 1-month period was -6 ± 1 mm, i.e., there was a slight drying from the beginning to the end of the period. The area and time averaged evapotranspiration can then be estimated as the residual term in the atmospheric water budget equation (storage term − convergence term = evapotranspiration − precipitation), resulting in 157 ± 32 mm of evapotranspiration, or

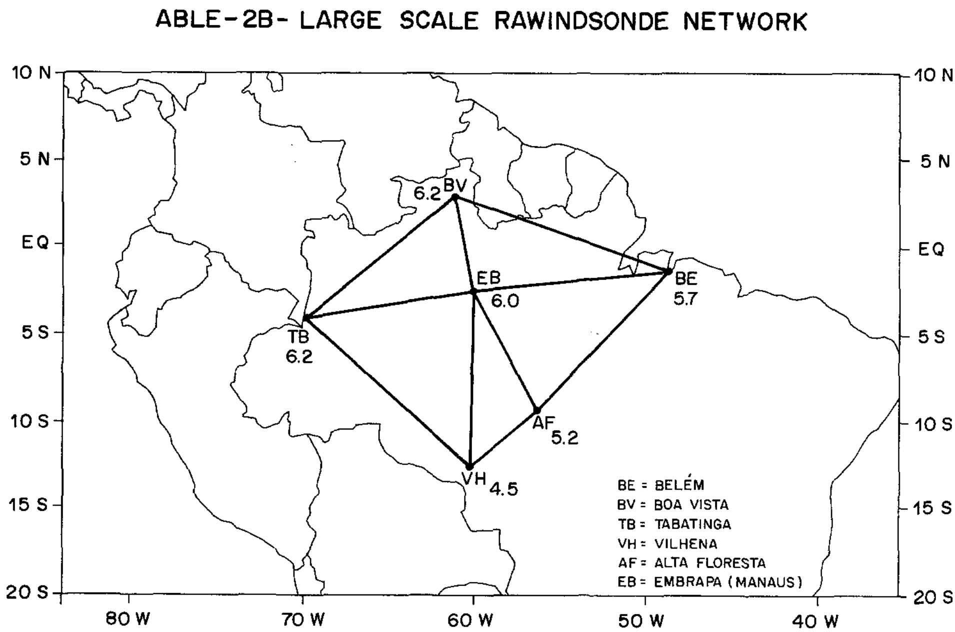

Fig. 4. Network of radiosounding stations used during the GTE-ABLE 2B Experiment (13 April–13 May 1987). The numbers next to the stations represent the mean precipitable water vapor content in cm for that 1-month period.

about 54% of precipitation, which again shows the importance of water-vapor recycling in Amazonia.

That the rates of water-vapor recycling within the basin must be high also can be inferred indirectly by examining the values of precipitable water vapor (the numbers next to stations on the map of Figure 4) in an east-west transect from Belem on the Atlantic coast westward to Tabatinga, which is about 2500 km inland. One can readily see that the values increase westward: 5.7 cm in Belem, 6.0 cm in Manaus, and 6.2 cm in Tabatinga. If oceanic water vapor were the main source of water vapor for precipitation, then one would expect precipitable water vapor to decrease as the air moves inland since rainfall would be extracting water vapor from the atmospheric column at a rate that would be higher than the rate of evapotranspiration into the column.

f. The estimated values of water-vapor fluxes entering Amazonia on the Atlantic coast are not sufficient to explain the values of rainfall observed in the region, that is, the total influx of water vapor into the region is smaller than the total precipitation.

g. There is almost no data on the atmospheric exchange of water vapor between the Amazon basin and Orinoco basin to the north.

h. The inflow of oceanic water vapor from the Atlantic into the Amazon basin is assumed to be about 8–10 × 10^{12} t/y. This is the values with the highest uncertainty in the water balance of Figure 1.

i. At the southern boundary of Amazonia the direction of water vapor fluxes is from north to south for almost the entire year. This shows that water vapor from Amazonia can influence water-vapor concentration in the atmosphere above the Brazilian Highlands.

4.2. *Precipitation*

In general precipitation is abundant along the Atlantic coast from the Guianas to about 3° S. In this coastal strip the total annual precipitation can be as high as 3500 mm. The main rainfall-producing mechanism is related to sea breeze-induced instability lines (bands of cumulus clouds aligned with the coastline). Often these lines propagate inland with phase speeeds typically of the order of 10–15 m/s. On rare occasions they even reach the eastern foothills of the Andes. Their high frequency of occurrence throughout the year (over 250 days in average near the Equator) is the reason behind the 'traditional afternoon showers' at the Brazilian river port city of Belem on the mouth of the Amazon river. From the coast precipitation decreases inland to the east-southeast, reaching values as low as 1600 mm/yr around Santarem, about 700 km from the coast. Rainfall increases again westward of this relative minimum and a broad maximum (annual totals greater than 3000 mm) is found over western Amazonia in the Brazil-Colombia border area. A possible reason for this broad maximum is the proximity of this region to the Andes mountains to the west, which might provide the conditions for low-level convergence of the airflow in that area. There is a secondary precipitation maximum in the southern portion of the basin (values as high as 2500 mm/yr). This region of maximum precipitation, which extends from southwestern Amazonia to the Atlantic, aligned on a NW-SE direction, is related to the interaction of frontal systems moving to tropical and subtropical latitudes with tropical convection over South America. This band of higher precipitation marks the northernmost position that frontal systems reach and where they show a tendency to become quasi-stationary.

The highest rainfall values in South America are found over localized areas on the eastern slopes of the Andes in Peru and on the western slopes in Colombia, where annual precipitation in excess of 5000 mm has been observed. These are caused fundamentally by mechanical uplifting of the prevailing low-level atmospheric flow by the Andean topographic barrier.

The mechanisms that explain the various precipitation maxima mentioned above are all apparently linked to either large-scale features (convergence provided by the Andes, frontal influences) or to other local and mesoscale forcings (topographic uplifting, diurnal land-sea temperature contrasts, etc). They do not appear to depend, to a first approximation, on the type of underlying vegetation. Yet, there is a wealth of observational evidence showing that evapotranspiration accounts for more than 50% of the precipitation (see Table I). This evidence suggests that the

TABLE I: Hydrological cycle of the Amazon Region; summary of the results obtained by different studies (adapted from Salati, 1987)

Research	Rainfall	Transpiration			Evapotranspiration			Runoff	
	mm	mm	%	mm/day	mm	%	mm/day	mm	%
Marques *et al.*, 1980	2328[a]	–	–	–	1260 (r)	54.2	3.5	1068	45.8
	2328[b]	–	–	–	1000 (r)	43.0	2.7	1328	57.0
	2328[c]	–	–	–	1330 (p)	57.1	3.6	998	42.9
Villa Nova *et al.*, 1976	2000[d]	–	–	–	1460 (p)	73.0	4.0	540	27.0
		–	–	–	1168 (r)	58.4	3.2	832	41.6
	2101[e]	–	–	–	1569 (p)	73.4	4.3	532	26.6
Molion, 1975	2379[f]	–	–	–	1146 (r)	48.2	3.1	1233	51.8
Ribeiro and Villa Nova, 1979	2478[g]	–	–	–	1536 (p)	62.0	4.2	942	38.0
		–	–	–	1508 (r)	60.8	4.1	970	39.2
IPEAN, 1978	2179[h]	–	–	–	1475 (r)	67.5	4.0	704	32.5
		–	–	–	1320 (r)	60.6	3.6	859	39.4
DMET, 1978	2207[i]	–	–	–	1452 (p)	65.8	4.0	755	34.2
		–	–	–	1306 (r)	59.2	3.6	901	40.8
Jordan and Heuvel-dop, 1981	3664[j]	1722	47.0	4.7	1905 (r)	52.0	5.2	1759	48.0
Leopoldo *et al.*, 1981	2089[k]	1014	48.5	2.7	1542 (r)	74.1	4.1	541	25.9
Leopoldo *et al.*, 1982	2075[l]	1287	62.0	3.5	1675 (r)	80.7	4.6	400	19.3
Shuttleworth, 1988	2636[m]	992	37.6	2.7	1320 (r)	50.0	3.6	–	–
ABLE-2B, 1987 (1 month)	290[n]	–	–	–	157 (r)	54.1	5.2	–	–

Observations: (r) – real evapotranspiration; (p) – potential evap.; [a] acrological method, applied for all Amazon Basin, period 1972/1975; [b] idem, for the region between Belém and Manaus; [c] by Thornthwaite method, for the region between Belém and Manaus; [d] Penman method, mean for the period 1931/1960; [e] idem, for Manaus Region; [f] climatonomic method, for all Amazon Region, mean for the period 1931/1960; [f] water balance by Thornthwaite and Mather method for the Ducke Forest Reserve, mean for the period 1965/1973; [h] Thornthwaite method for all Amazon Region and estimated for a period over 10 years; [i] idem, for various periods; [j] water balance, with transpiration estimated by class A pan-evaporation for San Carlos Region; [k] 'Model Basin' water balance and [l] 'Barro-Branco' water balance (Ducke Forest Reserve); [m] Adaptation of Penman-Monteith for the period Sept. 1983–Sept. 1985; [n] aerological method applied to the Brazilian AMazon Basin during ABLE-2B, April 13–May 13, 1987.

Amazonian forest is highly efficient in recycling water vapor back into the atmosphere. A different type of vegetation, such as grass, probably would not be as efficient in maintaining high evapotranspiration rates.

There is also a noticeable seasonal variation, with maximum rainfall in the Northern Hemisphere during July–August, and in the Southern Hemisphere during February–March (Salati and Marques, 1984). The average total annual precipitation for the Amazon basin has been estimated by various authors ranging from 2000 mm to 2400 mm. Recently Molion and Dalarosa (1990) have shown some evidence that these values might be underestimates because most rain gauges in Amazonia are located on the banks of large rivers and are possibly under the influence of a local diurnal river breeze circulation that would reduce precipitation over the river and its margins compared to forest areas distant from the river. With that in mind, a value of 2300 mm/yr was used for the water balance in Figure 3, which corresponds to a flux of 13.8×10^{12} m^3/yr.

4.3. *Evapotranspiration*

Accurate measurements or estimates of real evapotranspiration for the various ecosystems present in Amazonia are not available presently. Evapotranspiration estimates can currently be obtained through the following methods: (a) aerological calculation for large areas; (b) water-balance calculations for small watersheds; (c) direct micrometeorological flux measurements; (d) isotopic dilution calculation; and (e) calculations based on the surface energy budget (Thornwaite, Penman, Monteith methods). Evapotranspiration estimates made by various authors for Amazonia are summarized in Table I.

For the Amazon basin as a whole, values ranging from 1146 mm to 1320 mm, corresponding to 48–54% of the precipitation, were obtained. In the water balance shown in Figure 3, a value of 1200 mm/yr was used, corresponding to an annual input of 7.2×10^{12} t of water vapor from the surface into the atmospere.

4.4. *Recycling of Water Vapor*

In view of the results found for water-vapor fluxes and precipitation, it is possible to conclude that water vapor has to recirculate in the region, i.e., total precipitation is larger than the total influx of water vapor into Amazonia. Supporting evidence also comes from the distributions of isotopes (0–18 and D) of rainwaters of different areas of Amazonia (Salati *et al.*, 1979). Using current existing data it is estimated that 50–60% of rainfall originates from the recirculated water vapor through evapotranspiration.

5. Model Simulations of Amazonian Deforestation

Quantitatively estimating the effects that large changes in Amazonian ecosystems can have on the surface-energy and water budgets has been difficult because the equilibrium climate is determined by the momentum and energy exchanges at the Earth-atmosphere interface, interacting with complex dynamical processes in the atmosphere. Results of earlier model studies were generally inconclusive, and sometimes conflicting, about the regional (and global) climate changes following deforestation. The models were of two types: either energy-box models (Lettau *et al.*, 1979; Potter *et al.*, 1975) or crude resolution GCMs (Henderson-Sellers and Gornitz, 1984). In general, the latter lacked both spatial resolution and an adequate treatment of the land-surface processes. For instance, their resolution was typically 10° long. × 5° lat., which would cause the whole of Amazonia to be represented by a few grid-points. Also their representation of evapotranspiration processes was based on simple parameterizations. In Henderson-Sellers and Gornitz (1984) runoff was proportional to soil moisture and to the precipitation. These parameterizations, including the 'bucket hydrology' parameterization, were inadequate to represent evapotranspiration processes over vegetated surfaces (Sellers *et al.*, 1986) and make it difficult to represent the complex changes in soil hydrology following burning and land clearance. Henderson-Sellers (1987, her Table 1, pp. 468–469) summarized the main results from these earlier model studies.

Realistic models of biosphere have only recently been developed that can be coupled with realistic models of the global atmosphere (Dickinson *et al.*, 1986, Sellers *et al.*, 1986). The pioneering work of assessing climate impacts of tropical deforestations using these novel coupled biosphere-atmosphere models was that of Dickinson and Henderson-Sellers (1988), hereafter referred to as DHS. In DHS the National Center for Atmospheric Research Community Climate Model (NCAR CCM), coupled to the Biosphere-Atmosphere-Transfer-Scheme (BATS) of Dickinson *et al.* (1986), was used with a horizontal resolution of 7.5° long. × 4.5° lat. to study the effects of Amazonian deforestation. When the model's rainforests over Amazonia were replaced by degraded pasture, surface temperatures increased by 3–5 °C and evapotranspiration decreased over the region. The increase in surface temperature was attributed mostly to the decreased roughness length of the grass vegetation compared to that of forest and the reduction of evapotranspiration was mostly due to less absorbed solar radiation for grass given its higher albedo. Some difficulties were reported in the parameterization of incident solar radiation and of interception loss (Shuttleworth and Dickinson, 1989; Dickinson, 1989a, 1989b) that caused unrealistically high net radiation.

More recently two GCM simulations of tropical deforestation were conducted, one at the UK Meteorological Office (Lean and Warrilow, 1989, hereafter referred to as LW) and another at the Center for Ocean-Land-Atmosphere Interactions (COLA) (Shukla, Nobre, and Sellers, 1990, hereafter referred to as SNS). In LW the model's horizontal resolution was 3.75° long. × 2.5° lat. and all the model's vegetation north of 30° S in South America was replaced by grass. Although the total area in which the model's vegetation changed was almost twice that used in DHS and in SNS, their results were similar to those in DHS: surface temperature increased by 2.5 °C and evapotranspiration decreased for the pasture scenario compared to the forest one.

Additionally, it was found that simulated precipitation was reduced over Amazonia. As in DHS the increase on surface temperature was attributed to the decrease in roughness length. Table II (adapted from Table 2 of LW) summarizes the main results of their study.

In SNS the COLA GCM, coupled to the Simple Biosphere Model (SiB) of Sellers *et al.* (1986), was used with a horizontal resolution of 2.8° long. × 1.8° lat. –

TABLE II: Summary of surface variables for control (C) and deforested (D) simulations averaged over 3 years for Amazonia (from: Lean and Warrilow, 1989)

Surface variable	C	D	
Evaporation (m/d)	3.12	2.27	(−27.2%)
Precipitation (m/d)	6.60	5.26	(−20.3%)
Soil moisture (cm)	16.13	6.66	(−58.7%)
Runoff (mm/d)	3.40	3.00	(−11.9%)
Net radiation (W/m^2)	147.3	126.0	(−14.4%)
Temperature (°C)	23.6	26.0	(2.4 °C)
Sensible heat (W/m^2)	57.2	60.2	(+5.2%)
Bowen ratio	0.85	1.50	(+76.5%)

TABLE III: Mean surface-energy budget for Amazonia. The data are 12-month mean (January–December) values. Values are in W/m^2, except for B and a which are nondimensional, and T_s which is in °C. S is insolation; a is albedo; L_n is net upward longwave radiation; R_n is available radiative energy; E_t is transpiration plus soil evaporation; E_i is interception loss; E is evapotranspiration = E_t + E_i; H is sensible heating; G is ground heat flux; B is the Bowen ratio (H/E); and T_s is surface temperature (from: Shukla *et al.*, 1990)

	S	$(1-a)S$	L_n	R_n	E_t	E_i	E	H	G	B	a	T_s
Control	233	204	−32	172	91	37	128	44	0	0.34	12.5	23.5
Deforestation	237	186	−40	146	64	26	90	56	0	0.62	21.6	26.0
Difference	+4	−18	−8	−26	−27	−11	−38	+12	0	−0.28	+9.1	+2.5

TABLE IV: Mean water budget for Amazonia. The data are 12-month mean (January–December) values. Values E and P are in mm/yr; PW is in mm. P is total precipitation; E is evapotranspiration; and PW is precipitable water (from: Shukla *et al.*, 1990)

	P	E	$(E-P)$	E/P	PW
Control	2464	1657	−807	0.67	37.7
Deforestation	1821	1161	−661	0.63	35.4
Difference	−642	−496	+146	−0.04	−2.3
Change (in percent)	−26.1	−30.0	+18.0	−5.9	−6.1

i.e., the simulation with the highest horizontal resolution among the three studies – and the model's Amazonian tropical forests were replaced by degraded grass. The main results of SNS are summarized in Tables III and IV for the surface-energy and water balances in Amazonia, respectively (adapted from Tables 1 and 2 of SNS), and described below.

Surface and soil temperatures were warmer by 1–3 °C in the deforested than in the control cases. The relative warming of the deforested land surface and the overlying air is consistent with the reduction in evapotranspiration and the lower surface roughness length. The annual mean surface-energy budget (Table III) for Amazonia in the two simulations shows that absorbed solar radiation at the surface is reduced in the deforestation case (186 W/m^2) relative to the control case (204 W/m^2) because of the higher albedo (21.6%) for grassland compared to forest (12.5%). That plus the larger outgoing longwave radiation from the surface due to the higher surface temperature in the deforested case result in the amount of net radiative energy available at the surface for partition into latent and sensible heat flux being smaller in the deforested case (146 W/m^2) than in the control case (172 W/m^2). Also, as remarked in SNS, less precipitation is intercepted and re-evaporated as the surface roughness and the canopy water-holding capacity of the pasture are relatively small. Furthermore, the transpiration rates are reduced due to the reduced soil moisture-holding capacity for the soils under pasture.

An interesting result was that the reduction in calculated annual precipitation (642 mm) was larger than the reduction in evapotranspiration (496 mm), as seen in Table IV, which suggests that changes in the atmospheric circulation may act to reduce further the convergence of moisture flux in the region, a result that could not have been anticipated without the use of a dynamical model of the atmosphere, as noted in SNS. This, in turn, implies that runoff also decreased for the deforested case, a result also found in LW (Table II), since the decrease in precipitation was larger than the decrease in evapotranspiration.

Taken together the results of these three studies seem to suggest the existence of a significant sensitivity of the regional climate to the removal of the tropical forest. In general, the somewhat short period of integration in these studies precludes drawing conclusions on the significance of global climate changes or even climate changes in regions adjacent to Amazonia.

6. Discussion and Concluding Remarks

The conversion of tropical forested areas into pastures or other types of short vegetation will cause changes in the microclimate of the disturbed areas. If the size of the perturbed area is sufficiently large, even the regional climate may be altered. Depending on the scale of these alterations, they may cause climate changes at the global level and affect regions distant from the tropical forests.

6.1. *Local Changes in Climate*

Changes will occur in albedo and in energy and water balances. There will be a tendency toward less water infiltration and more runoff during rainy periods and less runoff during prolonged dry periods.

An important conclusion of the micrometeorological studies conducted at Ducke Reserve, near Manaus in central Amazonia (summarized in Shuttleworth, 1988), is that the annual flux of latent heat into the atmosphere is close to its potential value, that is, 20% smaller than the potential evapotranspiration during the dry season and about 10% above the evapotranspiration rate during the rainy season (implying a net transfer of sensible heating from the atmosphere into the canopy, i.e., negative Bowen ratio*).

Shuttleworth suggests that there might be a reduction of between 10 and 20% in the evapotranspiration for pastures as compared to the rainforest, mostly due to the higher albedo (thus, smaller available energy other things being equal) of grass compared to the albedo of tropical forests. That reduction, in turn, might cause rainfall to decrease by 10%, he suggested. Yet, this hypothetical scenario takes into account only changes in evapotranspiration due to changes in the available radiative energy. Important changes also would occur due to the decrease in surface roughness and at the soil level. Loss of top soil organic matter and soil fauna, compaction due to agricultural practices and overgrazing, and soil erosion may cause large changes in the physical and chemical characteristics of the predominantly clay soils of the Amazonian terra firme forest. Those changes likely would combine to reduce infiltration rates drastically, increase surface runoff during rainy periods, and decrease soil moisture in the shallower rooting zone of the grass vegetation primarily during the dry season. Decreased soil moisture availability also would contribute to reduce evapotranspiration.

Comparative measurements of the diurnal cycle of canopy and subsurface temperature at cleared and forested sites in Ibadan, Nigeria (Lawson *et al.*, 1981), and in Surinam (Shulz, 1960) showed a large increase of soil (> 5 °C) and air (> 3 °C) temperatures for the cleared areas compared to the forested ones. Not being in the shade of a tall canopy, the diurnal fluctuation of ground temperature and humidity deficit was much larger for the cleared sites in these two studies as well. Those changes in soil microclimate will have a profound effect on the biological, chemical, and physical processes in the top soil layer. Plants, animals, and microorganisms living in that layer will experience temperature, humidity deficit, and water stresses not present in the remarkably constant microclimate of the forest floor.

6.2. *Regional Climate Changes*

The summation of local climate change over a sufficiently large quasi-contiguous area (say larger than 1 million km^2) might change water-vapor transports and the water balance at a regional level with consequent changes in the energy balance. Climatic alterations and the scale at which they occur depend on the geographic location and its geomorphology. For instance, even small changes in the low-level wind regime on mountainous areas such as the Andean Cordillera can cause a large change in the temporal and geographical distribution of rainfall. It is not possible yet to predict accurately regional climate changes associated with the observed pat-

* It is not clear, though, how this condition could be maintained for a large area since a negative Bowen ratio would make the layer above the canopy more stable, therefore reducing the turbulent flux of latent heat from the canopy.

terns of deforestation by means of climate model simulations. An important reason for such limitation is that when current climate models are integrated in a control mode, i.e., attempting to mimic the observed climate, they commonly fail to represent important aspects of the regional climate. One problem is, of course, resolution. It is expected that only when model resolution becomes of the order of 100 km (current climate model resoltion is typically between 200 and 500 km) will the models probably capture the finer details of the regional climate. Yet, the results of recent climate model simulations of Amazonian deforestation, reviewed in the previous section, suggest the following changes at the regional level to be likely following extensive deforestation of tropical forests: increase in surface and soil temperature and in the diurnal fluctuation of temperature and specific humidity deficit, and a reduction of evapotranspiration and PBL moisture. In two of the three studies (LW and SNS), yearly averaged precipitation and runoff decreased for Amazonia as a whole for the pasture vegetation compared to forest. The annual reduction in rainfall in these two simulations was larger than the corresponding reduction in evapotranspiration, thus explaining the reduction in runoff. It is likely, however, that runoff will increase following rainy periods, that is, runoff (and river streamflow) would be higher after deforestation during the rainy season and decrease during the dry season.

6.3. *Global Changes*

Tropical forests contribute in many ways to maintain the present dynamical and chemical equilibrium of the atmosphere. Forests represent a carbon reservoir, both through their areal and root systems as well as through organic matter in the soil. Estimates indicate that tropical forests possess a reserve of carbon equivalent to 1.5–2.0 times the carbon store of CO_2 in the atmosphere. Therefore, conversion of forests into pastures will release CO_2 from the biosphere into the atmosphere, likely enhancing the greenhouse warming.

Forest burning associated with clearing processes for conversion into pastures also releases great quantities of particles and compound gases into the atmosphere. These particles cause changes in the atmosphere, especially in its chemical composition and energy balance.

To understand and predict any possible large-scale climate change due to tropical deforestation it is crucial to know to what extent the rainfall patterns will change when rainforests are converted into grasslands. It is well known that the tropical regions function as atmospheric heat sources through the release of latent heat of condensation in convective clouds. The heat so released drives large-scale tropical circulations (of the Hadley-Walker type) with ascending motion over the tropical regions, mostly over Amazonia, Tropical Africa, and the Indonesia–western Pacific region, and descending motion over the dry subtropics, primarily over the subtropical oceans. It is conceivable that a significant reduction in rainfall over Amazonia (say, greater than 20% reduction as the model simulations described in LW and

SNS suggest) might have an effect in these tropical circulations. However, it is unclear what these changes would be and how they would manifest themselves in terms of climate changes in the Tropics, but away from the perturbed areas, and in the extra-tropics. Regarding the extra-tropics, it is interesting to note the suggestion by Paegle (1987) of a possible link between tropical convection and quasi-stationary features of the large-scale circulation over North America. He suggests that the westward shift of the subtropical jetstream from the east coast of North America in boreal winter to the west coast in spring and a concomitant westward shift of the North American long-wave trough may be linked to the seasonal, northwestward migration of the area of rainfall maxima over Tropical South America from Central Amazonia in January–February to Central America in June–July.

Tropical forest areas also have a characteristic energy balance that contributes to the transport of energy as latent heat (water vapor) from the equatorial regions to those of greater latitude. This is particularly conspicuous in Central Brazil, southern Bolivia, Paraguay, and northern Argentina where, due to the generally southward low-level circulation, most of the water vapor present in those regions comes from Amazonia. Therefore, changes in atmospheric moisture in Amazonia due to deforestation might have an impact on the precipitation of the adjacent regions to the south.

So far we have focused our attention mostly on the Amazonian tropical forest. Can we say anything about climatic impacts arising from the removal of tropical forests in Equatorial Africa and Southeast Asia? It is likely that at the microclimate level the effects will be quite similar: higher near-ground temperatures and larger diurnal fluctuations of temperature and humidity deficit, increased runoff during rainy periods and decreased runoff during the dry season, decreased soil moisture, and, possibly, decreased evapotranspiration. The question whether there would be a significant change in precipitation is a complex one. For Southeast Asia large-scale changes in precipitation are less likely since the precipitation climate of that area of the western Pacific and Indian Oceans is controlled by large-scale features: on one hand, the precipitation distribution responds to the high sea surface temperatures (SST > 28 °C) that are conducive to large rates of evaporation besides a tendency for the low-level air to converge from areas of lower SST to areas with higher SST; these two factors enable cloud formation and high precipitation. On the other hand, land-sea heating contrast drives the monsoonal circulations of Southeast Asia. The monsoonal circulations account for the copious rainfall observed in that area.

In Africa there is, at least theoretically, the possibility that the removal of the tropical forest might influence the regional climate. A biophysical feedback mechanism as proposed by Charney *et al.* (1977) might cause an enhancement in aridification along the northern and southern boundaries of the forest. For reasons similar to the ones discussed in the earlier section, the changes in albedo, surface roughness, and soil moisture caused by replacement of forest by overgrazed pasture would result in decreased precipitation. That could, in turn, induce further clearings deeper into the forest. However, this question is not settled yet because interannual and longer-term rainfall variability in Tropical Africa is apparently also connected to planetary-scale phenomena, notably global SST distributions.

Finally, can we say anything on the ecological implications of the possibility of a future dryer and warmer climate in Amazonia following extensive deforestation? The decrease in precipitation suggested by the simulation studies for the deforested case is associated with a longer and more pronounced dry season. The authors in SNS remark that '[T]he lack of an extended dry season apparently sustains the current tropical forests, and therefore, a lengthening of the dry season could have serious ecological implications. Among other effects, the frequency and intensity of forest fires could increase significantly and the life-cycles of pollination vectors could be perturbed.... Changes in the region's hydrological cycle and the disruption of complex plant-animal relations could be so profound that once the tropical forests were destroyed, they might not be able to re-establish themselves'. The authors then conclude that a 'complete and rapid destruction of the Amazon tropical forest could be irreversible'. Therefore, there might be a tendency of 'savannization' of Amazonia if one recalls that the savanna vegetation is naturally more adapted to withstand fire and a long (6 months or greater) dry season. Amazonia is surrounded to the south, east, and north by savanna-like vegetation. Any trend toward 'savannization' in Amazonia would likely be seen first in the transition forests straddled between the rainforest and the savanna because in those areas the dry season is usually longer than in the rainforest. This implies that any increase in the duration of the dry season in those regions might make it unsuitable for the re-establishment of the rainforest.

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