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TROPICAL ECOSYSTEMS

Systems Characteristics, Utilization Patterns, and Conservation Issues

Edited by
W. Erdelen, N. Ishwaran, P. Müller

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Climatic Impacts of Tropical Land Use Practices

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Abstract

Tropical land use practices that might have impacts on regional or even global climate include clearing of rain forests, the burning of biomass, paddy field cultivation, cattle breeding, and overgrazing of semi-arid pasture-grounds. These land use types modify climatic parameters such as land-surface evapotranspiration, concentrations of atmospheric trace gases and aerosol particles, albedo and roughness of the earth's surface. The large-scale influence of changes of these parameters on temperature, precipitation, and general atmospheric circulation is discussed. Many effects appear to be small-scale effects which sometimes compensate each other. However, a variety of trace gases released from tropical areas under land use seem to be increasingly enhancing the earth's greenhouse warming.

Introduction

In recent years, information on human impacts on state and evolution of the global climatic system has grown considerably. This applies not only for urban-industrial activities in highly developed regions outside the tropics, but also for land use practices essential for subsistence in tropical and subtropical regions. Among the latter, the most important activities significantly impacting the climate are: (1) burning of biomass for clearing forests for shifting cultivation and combustion of fuel wood, grassland and agricultural wastes, (2) paddy field cultivation and cattle breeding, (3) large-scale clearing of natural forests (deforestation), and (4) overgrazing of pastures in semi-arid regions.

These activities not only modify microclimatic conditions in the areas affected, but may also influence the macroclimate on a regional or even global scale. In this paper only the latter aspect is discussed. Direct consequences of the activities listed for particular climatic parameters are outlined and possible implications for large-scale climatic modifications are discussed.

Effects of Land Use on Particular Climatic Parameters

Biomass Burning

Burning of biomass, up to 80 % of which is carried out in the tropics (Seiler and Conrad, 1987), has proved to be a major source of atmospheric trace gases (Crutzen *et al.*, 1979). These gases are known to have significant effects on the earth's climate. Some of them, such as carbon dioxide (CO_2), methane (CH_4) and nitrous oxide (N_2O) absorb specific wavelengths within the infrared and thus contribute to the earth's greenhouse-warming. In addition, chemical reactions of methane with hydroxyl radicals (OH) decrease concentrations of OH and thus lead to higher concentrations of other gases which are predominantly removed by reaction with OH (Crutzen, 1987).

Other gases, such as carbon monoxide (CO), nonmethane hydrocarbons (NMHCs) and nitric oxide (NO_x), are important in the photochemical production of tropospheric ozone (O_3) which is also a greenhouse gas. Some chemical compounds become important in the upper atmosphere, e.g. carbonyl sulphide (COS) which contributes to the production of stratospheric sulphate aerosols. Mean annual releases of trace gases due to biomass burning constitute substantially large percentages of those gases released globally or due to industrial activity alone (Table 1). The uncertainty factor in these estimates might be as high as $2^{\pm 1}$. More recent measurements made in the tropical savanna of Central Venezuela, during the dry season, have introduced additional complexities (Hao *et al.*, 1988): emissions of CO_2, CH_4 and N_2O from burnt and unburnt plots were similar, whereas emissions of CO_2 and N_2O increased considerably after burnt areas were subjected to simulated rainfall.

Apart from atmospheric gases, solid particles within the atmosphere will also be affected by biomass burning. Solid particles emitted due to biomass burning might be 10 % of the total released from urban-industrial, deflational and pyrogenic sources. Estimated annual particle emissions by human activities are about 300 megatons by urban-industrial activities, 100-250 by deflation caused by land use, and 40-60 by biomass burning (Bryson 1974, Root 1976).

Table 1. Estimated release of atmospheric trace gases (in teragrams) due to biomass burning and comparisons with industrial and total releases (mean values with factors of uncertainty up to $2^{\pm 1}$; see text for abbreviations). From: Crutzen *et al.* 1985, Seiler and Conrad 1987.

	Mean annual release	*% of industrial release*	*% of total release*
CO_2	2650	51.0	26.0
CO	1000	156.3	30.0
CH_4	79	111.3	18.5
NMHCs	35	?	?
COS	0.04	23.5	6.6
N_2O	1.5	75.0	11.1
NO_x	6	30.0	20.7

Paddy Fields and Fermentation by Ruminants

Certain types of land use are known to have impacts on the chemical composition of the atmosphere. Microbial activities in paddy fields lead to the formation of biogenic CH_4 which is transported through the rice plants before being released into the atmosphere (Seiler et al., 1984). Another biogenic source of atmospheric CH_4 is the enteric fermentation in animals, mainly in domestic ruminants. Two thirds of the world's stock of cattle is resident in developing countries (Seiler, 1984, p. 469). These biogenic sources of atmospheric CH_4 were 28% and 20%, respectively, of total emissions in 1980. CH_4 emissions from biogenic sources have increased considerably during the past 40-50 years due to the near doubling of the total area of rice cultivation and of cattle population (Pearman and Fraser, 1988).

The annual production rates of CH_4 increased 2.6% (paddy fields) and 1.0% per year (ruminants), respectively, between 1950 and 1980 (Table 2). These rates of increase were even higher than the rate of increase of CH_4 emissions due to biomass burning (+ 0.9% per year). Since the main sink for CH_4 is the chemical reaction with atmospheric OH, increasing CH_4 release tends to decrease atmospheric OH concentrations thus leading to further increases in concentration of other trace gases which are predominantly removed by reaction with OH (Crutzen, 1987).

Table 2. Annual production rates of CH_4 (in teragrams) from various sources for the years 1950 and 1980. *From: Seiler 1984, Seiler and Conrad 1987.*

	1950	1980
Paddy fields	18-35	70-170
Ruminants	49-69	73- 99
Other biogenic sources	19-77	43-101
Biomass burning	41-74	56-102
Industrial sources	24-25	55- 87

Deforestation

The large-scale conversion of tropical rain forests into grasslands or agricultural areas has gained widespread attention. Expected consequences of this activity on the earth's climate are believed to be the most significant among all human impacts in the tropics. A 20% decrease in evapotranspiration due to clearing of tropical rain forests has been estimated (Flohn, 1973). This would translate into an annual loss of latent heat of about 17 TW per 10^6 km^2 resulting in increased sensible heat of the atmosphere.

During forest clearing organic matter in biota and terrestrial humus is oxidized to release CO_2 into the atmosphere. In contrast to shifting cultivation or forestry, where regrowth occurs soon after clearing, conversion of forest to permanent agricultural land leads to a notably higher CO_2-release because soil organic matter is broken down (Liss and Crane, 1983). Two thirds of the area of tropical forest removed have been converted to agricultural land (Houghton *et al.*, 1985). Carbon release due to deforestation, occurring predominantly in the tropics, accounts for nearly 80 % of the total emissions from terrestrial biota (Woodwell *et al.*, 1983). These emissions have increased continuously since 1860. Emissions from fossil fuels overtook that from terrestrial biota only since the 1960s (Fig. 1). For 1980, net annual release of carbon into the atmosphere, due to tropical deforestation, was estimated between 0.5 and 4.2 Gt while the release from fossil fuels was 5.2 Gt (Houghton *et al.*, 1985). Difficulties in obtaining more precise estimates result from poor knowledge of the uptake of C by the oceans (roughly amounting to 2 Gt per year; Bach *et al.*, 1985), from CO_2-variations in the atmosphere due to varying temperature and large-scale El Niño events, and from possible effects of CO_2 fertilization of land biota (Kohlmaier *et al.*, 1985). Temperate forests might indeed still act as a global sink for CO_2 representing a 'recreational effect' after the large-scale clearings in the 19th century, whereas tropical forests and soils appear to be growing sources of CO_2 release since about the 1950s, mainly due to clearing of forest for permanent land use (Fig. 2).

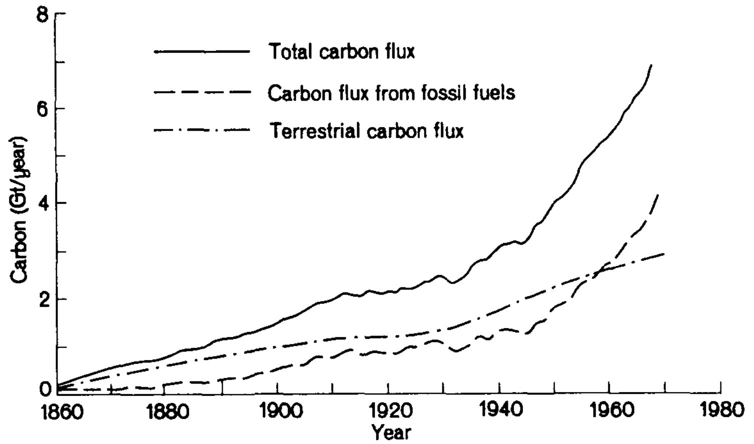


Figure 1. Global annual carbon flux into the atmosphere from 1860 to 1970. From: Moore *et al.*, 1980 quoted in Bach *et al.*, 1985.

Conversion of woodland to grassland also changes the surface albedo from about 15% to 20% (Baumgartner and Kirchner, 1980). Averaged zonal values of the mean annual increase in surface albedo that would result from global deforestation, i.e. conversion of all forests into grassland, are given in Fig. 3. In the tropics the largest changes (> 1%) are likely to occur within the equatorial belt. These changes would modify the radiation balance (less absorption of solar radiation) and might introduce additional variations, especially if degradation would continue, leading to further changes in surface albedo.

Deforestation also impacts the aerodynamic properties of the earth's surface which influence atmospheric circulation in the lower troposphere. These properties are commonly described by a roughness parameter r which may be estimated from vegetation height h and the empirical constants a = 1.03 and b = 0.86 (Baumgartner *et al.*, 1978):

$$\log r = a \cdot \log h - b$$

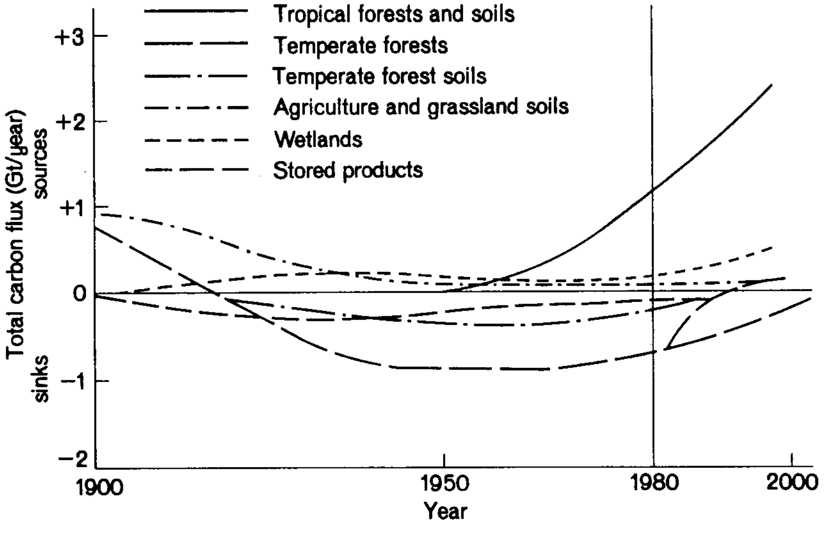


Figure 2. Total annual carbon emissions from 1900 to 1980, and projections beyond 1980. From: Loucks, 1980 quoted in Bach *et al.*, 1985.

A tropical rain forest with a mean vegetation height of h = 30 m would have a r value of 4.59 m, whereas a tropical grassland area with a height of only 1 m will have a r value of 0.14 m. After global deforestation there would be a general drop in r by some 15-20 cm within the tropics, still increasing near the equator where we presently find the highest tropical values (Fig. 4).

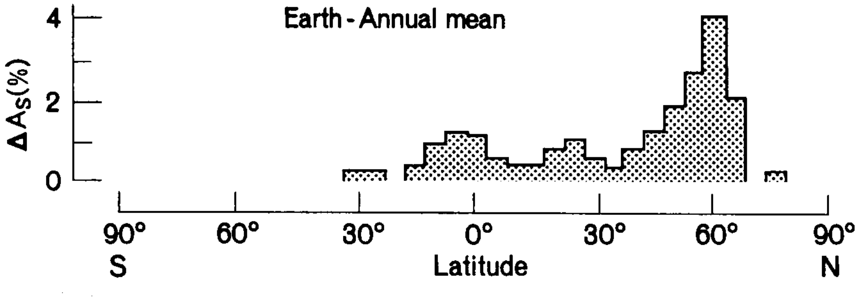
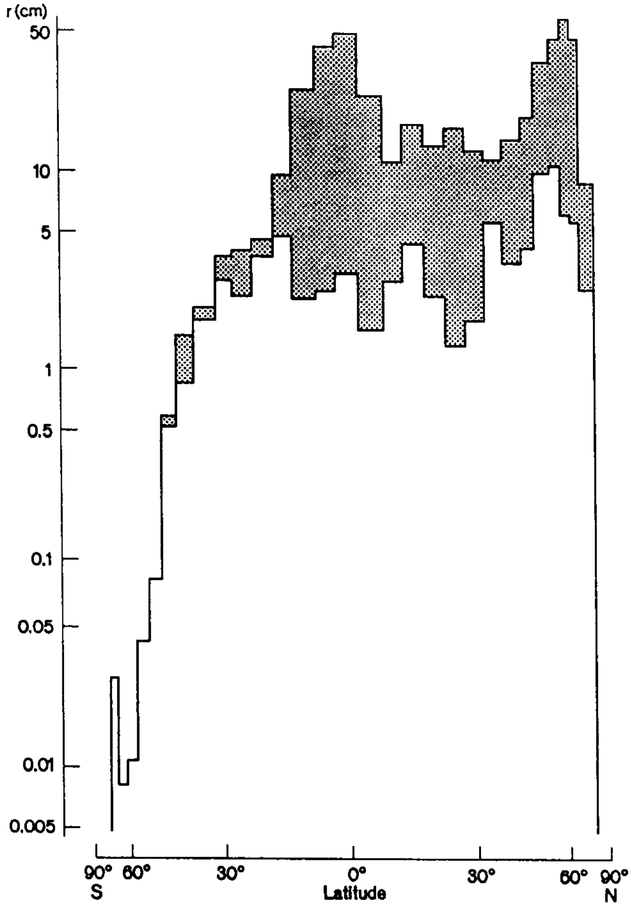


Figure 3. Zonally averaged changes of mean annual surface albedo (A_S) in different latitudinal zones due to global deforestation. From: Baumgartner and Kirchner, 1980.

Figure 4. Averaged zonal values of the surface roughness parameter r for present conditions (upper curve) and those after global deforestation (lower curve). From: Baumgartner and Kirchner, 1980.



Overgrazing

Overgrazing in semi-arid regions of the tropics, by growing livestock populations or further extensions of pasture lands into increasingly drier regions, is coupled with degradation of natural vegetation which may lead to changes in particular climatic parameters. Initially, surface albedo could rise by roughly 5% (Klaus, 1981), increasing up to 10-15% under conditions of desertification (Klaus, 1986). Reduced vegetation cover and increased surface albedo may, however, also result from naturally occurring periods of drought. The dry season albedo in the Sahelian region reached a maximum of roughly 30% during the peak of the drought in 1973 (Courel *et al.*, 1984). However, vegetation has recovered since then and surface albedo has decreased to 20% in 1979 (Courel *et al.*, 1984) despite the fact that the drought continued until the mid-1980s. This might be attributable to the fact that overgrazing was higher during pre-1973 years but decreased afterwards, because the drought continued.

Reduced vegetation cover will also increase dust emission due to intensified deflation and thus contribute to the anthropogenic particle emissions, one third of which are due to land-use-caused deflation. During drought years in the Sahel region mean aerosol concentrations at Barbados, which is affected by the North African dust plume, increased by a factor of 3

(Prospero and Nees, 1977). During the same time dust-storm activity in Mauritania and Sudan increased by factors of 6 and up to 5, respectively (Middleton, 1985). Although these changes are primarily caused by naturally occurring drought conditions, overgrazing probably exacerbated the effects.

Implications for Climatic Modifications

Tropical land use has been shown to influence parameters such as land-surface evapotranspiration, concentrations of atmospheric trace gases, concentrations of atmospheric particles, albedo of the earth's surface and roughness of the earth's surface. Climatic modifications that might result from these influences are discussed below.

Land Surface Evapotranspiration

Anthropogenic changes in land surface evapotranspiration are often thought to drastically alter the earth's climate, especially its mean precipitation, since the average land-surface evapotranspiration is about two thirds of the mean precipitation (Baumgartner and Reichel, 1975). General circulation models simulated precipitation during the month of July (Shukla and Mintz, 1982) for two extreme scenarios, i.e. the wet soil case, in which evapotranspiration was equal to potential evapotranspiration and the dry soil case with no evapotranspiration at all. These simulations yielded somewhat different results (Fig. 5). In the dry soil case Europe and most of Asia and North America lost the greatest part of their July precipitation but, within the tropics, rainfall either remained nearly constant (South America, Southeast Asia) or was about the same as the difference between precipitation and evapotranspiration in the wet soil case (North Africa). Accordingly, the horizontal transport of water vapor from the oceans and its convergence by the atmospheric circulation lead to humid conditions within the greater part of the tropics, even without any land surface evapotranspiration. Land surface July temperature, however, increased by about 15 to 25°C north of 20°S (Shukla and Mintz, 1982). This is due to the absence of evaporative cooling and decreased cloud cover. Although these calculations are based on rather unrealistic assumptions (extreme values of evapotranspiration and constant values of surface albedo), they provide useful estimates of trends in climatic modifications caused by human impacts on vegetation cover and soil moisture. In summary, reduced land surface evapotranspiration, for example due to large-scale deforestation, most likely causes significant changes in precipitation in regions outside the tropics, but only small or insignificant changes within the humid tropics themselves.

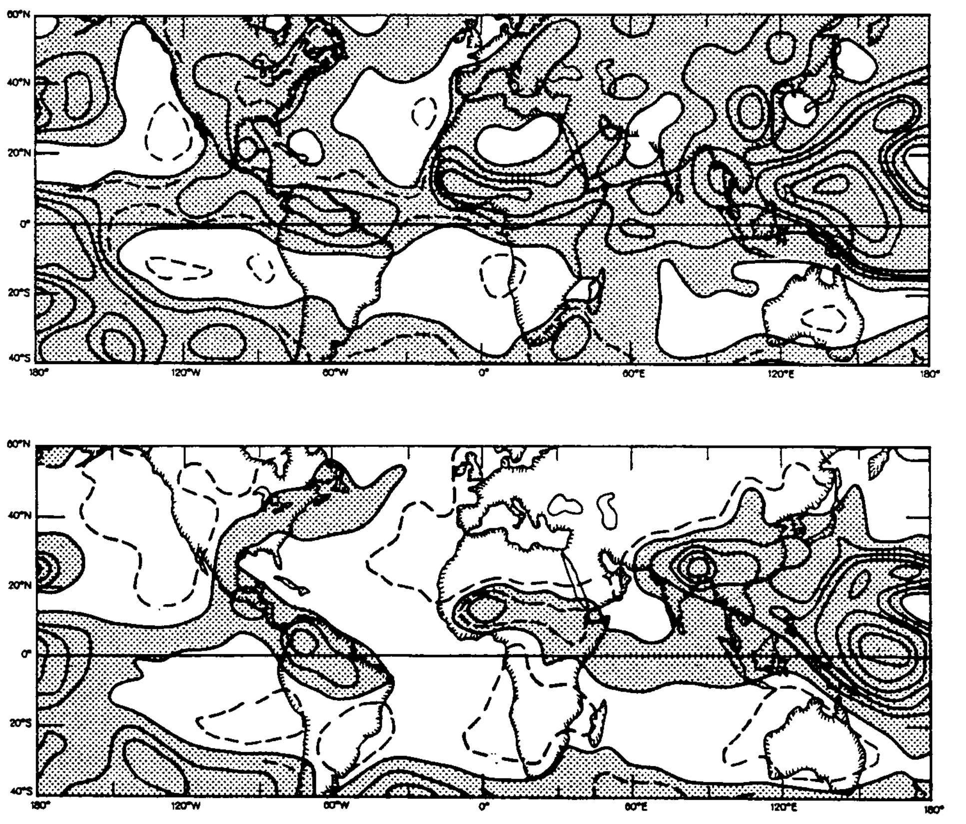


Figure 5. Simulated July precipitation (mm/day) for assumed conditions of potential evapotranspiration (upper part) and no evapotranspiration at all (lower part). Contour interval is 2 mm/day, areas above 2 mm/day are stippled. From: Shukla and Mintz, 1982.

Atmospheric Trace Gases

A number of trace gases released as a result of tropical land use are considered to be "greenhouse gases" which contribute to the earth's warming by selectively absorbing infrared radiation. Estimates for global temperature changes due to increased concentrations of greenhouse gases still differ considerably depending on the models used. Assuming a doubling of the atmospheric CO_2-content they vary between +1.3 and +4.8°C. Other greenhouse gases may also contribute to a rise in temperature. Tropospheric ozone (O_3), methane (CH_4), and chlorofluorocarbons (CFCs) currently change at rates greater than that of CO_2 (Tab. 3).

Table 3. Atmospheric trace gas concentrations, rates of change, and estimated global temperature increases due to CO_2-doubling and to time-equivalent increases of other trace gases, respectively (see text for abbreviations). From: various authors in Schönwiese and Runge (1988).

	Present concentration	Present rate of change (% per year)	Estimated global temperature effect (°C)
CO_2	347.0 ppm	+ 0.40	+ 1.3 to + 4.8
CH_4	1.7 ppm	+ 1.50	+ 0.09
N_2O	0.3 ppm	+ 0.25	+ 0.12
O_3	30.0 ppb	+ 1.00	+ 0.90
CFCs	0.4 ppb	+ 4.00	+ 0.50

Carbon monoxide (CO), nonmethane hydrocarbons (NMHCs), and nitric oxide (NO) which are released in significant amounts by tropical biomass burning, are involved in photochemical reactions thus indirectly contributing to the formation of tropospheric O_3. The extent to which CO_2-induced warming, between 1850 and 1980, may already have contributed to global temperature has been analyzed by Schönwiese and Runge (1988). Their multiple regression model allows to separate natural temperature variations (by keeping constant the CO_2-term) and thus to estimate temperature increase due to man-made greenhouse-warming only (Fig. 6). The differences between observed and naturally caused global temperatures had reached some 0.4 K around 1980. This value was still below the 90% level of statistical significance, but would rise to 0.7 K which is beyond the 95% level, if other greenhouse gases were also to be included (Schönwiese and Runge, 1988).

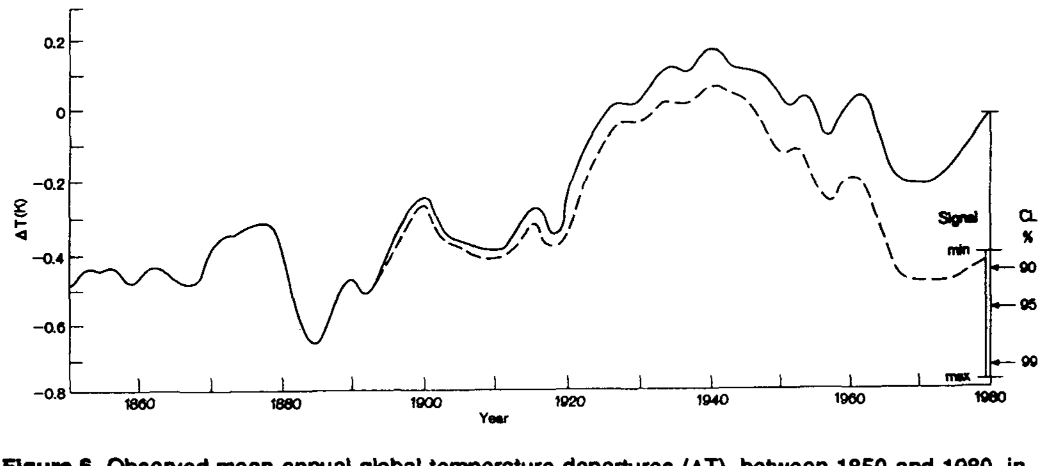


Figure 6. Observed mean annual global temperature departures (ΔT), between 1850 and 1980, in relation to 1980 (continuous line) and simulated ΔT without man-made CO_2-increase (dashed line). Consideration of additional greenhouse gases would enhance the signal above indicated confidence levels (CL). From: Schönwiese and Runge, 1988.

Hydrological parameters would also change in response to increased trace gas concentrations in the atmospher

e. A general circulation model (Manabe and Wetherald, 1980) has provided information about the changing differences between precipitation and evaporation over continental areas that result from a doubling of the atmospheric CO_2-content (Fig. 7). Humidity would increase north of about 50°N and between 15°N and 35°N, but would decrease in the latitudinal band between 40°N and 50°N. The arid belt would shift northwards thus causing profound changes in the sensitive subtropical regions. However, additional spatial variation must be taken into account, since the increased greenhouse-warming amplifies with latitude thus reducing meridional temperature gradients and inducing changes in the atmospheric circulation.

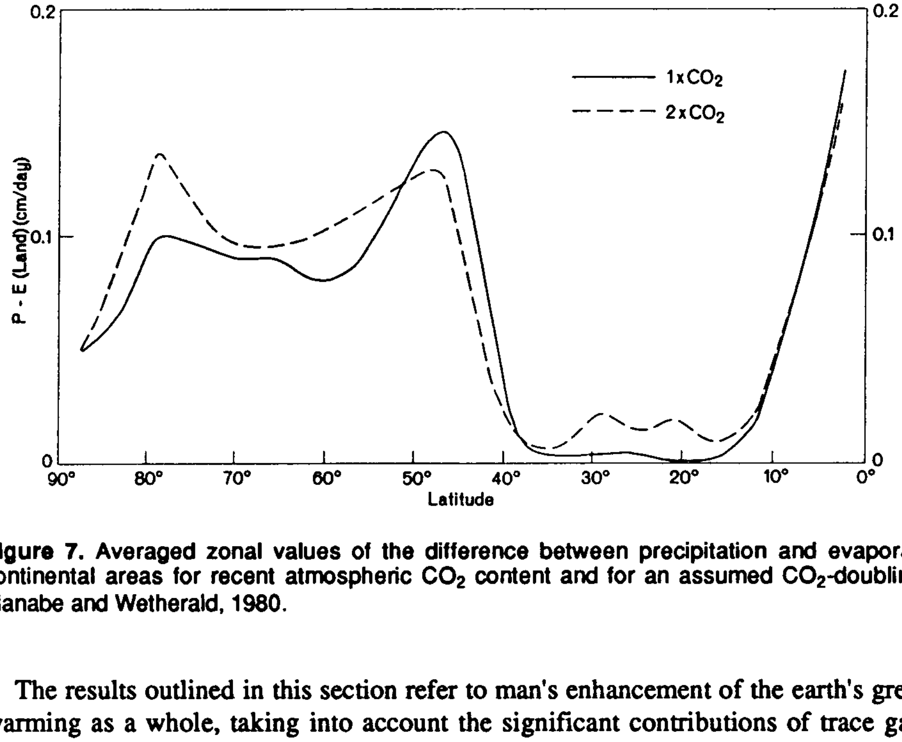


Figure 7. Averaged zonal values of the difference between precipitation and evaporation over continental areas for recent atmospheric CO_2 content and for an assumed CO_2-doubling. From: Manabe and Wetherald, 1980.

The results outlined in this section refer to man's enhancement of the earth's greenhouse-warming as a whole, taking into account the significant contributions of trace gases from industrial sources of fossil fuel combustion. Tropical land use alone, however, still contributes substantially to the increase in concentrations of greenhouse gases (see section 2.) and therefore partly causes the impacts discussed here.

Atmospheric Particles

Impacts of aerosol particle concentration on climate are more difficult to estimate, since particle size and radiation characteristics, as well as their rates of production and spatial distribution vary considerably. According to Grassl (1987) impacts on the radiation budget are similar to those of greenhouse gases only if there is a large amount of soot in regions of high surface albedo. Over the northern hemisphere, however, there is the opposite effect with increased albedo and radiative losses (Fig. 8) which amount to one third or, at most, one half of the expected gain from a CO_2-doubling (Grassl, 1987). The center of man-made albedo changes due to aerosol particles lies clearly outside the tropics, even during winter with widespread dry seasons in the tropics and increased biomass burning (Fig. 8).

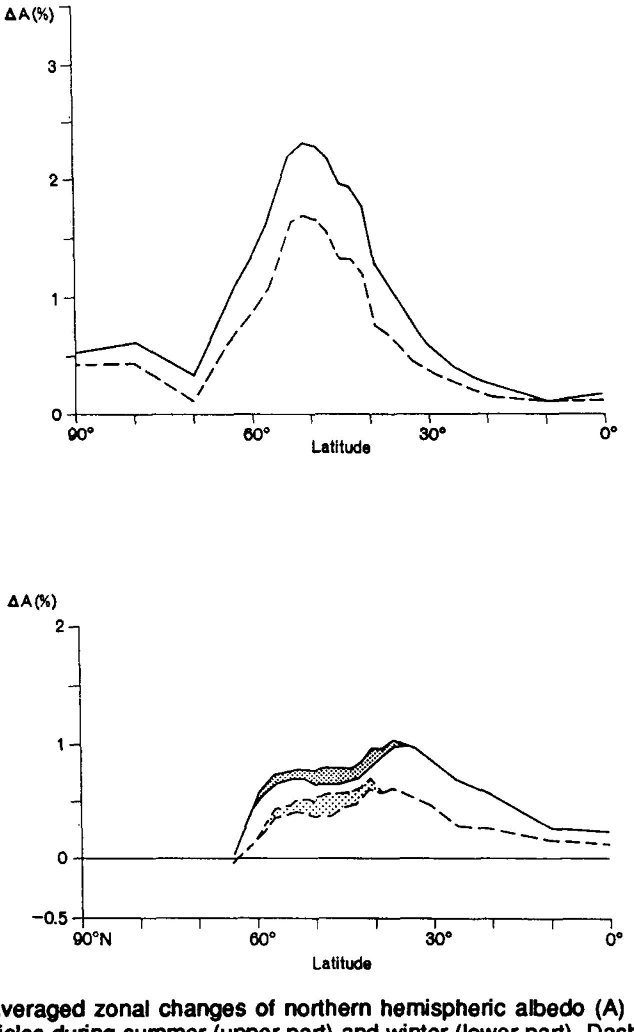


Figure 8. Averaged zonal changes of northern hemispheric albedo (A) caused by anthropogenic aerosol particles during summer (upper part) and winter (lower part). Dashed line for soot fraction of 20 %, stippled area for gas-particle-conversion. From: Grassl, 1987.

In respect of tropospheric dust concentrations originating from the Saharan desert regions, Klaus (1981) summarized radiation data obtained from GATE measurements. When vertically integrated over the layer between 600 and 6100 m, the energy source from incoming short wave radiation was 0.958 J $\times$ cm^{-2} $\times$ min^{-1} in the absence of dust, but only 0.669 J $\times$ m^{-2} $\times$ cm^{-1} under dusty conditions. Energy loss from outgoing long wave radiation was 0.172 J $\times$ cm^{-2} $\times$ min^{-1} in the absence of dust, but 0.209 J $\times$ cm^{-2} $\times$ min^{-1} under dusty conditions. Accordingly, the presence of dust particles leads to increase cooling of about 0.76°C for the whole dust layer (Klaus, 1981). This cooling supports vertical stability throughout the whole troposphere and thus may contribute to further reductions in rainfall. Dust concentrations in the Sahelian region, being significantly higher during dry phases, primarily respond to climatic conditions. Degraded vegetation, caused by overgrazing, however, will enhance wind erosion and increase dust concentrations. The troposphere in these semiarid tropics will thus become more stabilized.

Table 4. Simulated changes of mean global parameters due to increased surface albedo (from 7% to 25%) and increased runoff rate resulting from total tropical deforestation. From: Potter *et al.*, 1975.

Global parameters	Albedo increase only	Additional increase in runoff rate
Surface air temperature	- 0.3 °C	- 0.2 °C
Evaporation	- 0.9 %	- 1.0 %
Atmospheric water vapor content	- 3.2 %	- 2.8 %
Cloudiness	- 0.4 %	- 0.6 %
Precipitation	- 0.4 %	- 0.9 %
Residence times of atmospheric H_2O	- 2.3 %	- 1.9 %

The Earth's Surface Albedo

Increased surface albedo implies reduced surface absorption of solar energy and hence may result in further changes in global climate. Potter *et al.* (1975) used a two-dimensional (zonally and vertically resolved) atmospheric model to simulate these changes assuming total removal of tropical rain forests (Tab. 4). Thermal and hydrological parameters stabilized on a somewhat lower level, following an increase in surface albedo from 7% to 25% throughout the areas presently covered by tropical rain forests. Additionally, increasing runoff rates led to a lower decrease in temperature, but to higher decreases in evaporation, cloudiness, and precipitation. The atmospheric water vapor content, however, decreased less since the acceleration of the atmospheric water cycle (decreased residence times of atmospheric H_2O) was also less (see Tab. 4). The calculated values were not significant on a global scale (change of 0.2-0.3°C of mean annual surface air temperature or less than 1% of mean annual rainfall), even under the extreme assumption of total tropical deforestation. On a smaller spatial scale, however, more significant effects may be expected. For instance, simulated annual rainfall decreased up to 1.5-2.6% in high northern latitudes, whereas between 5°-25°N and S, respectively, there was even an increase in precipitation due to greater cooling in the upper tropical troposphere caused by reduced release of latent heat and subsequent steepening of lower tropospheric lapse rates (Potter *et al.*, 1975). Assuming widespread desertification, increased albedo around 20°N enhanced surface cooling, especially in high northern latitudes, up to 3°C through activation of the ice-albedo feedback mechanism (Potter *et al.*, 1980).

Regional consequences of changing surface albedo were studied by Charney (1975). He modelled radiation balance and departures of the tropospheric temperature field from radiative equilibrium due to thermally induced and frictionally controlled circulations in the North African atmosphere. With regard to the possible impact of Sahelian overgrazing on the regional climate, these temperature departures have been calculated for two contrasting values of surface albedo, viz. 14 %, representing an average for a continent covered by vegetation, and 35 %, representing common desert conditions. Fig. 9 shows the different deviations during summer for the latitudinal belt between 18° and 25°N, with lower positive values in the lower troposphere and higher negative values in the upper troposphere in the case of high desert-like albedo. These changes imply an enhancement of the mean sinking motion (roughly doubling in the middle troposphere) and cause additional drying. Without consideration of land-surface evapotranspiration this model suggests a tendency towards increasingly arid conditions with increasing surface albedo only. Thus, large-scale overgrazing may contribute to the perpetuation of distinctly dry conditions in the Sahelian region. Similar results have been obtained from a general circulation model (Charney, 1975) which yielded reduced mean rainfall between 18° and 26°N for the same increase in surface albedo (Fig. 10). Together with the indicated rise in average precipitation further south this corresponds to a southward shifting of the rain-bearing ITCZ (Intertropical Convergence Zone) by several degrees of latitude. However, such dramatic consequences may be expected only if a vegetational belt would be totally converted to a desert. The results discussed here are corroborated by calculations taking into consideration changing evaporation (Charney *et al.*, 1977) and by more recent simulations (Sud and Fennessy 1982, 1984).

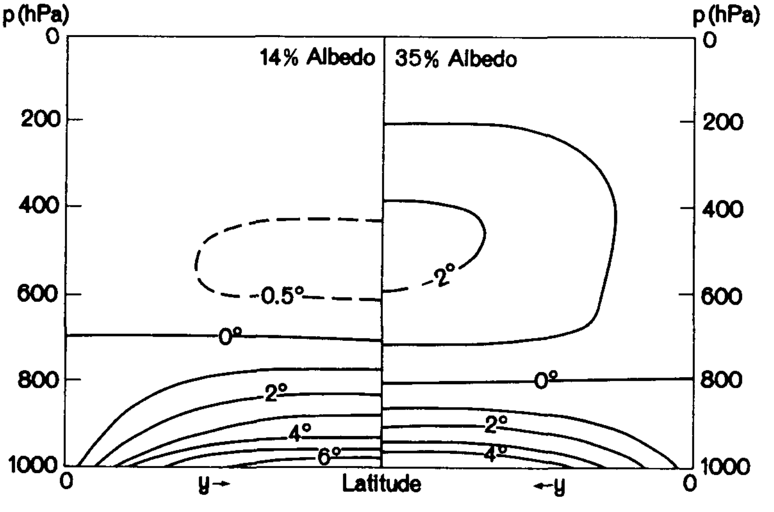


Figure 9. Deviation of summer temperature above the southern Sahara (roughly between 18° and 25°N with y coordinate pointing northward) from radiative equilibrium for albedos of 14% and 35%, respectively (vertical scale is in units of atmospheric pressure p). From: Charney, 1975.

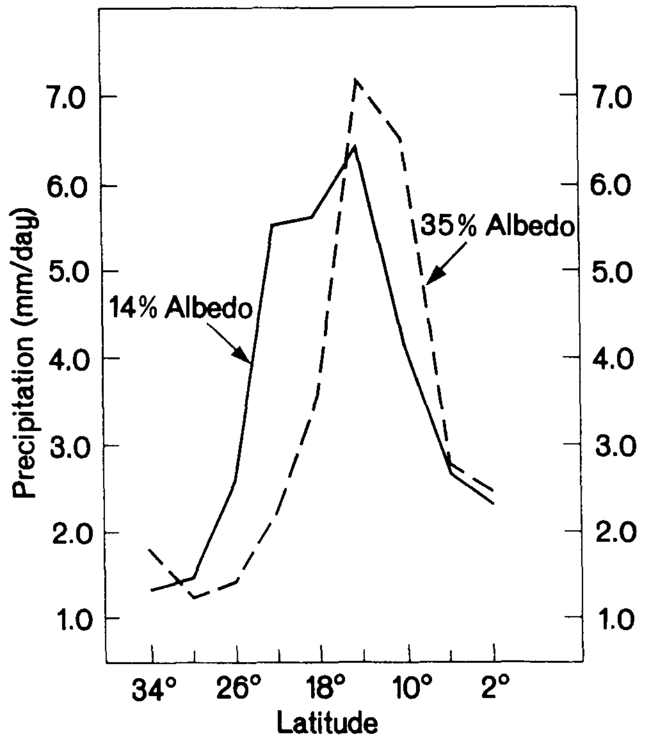


Figure 10. Simulated mean July rainfall over North Africa as a function of latitude for albedos of 14% and 35%, respectively. From: Charney, 1975.

Modifications of the Tropospheric Circulation

The general circulation of the tropical troposphere is characterized by large-scale easterlies above the friction layer which develop between the subtropical cells of high pressure and the equatorial trough of low pressure. Within these easterlies combined barotropic-baroclinic instabilities cause widespread wave developments known as 'easterly waves', especially in the middle and the lower troposphere. These waves cause most of the precipitation in tropical areas outside the equatorial belt. In regions with continental influence, within the lowest tropospheric layer equatorial westerlies develop which spread poleward into the summer hemisphere lagging behind the sun's zenith by 1-2 months. Frictional forces from the earth's surface finally cause ageostrophic deflections towards the lower atmospheric pressure. In the lowest layer this induces NE- and SE-trades, respectively, where tropical easterlies are present, but SW- and NW-monsoons, respectively, where equatorial westerlies have penetrated.

Tropical land use may influence these circulations and affect tropical rainfall conditions. This is corroborated by circulation studies of anomalously wet and dry months within different regions of the tropics (Jacobeit, 1989b). Referring to the semiarid region of northeastern Brazil (1° - 6°S, 38° - 49°W), daily grid point data of the horizontal wind components at the 200 hPa level during anomalously wet and dry months in the first half of the 1980s were analyzed using principal component analysis (see Tab. 5). The scores of the seven eigenvectors extracted (Fig. 11) represent the most important circulation patterns of the upper troposphere during these anomalous months. Each of these patterns dominates the upper tropospheric circulation in different months (Tab. 5).

Table 5. Mean monthly variances (in %) of daily wind fields at the 200 hPa level during anomalously wet (upper half) and dry months (lower half) in NE Brazil explained by the first seven principal components shown in Fig. 11. From: Jacobeit, 1989b.

		Principal Component						
	Date	1	2	3	4	5	6	7
	May 1984	20.3	1.8	4.6	43.2	3.0	1.9	3.4
Wet	Jan. 1985	3.7	13.4	23.6	3.4	5.7	5.8	23.9
	Feb. 1985	3.7	7.1	42.8	4.1	5.8	10.2	1.5
	Feb. 1981	2.3	14.8	9.6	1.9	20.9	25.5	4.6
Dry	April 1981	22.7	6.4	3.5	11.5	25.7	2.8	7.3
	Jan. 1983	9.2	50.2	6.9	1.7	5.1	3.7	2.6
	May 1983	54.9	4.7	3.4	10.1	4.5	1.1	1.7

A common feature of those patterns dominating during wet months is to have easterly components around NE Brazil, and of those patterns dominating during dry months to have westerly components (Fig. 11). Similar results were obtained for other tropical regions (Jacobeit, 1989b) and for whole monsoonal seasons (Jacobeit 1988, 1989a). These studies show that a strong development of upper easterlies is crucial for good monsoonal rainfall, whereas weakened easterlies or westerlies in upper layers tend to depress monsoonal rainfall. Accordingly, the formation of rain-bearing easterly waves in middle and lower tropospheric layers imply the existence of a well-developed easterly current reaching high enough into upper tropospheric levels (Riehl, 1979). Extension and intensity of the tropical easterlies, however, generally depend on meridional gradients of temperature and pressure within their borders. Generation of easterly waves requires a minimum of lateral shearing or of thermal contrast. Tetzlaff *et al.* (1985), for example, give a threshold value of 2-2.5 K/280 km for reaching conditions of baroclinic instability above the Sahelian region. Tropical land use is expected to affect these meridional gradients. Increasing surface albedo and rising concentrations of atmospheric particles due to overgrazing in the outer tropics would lower mean temperatures in these regions. Large-scale clearing of rain forests in equatorial regions, however, would raise mean temperatures due to additional sensible heat which is no more "required" for maintaining evapotranspiration. Altogether there would be a lowering of mean meridional temperature gradients in the tropics and a reduction in the easterlies' strength as well as in the generation and intensity of easterly waves. It is still impossible to make more precise quantitative statements, but the general trend towards reduced rainfall for monsoonal regions is obvious.

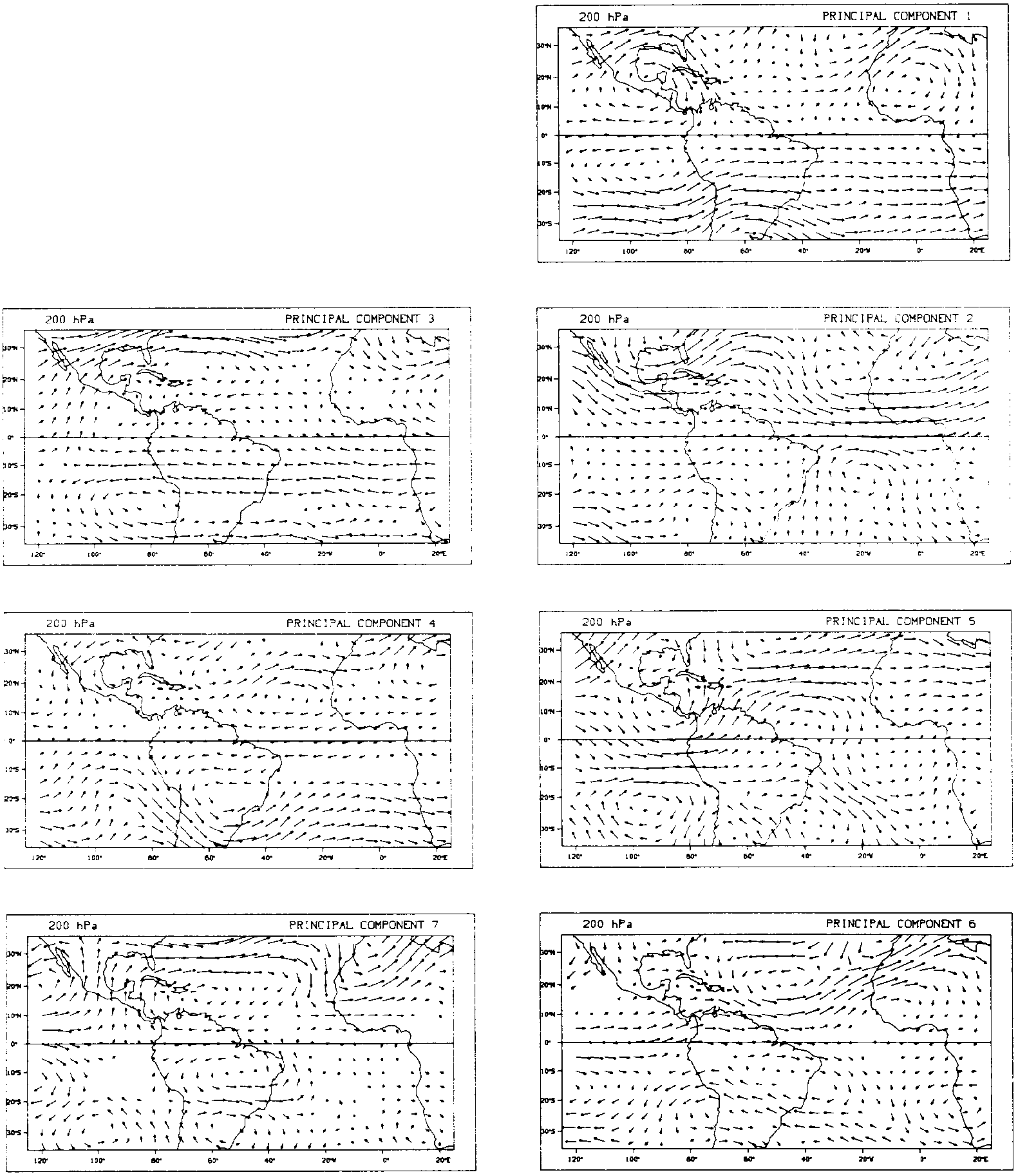


Figure 11. Scores of the first seven principal components of daily wind fields at the 200 hPa level during anomalously wet and dry months in NE Brazil. Circulation patterns on the left side dominate during wet months, those on the right side during dry months. See Tab. 5 for comparison. From: Jacobeit, 1989b.

A further argument arises from reduced surface roughness resulting from widespread conversion of tropical forests to grasslands or agricultural areas. Reduced roughness means reduced frictional force from the surface, and this again means reduced a geostrophic deflection of surface winds from the geostrophic direction parallel to surface pressure isobars (Fig.12). The angle of deflection, α (Kirchner, 1977) may be expressed as

$$\alpha = \frac{c}{\log\frac{v_g}{2w \times \sin\varphi \times r}} - d$$

where v_g = geostrophic wind strength at the surface
r = roughness parameter
w = angular velocity of the earth
φ = geographical latitude
c,d = empirical constants

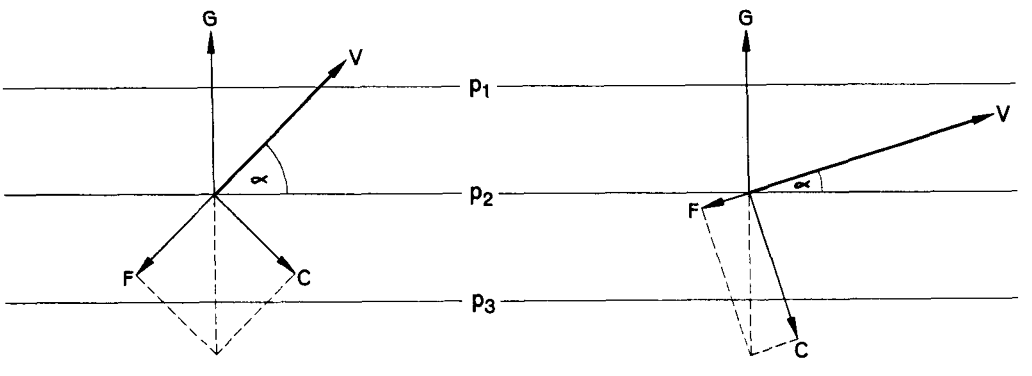


Figure 12. Deflection of surface wind from the geostrophic direction as a function of frictional force. v = wind vector, G = gradient force, C = Coriolis force, F = frictional force, α = angle of deflection, p_i = pressure.

Taking, for example, the summer monsoonal winds of the lower troposphere (southwesterlies in the northern, northwesterlies in the southern hemisphere), these would have poleward components smaller than at present if blowing over deforested areas with reduced surface roughness. As the geographical meridians converge with growing latitude, air masses flowing with a poleward component experience a tendency towards vertical stretching. This implies an upward motion favoring convection. Smaller poleward components, however, reduce vertical stretching and convective stimulation. As a consequence, a more zonal flow of monsoonal winds due to reduced surface roughness might lead to a reduced rising motion in the mean, regardless of all other conditions varying with different synoptic situations. For instance, for an air mass with vertical depth of 2000 m being transported from 5° to 15°N, α would be 34.4° in completely forested areas, where r = 4.59 m (calculation based on a mean latitude of φ = 10°, a mean geostrophic wind strength of v_g = 5m/s, and values from Baumgartner *et al.* (1978) of c = 173.58, and d = 3.03). The vertical stretching resulting from the meridians' convergence between 5° and 15°N, amounts to roughly 24 m and would be reduced to about 18 m (with $\alpha = 25.2°$) in completely deforested areas with r = 0.14 m. These values are indeed not very high with respect to the whole vertical depth of 2000 m. Of course, both assumptions, i.e. either complete forest cover or complete deforestation, are far from reality. However, the rate of change amounts to roughly 25% in the case mentioned above. Again, there is a general trend towards reduction of monsoonal rainfall. Further considerations, based on the stress-differential divergence mechanism, indicate similar changes for the near-equatorial coastal region of Northeast Brazil and yield considerably reduced annual rainfall due to the change in surface roughness (Roth, 1984).

Conclusions

The orders of magnitude of anthropogenic impacts of tropical land use practices on the regional or global climate apparently are not very high. Moreover, some of the model calculations were based on unrealistic assumptions, e.g. total deforestation or no evapotranspiration at all, and the influence of some effects shows opposing trends (e.g. greenhouse gases and albedo changes). Several trace gases released by tropical land use, however, seem to significantly contribute to the progressive enhancement of the earth's greenhouse-warming. Even minor man-made climatic changes might induce additional variations within the climate system. This would be true, particularly, if our climate system should prove to be almost-intransitive (Lorenz, 1976), i.e. that even small changes of boundary conditions are sufficient to induce completely different successions of climatic state transitions. Thus, in any case, man's activity is a critical factor for our highly sensitive climatic system.

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