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THE SIGNIFICANCE OF AMAZONIAN RAIN FOREST DEFORESTATION FOR REGIONAL AND GLOBAL CLIMATE CHANGE – A REVIEW

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Resumo. Este artigo é direcionado às questões de uso de terra e mudanças climáticas que afetam as funções biológicas, químicas e físicas da Amazônia, com particular ênfase à sustentabilidade e sua influência nas mudanças climáticas globais. O presente trabalho é uma iniciativa de alto grau de multidisciplinidade em pequena, média e grande escala, incluídos monitoramento à longo prazo, com intensivos trabalhos de campo, bem como a modelagem e análise de dados, focalizando aspectos hidrológicos, bioquímicos e climatológicos. Futuras atividades serão concentradas em fatores climáticos relevantes em florestas primárias, florestas secundárias e áreas de agricultura, investigação de efeitos de áreas de transição, pesquisas do efeito de fertilizantes de dióxido de carbono, emissões de gases de efeito estufa para flora e fauna amazônica e mudanças no ciclo hidrológico devido às mudanças climáticas, desmatamentos e subsequentes uso da terra.

Abstract. The paper addresses the question as to how land use and climate change will affect the biological, chemical, and physical functions of Amazonia, with particular reference to its sustainability and its influence on global climate change. Present research initiatives display a high degree of multi-disciplinarity and are conceived around small-, medium-, and large-scale frameworks. They include long-term monitoring in intensive data-collection campaigns, as well as modeling and data analysis, with a predominant focus on hydrological, biogeochemical and climatological aspects. Future activities should concentrate on: accurate measurements of climate-relevant factors in primary forest, secondary forest and agricultural areas; the investigation of edge effects; research into the carbon dioxide fertilizer effect; emissions of greenhouse gases by Amazonian flora and fauna, and changes in the hydrological cycle due to climate change, clear-felling and subsequent land use. *Accepted 8 July 1999.*

Key words: *Amazonia, global change, hydrology, carbon sink, biogeochemistry, climatology, rain forest.*

INTRODUCTION

The impacts of land-use changes in the tropics have only recently become a significant element in the discussion of climate issues, even though people and nations have been using, exploiting and degrading natural resources for thousands of years in order to satisfy their needs. In the course of this century, the global system has become exposed to enormous stress due to a combination of population increases, rising standards of living and technological advances, as a result of which the term "environment" has acquired an entirely new dimension. We are now entering an age in which global ecology and economy are extremely closely related. Systems can no longer be regarded in isolation, because the global equilibrium has been fundamentally disturbed. Large biotopes – e.g., the tropical rain forests – can have a stabilizing effect, while the significance for the planet as a whole of extensive natural forests such as those in Amazonia has not yet been established. However, global change can greatly affect the composition, growth, productivity and distribution of forest ecosystems.

The earth's climate is the result of a complex interplay between incoming solar radiation, thermal radiation emitted by the earth, characteristics of the

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planet's surface (land, ocean, vegetation, snow, ice) and the atmosphere. During the earth's 4.5-billion-year history, climate has varied and changed on a wide range of time scales due to natural causes, with human activities having no impact. These climate variations have caused extreme hardships and sometimes catastrophes for evolving ecosystems as well as, more recently, for human populations. During the past century the earth's climate has warmed by 0.3–0.6°C, while atmospheric CO_2 concentration has increased by 25% to over 350 ppmv. This increase in carbon dioxide is consistent with anthropogenic emissions. The warming of the past century may be attributable to increasing greenhouse-gas concentrations resulting from human activities, but it might also be due to natural variability. Humans are now playing an increasingly important role in determining the global habitability of our planet (McBean 1994).

Interactions between the atmosphere, oceans, land masses and biosphere play a central role in the determination of climate and climate variability. In order to gain a comprehensive understanding of the climate system, it is essential that each of these components be observed, understood and modeled in terms of its impacts upon and extent of control by the others (Fig. 1). It is now obvious that large-scale interactions between the tropical land surface and the atmosphere are key physical processes contributing to climate variability. Changes in land surface characteristics (vegetation cover, soil moisture, surface roughness), whether natural or human-induced, generate changes in the surface radiation balance and the fluxes of heat and moisture, which in turn can exert a pronounced effect on the variability of the lower atmosphere. In turn, atmospheric changes can exert feedback on surface processes, resulting in further changes in the surface hydrological regime, biological systems and chemical cycles (Salati 1985).

When the surface of the earth is altered, the most profound impact on climate is produced by the deforestation of extensive forest areas and the afforestation of previously unwooded regions, since these cause the greatest change in surface characteristics (Graßl 1990). The vegetation cover (or lack of it) has a strong influence on the amount of solar radiative heating absorbed by the land surface, due to variation in albedo (reflectivity). In addition to warming the soil, heat absorbed by the surface provides energy for evaporation and for heating the atmosphere directly (sensible heat). Thus changes in albedo can strongly affect evaporation and atmospheric heating, thus influencing the hydrological cycle and atmospheric circulation. Other aspects of vegetation cover,

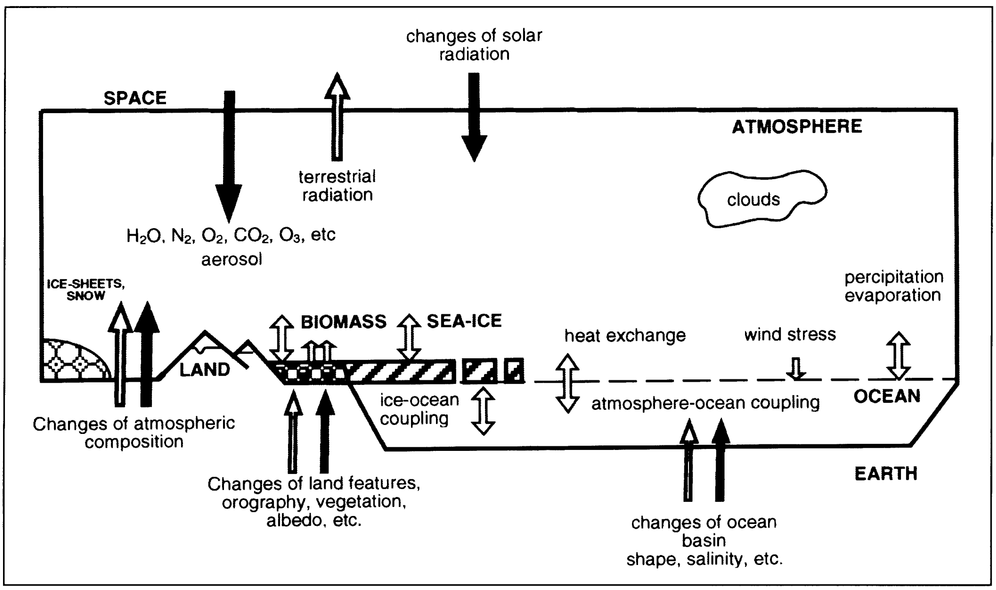

FIG. 1. Schematic illustration of components and interactions of the climate system (from Houghton 1984).

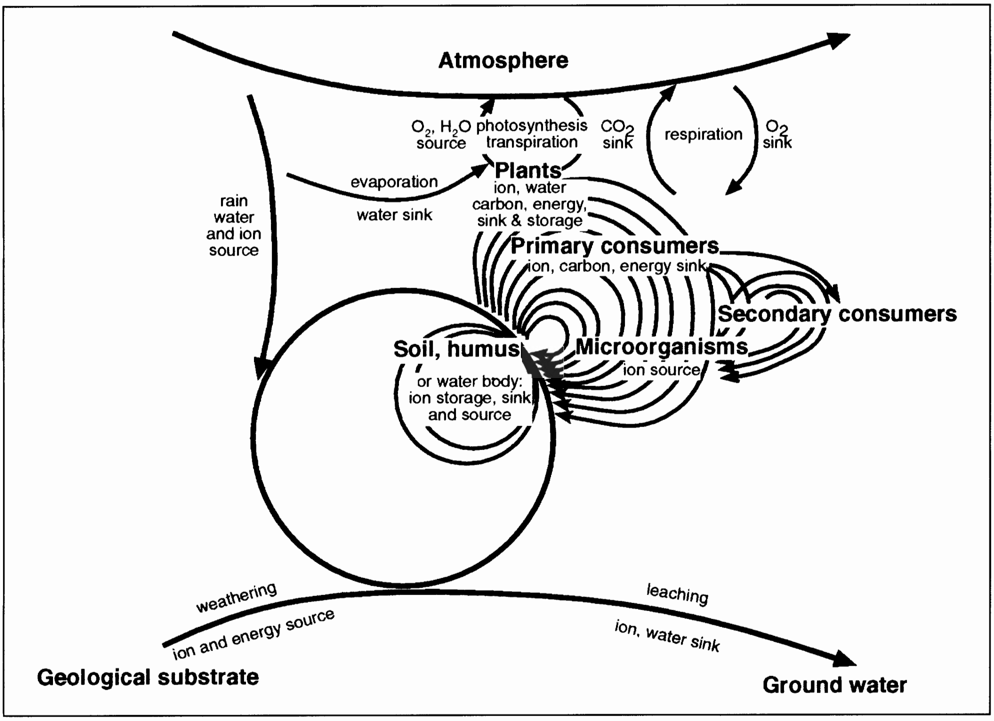

FIG. 2. Schematic representation of the atmospheric and biogeochemical cycles in ecosystems and the sink/ source relationships of the abiotic and biotic world in cycling energy, water, and mineral ions (Schulze & Zwölfer 1987).

such as aerodynamic roughness, stomatal resistance, canopy moisture capacity and rooting depth can affect the partitioning of incoming solar radiation between evaporation and sensible heat. The present state of knowledge on atmosphere-vegetation-soil interactions, linked by the water cycle and gas exchange, are shown in Fig. 2.

Of the three tropical climate zones where vegetation changes may have significant impacts on climate, i.e. the tropical rain forest, semi-arid areas and savannah, the tropical rain forest has received considerable attention. Here, evaporation rates are among the highest of any vegetation type, surface roughness is highest and, except for open ocean, albedo is lowest. Human-induced changes in the vegetation cover of small areas may have purely local impacts, but in the case of large areas the result may be major changes in the regional climate, which in turn may impinge upon regions remote from the area of change. Possible effects of global climate change on two different, well-studied ecosystems is shown in Fig. 3.

Trees and forests can influence global climate change, mitigate the effects of pollutants and greenhouse gases, and have positive feedbacks to the environment. Uncertainties about the effects of multiple interacting stresses on trees and forests can only be solved by an interdisciplinary research effort dedicated to understanding not only these interactions but also the exchange and feedbacks of forest ecosystems with the surrounding environment. Moreover, research into the impacts of climate change on forests must be conducted at multiple scales, including cell, leaf, tree, community and ecosystem levels.

Due to the sheer scale of land-use change in the Amazon Basin, the region has become a focal point of scientific interest in the fields of hydrometeorology, biogeochemistry, and ecology, leading to a consensus within the earth sciences community that an

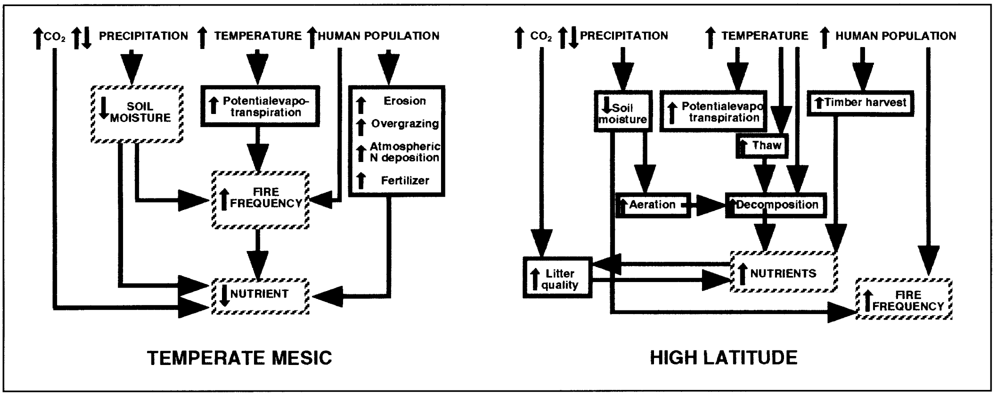

FIG. 3. Effect of projected global changes in climate and land use on resources and disturbance regime of a) temperate mesic ecosystems and b) high latitude ecosystems (Hobbie *et al.* 1994).

international, multidisciplinary research effort, initially built around comprehensive field experiments, should be put in place with a view to assessing the impact of deforestation on the local and global climate.

In 1990, the German parliament's Enquete Commission on "Protecting the Earth's Atmosphere" published a comprehensive report on "Protecting the Tropical Forests – A High-Priority International Task" (Deutscher Bundestag 1990). This report presents a detailed description of the present state of research regarding the climatic aspects of tropical forest ecosystems, as well as the causes and impacts of deforestation. Even now, almost a decade later, the Commission's recommendations for research and action to protect tropical forests and its extensive reference list have lost none of their pertinence.

The following sections present the current state of scientific understanding regarding the relationship between Amazonian rain forest deforestation and climate change, as well as state-of-the-art climate system modeling.

EMISSION OF GREENHOUSE GASES

Deforestation of tropical forests and subsequent use of the cleared area in various ways is disturbing the exchange of trace gases between atmosphere and biosphere in these regions.

Tropical forests and their ecosystems emit and take up numerous trace gases, thus contributing significantly to the global budget of these gases and to the chemical composition of the earth's atmosphere. These gases, e.g., CH_4, CO, N_2O and other nitrous oxides, have a direct effect on climate or are transformed by chemical reaction into secondary products of environmental and climatic relevance.

In South America, 5,660,000 hectares of tropical forest are converted annually by cutting and burning – the amount of burned biomass being 271,000,000 t year^{-1} (Goldammer 1993). The upper limit of forest and bush areas in degraded and/or seasonal forest formations potentially affected by fire each year, through which ground fires run regularly or which are subjected to a burning cycle of 15 years, is significantly higher at 81,023,000 hectares – or 460,000,000 t year^{-1} burned biomass (Goldammer 1993). Tropical forest clearing and burning in Brazil has been estimated to produce a net carbon input to the atmosphere of 0.3–0.5 Gt C year^{-1}, or approximately 30% of the global biotic net flux. Fig. 4 shows the deforestation of Rondonia as an example of tropical deforestation.

Biomass burning is now recognized as a major source of important trace gases – including CO_2, NO_2, CO and CH_4 – and aerosol particles. It takes many forms: burning of forested areas for land clearing, extensive burning of grasslands and savannahs to sustain grazing lands, burning of harvest debris, and use of biomass fuel for heating.

Emissions from biomass burning constitute a major perturbation of global atmospheric chemistry, especially in the tropics. Here, satellite observations have shown high levels of O_3 and CO over vast areas of Africa, South America, and the tropical Atlantic and Indian Oceans. Recent studies have linked this phenomenon to biomass burning plumes, and demonstrate that pyrogenic emissions affect regional ozone concentrations and the oxidative characteristics of the tropical atmosphere. Particulates affect regional global radiation budgets on account of their light-scattering effects and their influence on cloud microphysical processes (Andreae *et al.* 1994). The impacts of deforestation by fire will be discussed in the section on "Edge Effects".

The emissions of greenhouse gases and smoke particles from biomass burning that are of crucial significance for biogeochemical cycles and climate-forming processes are presented in the following sections.

THE PHOTOCHEMICAL OXIDANT CYCLE

The hydroxyl radical, OH, is the primary oxidant in the troposphere responsible for the removal of major pollutants released into the atmosphere by anthropogenic and natural processes. For example, carbon monoxide is converted to carbon dioxide, sulfur dioxide to sulfuric acid, nitrogen oxide to nitric acid, and methane to carbon monoxide. Thus, OH plays a key role in determining the lifetime of trace species in the atmosphere and whether the environmental impacts of the species emitted (or produced) will be confined to source regions or transported and distributed over global scales. Gases which do not react with OH have very long lifetimes and are transported into the stratosphere, where they can be broken down chemically. In the case of species like N_2O and the fully halogenated chlorofluorocarbons, the chemical destruction processes in the stratosphere result in the depletion of the ozone layer.

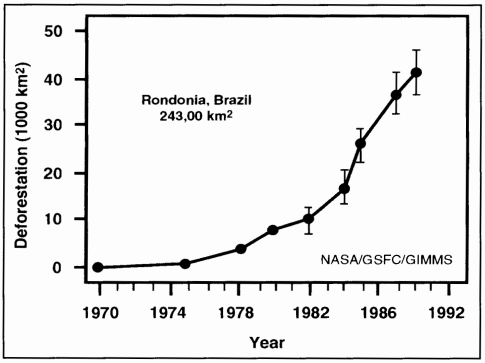

FIG. 4. Area deforested in the state of Rondonia against time over the last two decades (Crutzen & Carmichael 1993).

The OH radical is produced by photolysis of ozone through absorption of solar ultraviolet radiation, followed by the reaction of the electronically excited oxygen atom "O" with water vapor

$$O + H_2O = 2OH.$$

In the background troposphere, prevailing concentrations of OH are governed by the photochemical oxidant cycle, which is controlled by temperature, water vapor, the intensity of ultraviolet radiation, and the abundance of carbon monoxide, methane, and nitric oxide (Crutzen & Carmichael 1993). The key photochemical processes are shown in Fig. 5.

CLIMATE IMPACTS OF SMOKE AEROSOL

Near source regions, smoke will be optically dense, reducing reflectivity. The direct radiative effects of smoke aerosol from biomass burning are of some significance for climate, and the indirect effects through clouds are possibly of equal significance. Together, they may increase the global reflection of solar radiation by up to 2 Wm^{-2}, comparable to that for sulfate aerosol, and each about as large as the present-day warming from greenhouse gases of 2.4 Wm^{-2}. It is likely that some aspects of the estimated cooling from aerosols are now exaggerated, since otherwise a substantial global cooling should have been observed over the last century. It is difficult to reduce the estimated direct effects by more than a factor of two, but indirect effects could be uncertain by at least twice as much. It seems likely that the net cooling from aerosol has been at least half that of the current greenhouse warming (Dickinson 1993).

During the fire season, aerosol concentrations in the entire Amazon Basin are very high. Far from burning plumes, the mass concentration of inhalable particles exceeds 300 $\mu g\ m^{-3}$. Large amounts of fine particles are injected into the atmosphere, where they travel over long distances. The elements emitted are K, Ca, S, Cl, P, Si, Al, Mg, Fe, Mn, Ti, Zn, Cu, Ni, Pb, V, Cr, Br, Rb, and Sr. The emission of trace elements and heavy metals into the global atmosphere due to biomass burning can be very high. Data from continuous ground-based aerosol monitoring in Cuiabá indicate substantial levels of biomass-burning

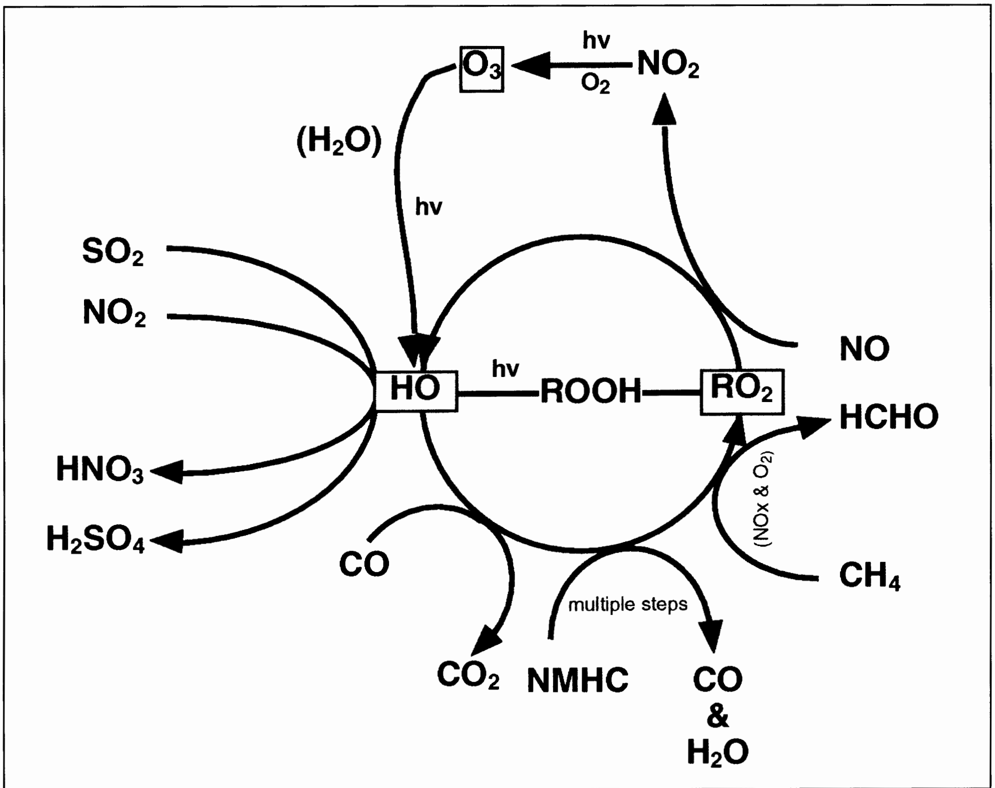

FIG. 5. Schematic representation of the photochemical oxidant cycle. In this figure, R represents H, CH_3 or higher carbon-number fragments.

aerosol far from the source regions. Aircraft measurements show an average atmospheric aerosol loading of more than 350 µg m^{-3} at 1000 m altitude.

In September 1991, the INPE Bandeirante aircraft collected aerosol samples at an altitude of 1000–3000 m on flights covering the Amazon Basin and Cerrado region. The collected aerosol particles were "aged" biomass-burning emissions, and represent an average composition on a basin-wide scale. Aerosol mass concentration varied from 30 mg m^{-3}, for areas little affected by biomass-burning, to more than 300 mg m^{-3} for regions with intense burning. These are average concentrations for large areas where no impact from any particular plume is evident. Visibility in the Amazon Basin as a whole was typically 500–1000 m; in areas with very low visibility the figure was only 200 m (Artaxo *et al.* 1993).

Several essential plant nutrients (such as P, S, N, and others) are transported in large amounts in the atmosphere as a result of biomass burning processes. Smoke plumes in South America show very high concentrations of sulfur and soot carbon, which affect cloud formation mechanisms and the radiation balance of the Amazon atmosphere (Fishman *et al.* 1993). Most of the aerosol particles are water-soluble and can be active as cloud condensation nuclei, having the potential to change the cloud formation mechanisms in the Amazon Basin and other regions of the planet (Artaxo *et al.* 1993).

EMISSIONS OF NITROUS OXIDES AND THEIR IMPACTS ON THE OZONE CONCENTRATION

In recent decades the tropospheric ozone concentration in the tropics has increased sharply during the dry season – the time of forest fires. Over large areas of Central Africa and tropical Latin America it has by now reached values comparable to those observed over polluted regions in the Northern Hemisphere. In Brazil, ozone concentrations of 50 to 60 ppb are measured even near the ground and values of 60 to 70 ppb in the middle troposphere. Over undisturbed tropical forests, tropospheric ozone concentrations are very low, often less than 10ppb. These low values are explained by dry deposition of ozone in tropical forests and by the destruction of ozone in a chemical reaction with isoprene, a hydrocarbon emitted by tropical forests. Ozone is toxic for humans, animals and plants. However, the impacts of increased ozone concentrations on vegetation have been little studied so far. By examining physiological changes (gas exchange, water-use efficiency), Grulke & Miller (1994) succeeded in identifying an age-dependent sensitivity of the *Sequoiadendron giganteum* redwood to increased ozone levels (up to 10 times), with sprouts up to the age of 5 years showing a greater sensitivity. If these results also apply to juvenile plants of other species, as is very likely, regeneration of a forest after deforestation by fire appears questionable.

In an extended review, Vitousek *et al.* (1997) discuss sources and consequences of human alterations of the global nitrogen cycle. They state that nitrogen inputs serve human needs such as agricultural production, while environmental consequences are considered to be serious and long term.

Ozone is produced in the troposphere by photochemical disintegration of CO, CH_4 and other hydrocarbons, in the presence of sufficiently high NO_x concentrations. In general, NO_x stands for NO and NO_2, with a ratio of 85:15 in the smoke of forest fires, because NO_2 is much less stable than NO. If NO_x concentrations are lower than ca. 5–10 ppt, ozone will be broken down by the same compounds. Thus, nitrous oxides are highly significant for the formation of ozone in the troposphere.

In the tropics, nitrous oxides are produced in large amounts by biomass burning during forest and savannah fires, as well as by lightning discharge in the atmosphere. NO_x emissions by tropical biomass burning are estimated to contribute about 20% to global NO_x emissions, almost one quarter of that portion resulting from deforestation by fire, the remainder from savannah fires. In addition, atmospheric NO_x concentrations are further enhanced since there are less plants available for NO_x fixation due to deforestation. According to initial assessments, pristine tropical forests take up the greatest part of natural NO_x emissions from soils – in the order of 4 million tons. Amounts of NO_x emissions from lightning discharge are still not known. Estimates range from 3–8, to as much as 100 million tons per year nitrogen in the form of NO_x. Downcurrents of thunderstorms can even cause this to reach the lower troposphere.

Under natural conditions in the biosphere, N_2O is produced mainly in soils, where it is also consumed simultaneously. Emissions are therefore low. For soils in tropical forests, N_2O emission rates of about 40 $\mu g\ m^{-2}$ have been measured, with a total emission of N_2O nitrogen from natural soils (tropical and temperate regions) of approx. 6 Tg N $year^{-1}$.

Production and destruction of N_2O occur within the metabolic pathways of denitrification, nitrification, nitrate dissimilation, and nitrate assimilation. Nitrous oxide is produced and consumed not only by single microbial species, but also by microbial teams. It is usually reduced by micro-organisms that use it as an electron acceptor when oxygen or nitrate is not readily available, that is, under anaerobic conditions. On the other hand, N_2O may be oxidized by reactions involving soil catalase activity. Hence, there are many pathways for the production and destruction of N_2O, processes which are affected by various environmental parameters such as temperature, moisture, redox potential, organic carbon content, pH, nitrate content, the presence of plants, carbonate content, and others. The N_2O flux at the soil-atmosphere interface also depends on the location of the N_2O-producing and N_2O-consuming micro-organisms and their relative activity within the soil column. But very little is known about the flux rates between uncultivated soils and the atmosphere (Seiler & Conrad 1987).

METHANE EMISSIONS

Biospheric methane release is now attributable primarily to human activities, with a major proportion, perhaps 80%, ensuing from agricultural activities in the tropics – such as biomass burning, domestic cattle husbandry and, perhaps even more important, from rice production in the paddy fields of East Asia. Methane is now increasing in the atmosphere at a rate of 1–1.5% annually. With a projected annual population increase of almost 3% in many such areas, a continuing increase in atmospheric methane is probable.

The global distribution of sources and sinks of methane indicates dominant sources in the tropics.

Global methane destruction by reaction with OH is estimated to be 320 Tg $year^{-1}$. The total mass of CH_4 in the atmosphere is around 500 Tg. For methane to increase at the observed average rate of 1.2% per year, there must be a total input of about 380 Tg CH_4 per year. Several sources can be confidently estimated. E.g. enteric fermentation in domestic ruminants, especially cattle, produces about 80 Tg CH_4 $year^{-1}$, with less-developed and industrialized countries accounting for similar portions. Coal mining produces 34 Tg CH_4 per year worldwide, given emissions of 18–19 m^3 methane per ton of coal. With a CH_4-to-CO_2 emissions ratio of between 0.8 and 1.6% and the previous estimates of carbon burned per year, biomass burning in the tropics produces 20–70 Tg CH_4 annually. An average of 45 Tg is adopted here. Leakage of natural gas from distribution systems could release an annual 33 Tg, if the 4% estimated for Switzerland is assumed globally. Digestion of plant material by termites is a minor source of methane. The annual release rate from rice fields is between 20–150 Tg CH_4, with a median value of 60 Tg CH_4. The total area of marshes and swamps in the world equals about 2 x 10,000,000 km^2, of which 75% is located in the tropics. The tropical natural wetland area is roughly equal to the cultivated rice paddy area. Organic matter production in tropical wetlands is 4 kg m^{-2} $year^{-1}$, a rate that extrapolates to an annual tropical wetland organic matter production of 6,000 Tg, more than 80% of the total from all marshes and swamps in the world. A methane yield of about 1.5% from the decay of this material would, therefore, yield the required methane flux of 90 Tg $year^{-1}$. Such a low efficiency of CH_4 production indicates that many variables must influence the release of CH_4 from different ecosystems and that results from global extrapolation of measurements at a handful of sites will probably always remain extremely uncertain. Crutzen (1987) puts the current annual anthopropogenic production of methane at 290 Tg and natural production at about 90 Tg which yields to the above mentioned yearly amount of 380 Tg.

Aerated soils were long considered to be an insignificant source of atmospheric CH_4. Field measurements carried out in different ecosystems now show that atmospheric CH_4 is actually deposited in aerated soils and that deposition may be a significant sink for atmospheric CH_4. The processes responsible for the observed deposition of atmospheric CH_4 are still unknown. They are aerobic, requiring a well-aerated soil. A change from CH_4 emission to CH_4 deposition has been observed when a flooded soil slowly dried up. Additional CH_4 uptake occurs in tropical forests, which have an overall area of about 24 x 10,000,000 km^2. If we assume that the CH_4 loss observed in the Amazon Basin applies to the humid tropics in general, the uptake rate for the overall area is about 25 ± 12 Tg $year^{-1}$. An additional 7 ± 3 Tg $year^{-1}$ is oxidized at higher latitudes, so the global CH_4 consumption by soils may account for 32 ± 16 Tg $year^{-1}$, about 10% of the oxidation rate by reaction with OH (Seiler & Conrad 1987). Conversion of forest to cattle pasture resulted in a diminution of the methane sink, because soil compaction by cattle leading to poor soil aeration changes the balance between methane consumers and producers. In poorly drained soils, methane production may reach significant levels, with a net methane sink possibly becoming a net methane source.

The reactions that take place during the oxidation of methane in the atmosphere are particularly important in photochemistry and will be therefore discussed here.

The oxidation of methane starts by reaction with OH. Subsequent oxidation steps and products depend decisively on the availability of nitric oxide. If sufficient NO is present, the oxidation pathway to carbon monoxide yields an average net gain of about 2.7 ozone molecules and 0.5 hydroxyl radicals per methane molecule oxidized. The oxidation of CO to CO_2 produces another ozone molecule without affecting the total odd-hydrogen concentration.

The situation is very different in environments with low NO volume mixing ratios. The oxidation of CH_4 to CO leads under such conditions to an average net loss of about 3.5 odd-hydrogen and 0.7 ozone molecules for each reacted CH4 molecule. The further oxidation of CO to CO_2 leads to the loss of one additional ozone molecule without affecting odd hydrogen.

THE ROLE OF THE TROPICS IN CHANGING INTERACTIONS BETWEEN OH, CH_4, NO, AND O_3

The tropics are very important for the photochemistry of the atmosphere and for the cycling and removal of many important atmospheric trace gases. This importance is due to the efficient production of hydroxyl radicals, the "cleaners" of the atmosphere. Hydroxyl radicals are increasingly being removed in the tropical atmosphere by reactions with methane and methane oxidation products. In these environments, the loss rate of methane from the atmosphere is no longer linearly dependent on its atmospheric concentrations. A continuing growth in atmospheric CH_4 is expected to lead to a decrease in global average hydroxyl radical concentrations and to an increase in the concentration of the many gases removed predominantly by reaction with OH. On the other hand, in parts of the atmosphere containing supercritical NO concentrations, such as the upper troposphere, lower stratosphere and the continental boundary layer of the industrial mid-latitudes, as well as parts of the tropics, there will be more ozone production. These changes in methane, hydroxyl radical and tropospheric ozone concentrations could clearly have major impacts on the future composition and photochemical functioning of the earth's atmosphere and on global climate.

The production and atmospheric release of CH_4 now seems to originate mostly from land clearing and agriculture activities in the tropics, especially increasing biomass burning and rice production in the paddy fields of East Asia. With an expected annual population growth rate of 2.2% in the developing world, indeed 2.5–3% in Latin America and almost 3% in East Asia, atmospheric CH_4 content is likely to show continued growth, currently at 1–1.5% $year^{-1}$. The consequences of this increase for the photochemistry of the earth's atmosphere should receive considerable attention (Crutzen 1987). Studies of sources and sinks of methane and many other important trace gases should have high priority.

CARBON MONOXIDE EMISSIONS

Carbon monoxide is emitted in the tropics from forest and savannah fires. Small amounts are generated as direct emissions from forests. In addition, CO is produced by the oxidation of non-methane hydrocarbons emitted from forests (especially isoprene). CO production of around 1,011 molecules cm^{-2} s^{-1} from sub-tropical forest leaves has been measured. Given high leaf area indices and a large UV flux at the equatorial site investigated by Kirchhoff & Marinho (1990), it is plausible that direct plant emissions account for the bulk of the implied source they measured. Emissions of this magnitude would yield about 35 Tg CO $year^{-1}$ for the Amazon Basin. Recent studies demonstrated that substantial sources arise either from oxidation of biogenic hydrocarbons, or from metabolic processes in the forest. These non-combustion sources are little understood in quantitative terms; they operate in every season, and may play a major role as regards the concentrations of CO in the Amazon region, and in the tropical region globally. CO emissions from tropical biomass burning are estimated to contribute about 30 to 40% to worldwide CO emissions. Direct CO emissions from tropical plants may amount to 5–6% of total emissions, and CO from hydrocarbon oxidation to 15–20%.

The calculated global budget of sources and sinks of carbon monoxide contains some well-known quantities, such as the amount resulting from methane oxidation, which can be estimated from the calculated OH and observed CH_4 distribution, the transport of CO from the Northern Hemisphere to the Southern Hemisphere, the uptake of CO by soils, and the oxidation of CO to CO_2 by reaction with OH, which is likewise estimated from the calculated global distribution of OH and the observed distribution of carbon monoxide. Analysis of these production and destruction terms indicates the need for an additional large source of CO in the tropics equal to about two thirds of the total source strength.

Measurements of carbon monoxide in the boundary layer to 3 km altitude over the Amazon forests (near Manaus) in August 1980 show average carbon monoxide volume mixing ratios of about 300 ppbv, indicating the existence of a large source of CO, since Atlantic and free tropospheric volume mixing ratios at the same latitudes are about 100 ppb. A trajectory analysis indicated a travel time of about two days for boundary layer air between the Atlantic and Manaus. Most prominent among the hydrocarbons emitted from forest vegetation is isoprene, but other organics are also present in significant amounts. The oxidation reactions of the natural hydrocarbons are not known precisely and depend, in turn, on the availability of NO, about which there is very little observational knowledge (Crutzen 1987).

CARBON DIOXIDE EMISSIONS

Even though CO_2 emissions from extensive forest burning in Amazonia account for only 10-30% of total emissions, they have a noticeable impact on the atmosphere and thus on global and especially regional climate. Subsequent to cutting down an area of tropical forest, CO_2 is released primarily from burning and/or decay of biomass. Additional CO_2 comes from oxidation of organic carbon fixed in the soil.

In the case of tropical deforestation by fire, only about 30% of the above-surface biomass is burned directly. The remainder decays in the following years or is burned year by year along with agricultural waste produced in these areas. Subsurface biomass, mostly tree roots, is transformed almost exclusively by microbial processes.

Use of wood for domestic fires and as construction material is another source of carbon dioxide emissions in tropical forests. Carbon fixed in firewood is quickly released into the atmosphere as CO_2. In the production of construction material, almost half of the felled biomass is left behind in the forest; about 10 to 15% of it decays per year. Carbon contained in construction material is generally fixed for longer time spans. However, the portion of construction material in the total biomass destroyed is very small. If the wood is processed as paper, this carbon will also be emitted to the atmosphere very quickly. Carbon is released as CH_4 from paper deposited in waste disposal sites. Carbon released from deforestation by fire and from decomposition can be taken up again by a secondary forest within 30 years (Curran *et al.* 1995).

CARBON FIXATION BY VEGETATION

The CO_2 sources listed in the previous section are juxtaposed essentially by four potential CO_2 sinks, namely:

- fixation of CO_2 by successor vegetation after deforestation;
- formation of graphite carbon from biomass burning;
- increased CO_2 uptake by vegetation due to a possible CO_2 fertilizer effect;
- storage in construction material.

The terrestrial biosphere (vegetation and soils) could remove at most a third of the CO_2 emitted if all fossil fuel reserves were burned (Tans & Bakwin 1995). In the context of CO_2 fixation, it appears important to distinguish between recent carbon concentrations and future ones. Under present conditions, vegetation in general (Esser 1990) as well as tropical rain forests can be regarded as carbon dioxide buffers. For example, in undisturbed rain forest areas of Rondonia, Grace *et al.* (1995A) found an average CO_2 uptake resulting in an average CO_2 fixation by the vegetation of 0.05 mol m^{-2} and day. A former assumption that Amazonia's tropical rain forest is in an exactly balanced steady state was not confirmed by these results. Thus, under present conditions, 11 mol C m^{-2} $year^{-1}$, or 1.32 tons C ha^{-1} $year^{-1}$ are being fixed. Long-term monitoring of plots in mature humid tropical forests concentrated in South America revealed that biomass gain by tree growth exceeded losses from tree death. The data suggest the Neotropical forests may have taken up as much as 0.4 Pg C $year^{-1}$ during 1975–96 (Phillips *et al.* 1998). This is an essential factor in the hitherto unbalanced carbon balance calculations, and due to this buffer capacity the tropical rain forest (primary forest) acquires great significance for the global climate. However, Prentice & Lloyd (1998) pointed out that climate variability causes large swings, so that in El Niño years these ecosystems can become a source of carbon release into the atmosphere. Additional information on how climate variations like El Niño affect terrestrial carbon storage are given by Tian *et al.* (1998), pointing out the significance of soil moisture as an important control of carbon storage.

However, with regard to the fertilizer effect resulting from enhanced CO_2 concentrations, as described by the Second Report of the Enquete Commission in 1990, certain distinctions need to be made. According to recent publications, it is important to distinguish between present and future CO_2 concentrations. So far, vegetation was able to buffer part of the carbon dioxide production during the time span from the beginning of the industrial revolution to today. This, however, is unlikely to happen again, given the recent increase in atmospheric CO_2 concentrations.

There are several reasons: the earth's forest cover has decreased, so long-term CO_2 fixation by scrub vegetation as successor vegetation to forests has no significance, as is the case with any CO_2 fertilizer effect on agricultural plants, since the carbon fixed will be released again during the course of the year. In boreal latitudes, 50% of the existing forests are already damaged (Second Report of the Enquete Commission 1994), and forest areas in tropical America have decreased by 42% (Jackson 1983). The Enquete Report cited above forecasts a further reduction of 50% by the year 2025. Only the reduction of this storage potential might suffice to call any kind of carbon dioxide buffering into question. Furthermore, the expected future CO_2 concentration, which will be almost twice as high as today, makes a CO_2 fertilizer effect on vegetation appear questionable. For physiological reasons, there is little likelihood that forest vegetation will react with increased growth.

Applying future climate conditions, including increased CO_2 concentrations, to *Pinus ponderosa* in a field experiment, De Lucia *et al.* (1994) found no indications of enhanced growth and productivity. For *Castanea sativa*, El Kohen & Mousseau (1994) describe that an increase of the atmospheric CO_2 supply up to 700 µmol mol^{-1} had no positive effect on leaf area and shoot growth without additional mineral fertilization of the soil. However, total biomass production increased by 20%, indicating a shift in the root-shoot ratio. Those test plants that did not receive additional mineral fertilization displayed a decrease in nitrogen and chlorophyll content together with an accumulation of starch in the course of the experiment. Thomas *et al.* (1994) also found a positive effect on photosynthesis capacity under enhanced CO_2 conditions only if N and P fertilizer is supplied in addition. A shift in the biomass ratio between root and sprout of *Prunus avium* under increased CO_2 conditions with additional mineral fertilization is also described by Wilkins *et al.* (1994), while high CO_2 supply without fertilization had no effect on biomass production. Although model calculations by Melillo *et al.* (1993) suggest increased net primary production by tropical ecosystems under enhanced CO_2 conditions, Körner & Arnone (1992) reported no significant changes in leaf area index or water and nitrogen consumption, or in stomatal behavior of understory and overstory plants of the humid tropics under 2 x CO_2 concentrations.

In studies on *Quercus petraea* and *Pinus pinaster*, Guehl *et al.* (1994) demonstrated a direct relationship between the CO_2 fertilizer effect (CO_2 concentrations of 700 µmol mol^{-1}) and the plant's water supply. Under unfavorable watering conditions, the increased biomass production of *Q. petraea* is reduced to less than one third, and *P. pinaster* shows no increase at all in biomass production.

Overdiek & Forstreuter (1994), however, report reduced evapotranspiration and an improved transpiration coefficient (transpiration/net CO_2 uptake) for *Fagus sylvatica*, i.e., with a doubling of atmospheric CO_2 concentration, water loss from evapotranspiration decreases (up to 14% at temperatures between 14° and 25°C). Baldocchi (1994) also describes a reduction in stomatal conductivity but no decrease in photosynthesis activity if the CO_2 supply is increased. In a model simulation, a temperature increase of 5°C and a CO_2 concentration of 660 ppm lead after 100 years to a reduction in soil water in a forest ecosystem dominated by conifers (Väisänen *et al.* 1994).

In summary, only very few of the investigated tree species increased their biomass production, thus fixing CO_2, if the atmospheric CO_2 supply is doubled. It seems instead that, over the long term, atmospheric CO_2 concentrations of around 700 µmol mol^{-1} will only have a positive effect on total biomass production if at the same time mineral fertilizer is supplied and watering conditions are favorable. In this context, the increase of the root/sprout ratio under increased CO_2 conditions should be emphasized. This, in turn, leads to a shrinkage in leaf area, i.e. the percentage of photosynthesizing organs decreases and the initially positive effect is negated. For complex ecosystems like the rain forest, Körner (1994) summarized: "The more complex and stable a system is, the closer its long-term carbon gain approximates zero and the greater the relative biological effectiveness of CO_2 enhancement initially will be (while the absolute gain may still be small)."

In the future (with CO_2 concentrations of approx. 700 µmol mol^{-1}), the remaining areas of tropical rain forest, especially in Amazonia, cannot be considered as a CO_2 buffer for the reasons stated above, unless they are fertilized. This still leaves the question open as to what the long-term effect of fertilization – mineral and/or CO_2 – would be.

The best estimate of potential sequestering from all processes in the closed forest landscape is 1.5–3.2 Pg C year^{-1}, or about 31–58% of the current CO_2 emissions generated by the combustion of fossil fuels. Most of the potential carbon sequestering is due to biomass accumulation in recovering forests (72%), followed by accumulation in soils (11–17%) and in coarse woody debris (11–12%) (Lugo & Brown 1992).

Soils in tropical rain forests can store carbon for long periods of time. Schlesinger (1990) confirmed this trend when he showed that soils can continue to accumulate "Soil Organic C" over thousands of years. Large losses of Soil Organic C are associated with permanent agriculture, but not with short-term use by shifting agriculture. Soils recover their Soil Organic C following abandonment of agriculture to forest succession, with the rate depending upon life zone. Forests in moist or wet life zones recover Soil Organic C faster than in dry life zones, but all appear to reach levels of Soil Organic C approaching those of nearby forests in about the same length of time –

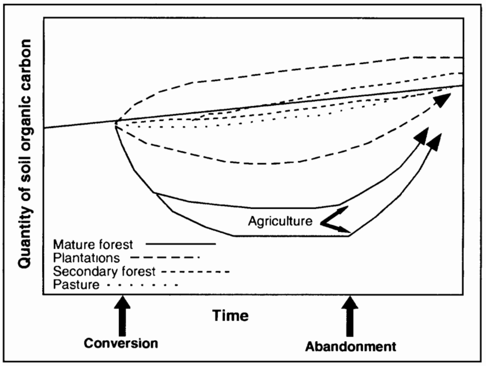

FIG. 6. Suggested patterns of change in soil organic carbon in soil subjected to different land uses in the tropics. This model ignores variations due to soil type and mechanical conversion of sites. The type of agricultural use or plantation species planted changes the slope of the curves (Lugo & Brown 1993).

about 50 years – (Brown *et al.* 1992). Conversion of forests to pastures often results in no loss or gain in Soil Organic C compared to nearby native forests. Furthermore, resampling of cultivated sites that had been converted to pastures showed increases in Soil Organic C over several decades. In other words, soils under pasture can be C sinks, regaining some of the C that was released during clearing. High rates of root production by grasses may explain why pastures accumulate Soil Organic C. Soil organic matter will recover under forest plantations at rates similar to or faster than secondary forests. Differences in rates of Soil Organic C accumulation are caused by differences in species and environmental factors. Some species produce more litter and roots than others, thus producing more organic inputs which eventually influence Soil Organic C (Fig. 8). Clearly, this species effect must be considered in plans for enhancing C sequestering (Brown *et al.* 1992).

From comparable Soil Organic C data from a variety of tropical land uses we estimate that tropical soil can accumulate between 168 and 553 Tg C year^{-1}. The greatest potential for carbon sequestration in tropical soils is in the forest fallows which cover some 250 million hectares. Increased attention to Soil Organic C by land managers can result in greater rates of carbon sequestration than predicted by current Soil Organic C models (Lugo & Brown 1993). On a global basis, particularly high values of Soil Organic C (> 300 Mg ha^{-1} to 1 m depth) are reported for forest stands on inceptisols in Venezuela, from lower montane sites in Papua New Guinea, and from wet sites in Colombia. Part of this variability represents differences among plant associations, part is related to soil type and mineralogy, part is related to location (e.g., aspect, microtopography, etc.), and part is climatically induced. As water availability increases, there is a greater diversity of possible plant and geomorphological associations, each with particular Soil Organic C content (Lugo & Brown 1993) (Fig. 6).

HYDROLOGICAL CYCLES

The global hydrological cycle is most intense in the tropics and subtropics, where it is characterized by enormous spatial variability.

Extensive land-use changes can affect climate and also the hydrological cycle; according to Salati (1987), the following climate-relevant parameters will be disturbed:

- energy balance on a microclimatic scale
- regional energy balance
- water balance on a microclimatic scale
- water balance in representative hydrographic basins
- water balance at the regional level (Amazon Basin)
- biogeochemical cycles on basin and regional scales
- interaction among water, terra firme forest, and flooded-land forest.

Since changes in climatic elements are mostly caused by changes in the hydrological cycle, or affect it directly, it will be dealt with in greater detail in the following section. Fig. 7 shows a summarized representation of the water cycle.

The water budget has been determined for a 25-km^2 model area near Manaus. Here, it was found that about 75% of precipitation is returned to the atmosphere as water vapor through evapotranspiration from plants. Of this 75%, an average 25% is directly intercepted by the plant surfaces and evaporated, while the approx. 50% remaining is transpired by the plants. Only about 1/4 of the total rainfall contributes to the surface runoff in rivers and streams (Fig. 8). However, for the entire Amazon Basin region, these values can only be calculated at present as averages from various measurement sites, and they clearly deviate from those found in the model study due to differences in vegetation types and soil characteristics (Salati 1985).

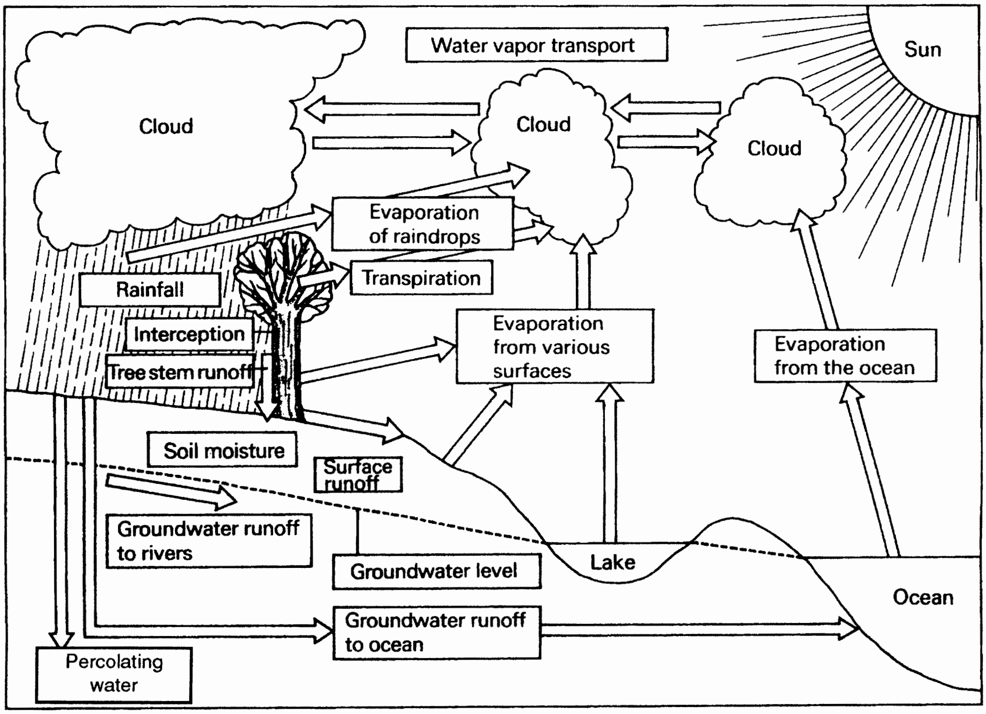

FIG. 7. Schematic representation of the water cycle (Enquete Report 1994).

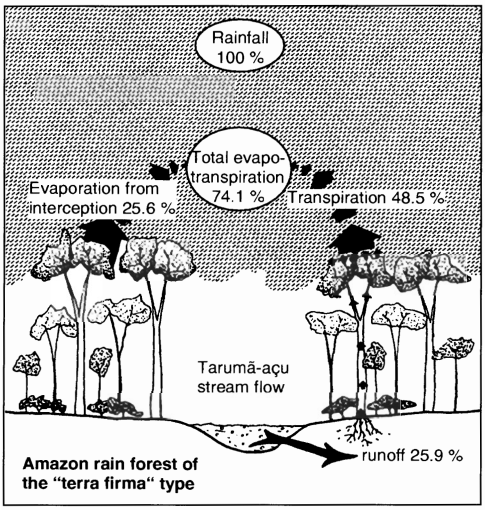

FIG. 8. Water balance of the model basin (Salati 1987).

The amount of precipitation captured by the plant surface and evaporated again is closely related to rainfall intensity and duration (see Fig. 9).

Due to the high portion of evaporation, the amount of precipitation in the entire Amazon Basin is very high, although the western parts are several thousand kilometers away from the ocean. According to model calculations, one water molecule is recycled up to eight times in the precipitation-evaporation cycle over the Amazonian rain forest, until it reaches the Andes with the westward-directed air current from the Atlantic.

Looking in greater detail at the water cycle within the rain forest, three internal cycles can be distinguished (Fig. 10). Cycles 2 and 3 are closely linked, while Cycle 1 in the scrub of the forest functions independently at high humidity, especially at night (Grace *et al.* 1995 B), and it depends more strongly on the morphology, anatomy and physiology of the vegetation rather than on precipitation directly.

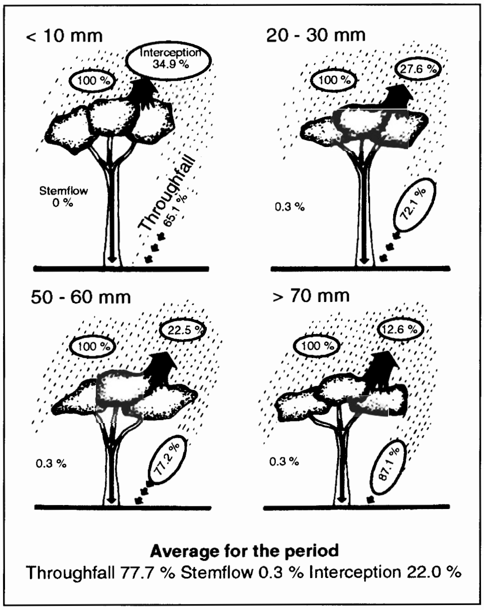

FIG. 9. Interception of rain by tree canopy in the model basin, as a function of intensity of precipitation (Salati 1987).

Sellers (1987) demonstrated that the energy and water balance of a region could be highly dependent on plant physiology and morphology: the former factor controls the transpiration rate, while the latter has a marked effect on the annual interception loss rate (that proportion of rainfall that is held on plant leaves as liquid water and then re-evaporates to the atmosphere without reaching the bulk soil moisture store). Plant structure (Fig. 11) affects the absorption of radiation in a region by the trapping of light in multiple reflections between plant elements (Sellers 1987).

Temperature, humidity and wind are three important climate factors that strongly depend on each other and on vegetation. Wind profiles in Amazonia's tropical rain forest compare well with those described for forests in temperate latitudes. Near-surface wind velocities are between 0 and 1 m s^{-1} and hence near the limit of measurability (Molion 1987).

Temperature and humidity are relatively stable throughout the year, with a daily amplitude in the forest clearly smaller than in the upper canopy level. On days with rain showers, temperature variations are greater than on purely rainy or sunny days. On dry days, temperatures in the forest vary between 25° and 32° C. On days with rain showers, temperatures vary during the day in the range of 21° to 27° C in the forest and 21° to 30° C in the upper canopy level. The water deficit of the air is 30 to 50 % in the upper level of the treetops, while near the surface it is nearly saturated, even on dry days.

Almost 75 % of the absorbed solar radiation is consumed by evapotranspiration, and only 25 % heats the air. The main significance of the forest, however, lies in water recycling (approx. 50 % of the precipitation evaporates and is returned to the atmosphere), leading to a continuous transport of vapor within the Amazon Basin. Therefore, the distribution of precipitation in the entire area of the Amazon is much more even than it would be without a forest (no forest: high precipitation near the coast, low precipitation in the interior regions). In summary, vegetation affects humidity, wind (surface roughness) and thus the stability of air stratification, air pressure, precipitation and the absorbed solar radiation.

INFLUENCE OF ALBEDO

Depending on the succession vegetation, extensive deforestation can cause changes in the hydrological cycle. Destruction of secondary forests would cause hardly any changes in the water cycle compared with a primary forest. If, however, tropical forest is converted into pasture, the impact on the water cycle is rather significant. Precipitation decreases, thus extending the dry season. During the dry season, evaporation is reduced to almost zero because there is hardly any water available. Pastures dry out, increasing local albedo, which, in turn, causes further

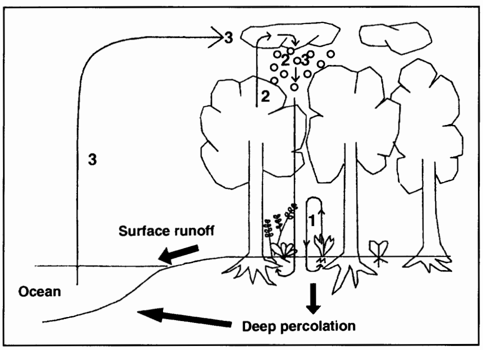

FIG. 10. Details of the hydrological cycle in a tropical rain forest (Lösch, pers. comm.).

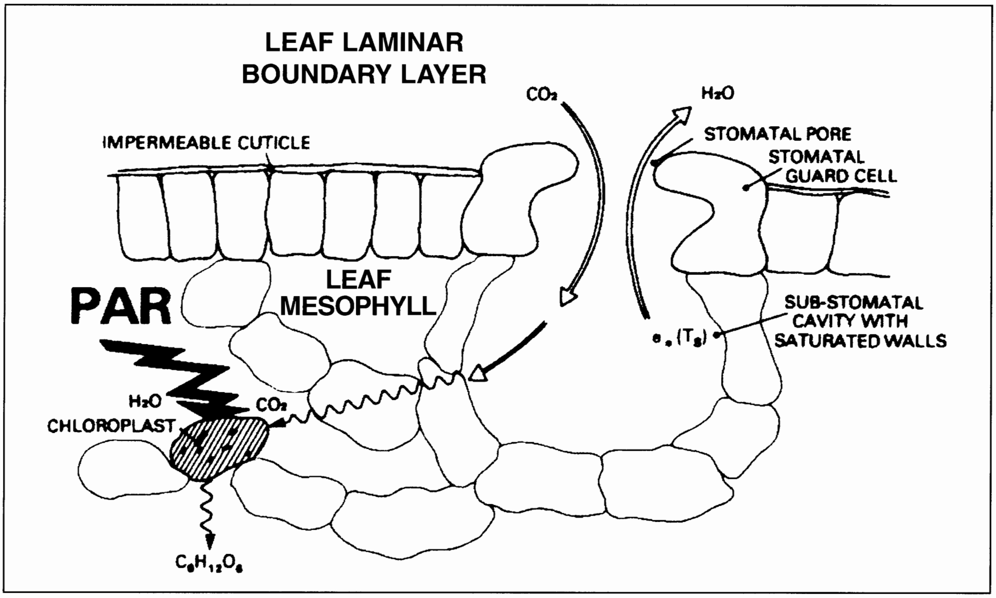

FIG. 11. Processes of gas exchange and photosynthetically active radiation (PAR) interception at the leaf level. Carbon dioxide diffuses in through the stomatal pore, the aperture of which is under active control. There is an inevitable loss of water vapor over the same route since the internal tissues are saturated. Carbon dioxide and cellular water are combined into organic molecules in the mesophyll chloroplasts. These have chlorophyllous structures that are highly absorbent in the 0.4–0.7 μm spectral region.

evaporation and less precipitation. With an albedo of approx. 13 %, tropical rain forests are the darkest natural land surface ("black hole" vegetation). Here, 87 % of the incoming solar radiation is absorbed by evaporation, heating of the forest and the lower atmosphere as well as by photosynthesis. Destruction of the tropical rain forest increases albedo. The amount of increase depends on the type of succession vegetation, e.g. a secondary forest has a slightly higher albedo. The species composition of the forest as well as the morphology and anatomy of the individual species have an effect on albedo. Radiation absorption by individual leaves is influenced to a considerable degree by structural properties. The outer cuticle surface of the leaf epidermis reflects incident light specularly as a polarized component. The unpolarized, diffuse part of the reflection is influenced by leaf pigmentation and wavelength, which demonstrates that it emanates from the internal leaf structure. Also, leaf pubescence has little influence on energy input, but reflection of infrared (IR) radiation from the upper leaf surface is higher if the lower surface is hair-covered (Lösch 1994). Nutritional conditions also affect the reflective properties of vegetation and hence albedo (Lösch 1994), i.e. comparable forest formations can have different albedos caused by different nutritional conditions. If rain forest is succeeded by savannah albedo increases to 15 to 19 %, for pasture to as much as 18–21 %.

Destruction of tropical forests changes the earth's radiation balance by increasing CO_2 emissions, the resulting increase in the greenhouse effect thus counteracting the increase in albedo. The former effect is nearly three times greater than the latter: increased additional greenhouse effect will reduce the albedo effect by almost 40 %. Considering the case that an area of 1 million km^2 of Amazonian tropical rain forest were converted into pasture and savannah, the increase in albedo would reduce the average global radiation balance by 0.0172 Wm^{-2}. In contrast, the atmospheric CO_2 concentration would increase to such an extent that the radiation balance of the earth-atmosphere system would rise by 0.045 Wm^{-2}.

The increase in albedo means a reduction in solar radiation absorbed by the earth's surface and of the net energy input at the upper zone of the atmosphere.

Lower absorption of short-wave solar radiation at the earth's surface causes a decrease in radiative and latent heat flux as well as in the heat flux into the soil, thus leading to a cooling of the surface. If the net energy input of solar radiation into the earth-atmosphere system decreases, heating of the tropical troposphere will be lower. This will result in several feedback effects involving the water cycle and tropical atmospheric circulation according to model calculations, leading to a reduction of precipitation rates and an increase in near-surface temperatures, although less energy is available for heating the troposphere.

IMPACTS OF EXTENSIVE DEFORESTATION ON THE WATER CYCLE

Climate in the humid tropics is controlled by the trade winds among others. The trade wind systems of both hemispheres converge in the tropical convergence zone, and the air masses are forced to rise into higher strata. Numerous shower and thunderstorm clouds form. During condensation, heat is released and the upper troposphere warms up.

Since about 80% of the energy from solar radiation is used for condensation, small changes in the condensation rate can be of great significance for the tropical energy budget. Following destruction of forests, albedo increases, and less energy is thus available for the tropical regions. The atmosphere, with the exception of the lowest stratum of the troposphere, receives less heat for warming, precipitation decreases and the entire tropical circulation weakens. From adjacent areas, less air and, in turn, less water vapor moves into these regions, both of which are necessary for the maintenance of the upward air flow. This causes a further decrease in precipitation, and together with the atmospheric circulation the water cycle will weaken as well, affecting in turn the heat uptake at mid- and high-tropospheric levels.

Weakening of the entire water cycle and atmospheric circulation leads to further changes in the radiation budget, in addition to the reduction in precipitation over those areas where tropical forests have been destroyed. The earth's surface warms up, because less of the solar radiation is used for evaporation and a larger portion is therefore available for heating the earth's surface directly. Outgoing infrared radiation into space increases. Atmospheric water vapor content decreases, since less water is available for evaporation. Therefore, less radiation is absorbed by water vapor in the atmosphere, where the temperature increase is then smaller. Soils and vegetation dry out further, albedo increases and the solar radiation absorbed by the earth is reduced even more.

The vertical profile of temperature and humidity (dew point) after clear-felling shows the following changes: temperature of near-surface air layers increases by 2–3° C (on average) and soil temperature by 3–4° C subsequent to conversion to grassland. In the forest, the dew point is 25° C and in cleared areas 20° C. Over these areas, small clouds form at altitudes of approx. 1800–2500 m, while over forests, cloud formation begins as low as 800 m and extends up to over 4000 m.

Plant and soil properties affect the hydrological cycle. Low stomatal resistance indicates high evaporation rates. Grass (often C4 plants) has a higher stomatal resistance and thus lower evaporation rates than forest vegetation. Less of the incoming solar radiation is used for evaporation and the air temperature over pasture rises higher than in forest. Root depth affects the field capacity of the soil: low root depth (e.g. grass on pastures) results in a low penetration depth of water, while soil dries out faster and soil temperature rises. Field and infiltration capacity are interdependent: lower root depth and, therefore, lower field capacity causes decreased infiltration. During and after a rain shower, water penetrates to shallower depths and in lower amounts, runoff increases, less water is stored in the soil, which then dries out faster.

Disregarding general atmospheric circulation, the impression might be gained that climate in deforested tropical areas would have to become drier. However, regional climate, including the water cycle, is closely coupled with the global hydrological cycle and global atmospheric circulation, both controlling the site and movement of tropical low-pressure areas. Amounts of precipitation are expected to decrease only on average for all those areas where tropical forest has been cleared. This reduction amounts to at least 20%. Evaporation will be reduced by 30–40% and soil humidity by up to 60%. However, if the entire atmospheric circulation is shifted in certain areas it may become wetter.

A decrease in rainfall has a negative impact, especially on all succeeding ecosystems, because the dry period is extended. If monthly amounts of precipitation are below 50 mm for longer than three months, it will be impossible for a rain forest to grow on the cleared area. Consideration must be given to the fact that even under favorable conditions – i.e. a rather small cleared area – regeneration times of between 100 and 300 years are necessary before a new dynamic equilibrium is established. For extensively deforested areas, the time needed for regeneration – if it is allowed at all – is between 300 and 1000 years, depending on soil type and size of the area.

The remaining forest is threatened by a decrease in precipitation and the concomitant dessication of the vegetation. On the one hand, increased dryness further impairs the regeneration of the forest, and on the other dry forests are more easily set on fire. In the southern part of Amazonia amounts of precipitation are already barely sufficient to guarantee the existence of the tropical forests, which could be threatened if only small reductions in rainfall occurred. Using rainfall data, satellite imagery and field studies, it is estimated here that half of the closed forests of Brazilian Amazonia are dependent on deep root systems to maintain green canopies during the dry season. Evergreen forests in northeastern Pará state maintain evapotranspiration during five-month dry periods by absorbing water from the soil to depths of more than 8 m. In contrast, most pasture plants substantially reduce their leaf canopy in response to seasonal drought, thus reducing dry-season evapotranspiration and increasing potential sub-surface runoff relative to the forest they replace (Nepstad *et al.* 1994). There, too, pasture causes a decrease in precipitation with significant negative feedbacks for the remaining forest and for the regeneration of forest in this region.

In areas where agriculture is permanently established, not only the ecosystem is changed, because thousands of species are replaced by a few, but also nutrient cycling in the soil is accelerated, necessitating fertilization. Also, water and energy cycles are modified as a natural consequence of the altered plant cover. Loss of water by runoff into big rivers increases during the rainy season, especially during severe rains, and the availability of water for evaporation and transpiration is reduced.

Water and energy cycles are modified at the microclimate scale, relative humidity is reduced and temperature is higher. Variations in the amount of vapor condensing in the higher part of the troposphere (lack of evapotranspiration) may also influence climate on a global scale; during evapotranspiration, solar energy is transformed into latent heat. This heat is subsequently released in the atmosphere when water vapor condenses to form clouds. This energy plays a significant role in upper tropospheric circulation. On the other hand, part of this vapor is transferred to higher latitudes where, upon condensation, it releases energy. This is one of the important mechanisms for transferring energy from the equatorial to the polar regions. The extent to which the deforestation of the Amazon would influence general atmospheric circulation is not known. However, it is clear that some changes would be induced (Salati 1987).

The features of tropical deforestation believed to be of significance for the climate are:

- changes in local hydrology and the water balance, especially increases in runoff and decreases in evapotranspiration
- an increase in surface albedo (and possibly infrared emissivity)
- perturbation of turbulence characteristics in areas where tall and diverse forests stands are replaced by low crops or grazing land
- perturbation of the carbon cycle, probably adding to the already increasing level of atmospheric CO_2
- addition of particulates to the troposphere both directly from combustion and from increased wind-blown dust from exposed and drier soil surface (Henderson-Sellers 1987).

The effects of increased CO_2 concentrations as envisaged for North America, but also transferable to Amazonia, were summarized by Goldammer (1993) as follows. Under a 2 x CO_2 climatic regime, the frequency of drought events will increase dramatically (GISS model simulation for North America (Henderson-Sellers & Gornitz. 1984, Rind *et al.* 1990). The Palmer Drought Severity Index as well as the newly developed Supply-Demand-Drought-Index (SSDI = difference between precipitation and potential evapotranspiration) will increase significantly, both indices are an expression of the difference between atmospheric water supply and the humidity required. While at higher latitudes the relatively cool atmosphere can be more easily saturated with additional humidity and precipitation increases, the rather warm middle and low latitudes are characterized by increased evapotranspiration together with a deficit in soil humidity and increased vegetational water stress. In North America, extreme drought events today occur with a frequency of 5 % (relative to total time); by the year 2050, this figure will have increased to 50%. According to this model similar conditions are expected for the tropics.

This discussion clearly illustrates that the hydrological and energy cycles over Amazonia are fundamentally linked to the planetary-scale features of global circulation. The intense upward motion over Amazonia during the rainy months is energetically linked to the subtropical jetstream of the Northern Hemisphere and to compensating subsiding branches and circulation features over neighboring ocean areas of both hemispheres. All these features are in turn linked through teleconnection mechanisms to the circulation in more remote areas of the globe.

Observational studies and general circulation model (GCM) experiments show that this non-uniform distribution of time-averaged cloudiness and precipitation provides the largest asymmetric component for diabatic forcing for the time-averaged planetary-scale motions in the tropics (Wilson 1984). Tropical cloud systems have a direct impact on solar radiation, mainly by increasing the planetary albedo. High cirrus clouds of the type produced by the intense convective complexes of the tropics are extremely efficient regulators of heat loss to space.

Latent heating of the atmosphere is a major atmospheric energy source in the tropical regions with heavy rainfall. Seasonally-averaged fields of outgoing longwave radiation (OLR) show extensive minima (precipitation maxima) over South America/Central America and equatorial Africa, with a third and more extensive area of heavy convection indicated over the warm waters and adjacent land areas of the Asian-Australian monsoon region. These are the three major atmospheric heat sources in the tropics.

The impact of large tropical heat sources, such as the Amazon and Central Brazil during the summer, has also been shown to have a significant impact on the extratropical flow pattern. Other researchers have identified the propagation of long waves from the tropical heat source toward polar regions, and similar wavetrains are also expected to emanate from South America. More recently, it was shown that the upper level divergence over Central Brazil during the summer season and the associated south-eastward extension, known as the South Atlantic Convergence Zone (SACZ), act as a source region for some important teleconnection patterns in the Northern Hemisphere winter, such as the Eurasian pattern. Strong influences are also observed in the North Atlantic, affecting the eastern coast of the North American continent.

CLIMATE MODELS, MODEL PREDICTIONS

Like global climate predictions, assessment of the effects of Amazonian deforestation on climate, both regional and global, is reliant on climate models. As far back as its 1990 and 1992 reports, the Intergovernmental Panel on Climate Change (IPCC) stressed that the current generation of coarse-grid global coupled ocean-atmosphere general circulation models were capable of simulating large-scale features well, but were less successful at simulating regional climate (half-continent scale and smaller) (Houghton *et al.* 1990, 1992). In April 1994, the IPCC Working Group 1 Initiative on Evaluation of Regional Climate Simulations (Cubasch *et al.* 1994) reviewed the problems involved in simulating regional climate, and summarized work in progress regarding regional climate evaluations to identify gaps. The group noted that considerable resources will be devoted in future to the study of regional climate and climate impacts.

The interactions of the atmosphere, oceans, land masses and biosphere are central to the determination of climate and climate variability. To obtain a comprehensive understanding of the climate system, each of the components must be observed, understood, and modeled in the context of its impacts upon and extent of control by others. Tropical land and atmosphere form a highly interconnected system. The surface fluxes are coupled to the surface net radiation flux, the vegetation state, and the profiles of temperature and water, both below the surface and up through the atmospheric planetary boundary layer (PBL). These processes at the land-atmosphere interface are influenced in a fundamental way by topographic features and the highly heterogeneous character of the land surface layer. Consequently, the coupling of fluxes of heat and moisture across the interface may vary on spatial scales ranging from meters to thousands of kilometers. Observing and modeling these coupled land surface/atmosphere processes is crucial to an understanding and simulation of climate system interactions. A key challenge is to develop sub-components which will function efficiently over Amazonia in coupled models. The regional and possibly global effect on climate of continued deforestation of the Amazon Basin is a current question of major importance and scientific interest.

Models used to study the response to tropical deforestation contain both simple and relatively complex land-biosphere components. However, the coupling between the atmosphere and land-surface fluxes of heat and momentum is highly nonlinear, with many feedback mechanisms. One such mechanism is cloud feedback. Another important feedback mechanism in the context of tropical deforestation arises in connection with large-scale moisture convergence into the region. A reduction of evaporation may tend to reduce precipitation locally. However, if there is simultaneous surface warming, this causes a lowering of pressure in the low troposphere, which might drive larger moisture-laden airstreams into the basin and increase moisture convergence and precipitation, thus counteracting the initial decrease in precipitation. In the simulation experiments in general, although there was surface warming, large-scale moisture convergence in fact decreased for deforested scenarios. The reasons for this are complex and there is no simple interpretation. Large-scale moisture convergence might have decreased due to lower friction from the reduced surface roughness of grassland. Alternatively, there may be a complex interaction with convective precipitation. A reduction of convective precipitation induced locally by a decrease of evaporation may have the net effect of reducing the potency of the Amazon heat source through diminished release of condensational heat in deep cumulus clouds. A weakened heat source drives less moisture convergence from the Atlantic Ocean. Thus, there is clearly a need for further development of the parameterization schemes of the models as well as other aspects of the simulations in order to firmly establish a more likely scenario for the hydrologic effects of Amazon deforestation. These include multi-year runs to resolve natural variability in the model, more complete validation of the surface-biosphere parameterizations from field observations, better resolution of topography, and the incorporation of ocean-atmosphere coupling.

Accurate representation of the terrestrial biosphere in models of the earth system is an ongoing

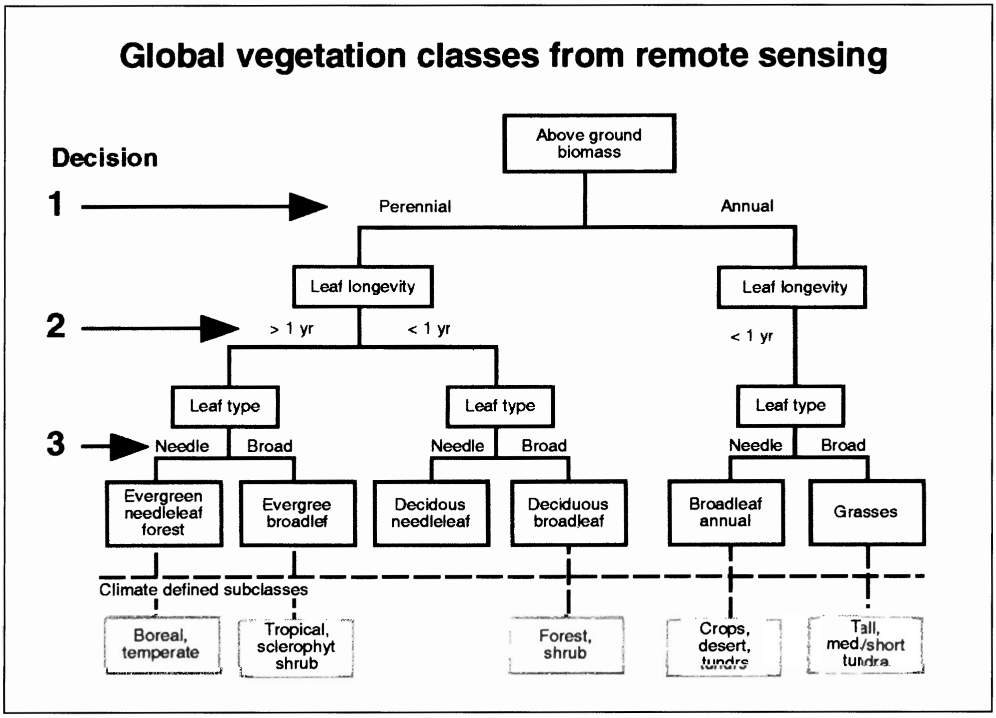

FIG. 12. A flowchart of the global vegetation classification logic. Each simple box identifies the variable being defined, and each decision point is illustrated. The final six classes of vegetation are shown in bold. Below the dotted line, potential-climate defined subclasses corresponding to more common classification schemes are suggested (Running *et al.* 1994).

challenge. The range of climates, geomorphic substrates, natural disturbances and human encroachments occurring globally has produced an incredible diversity of terrestrial vegetation. Only recently, global vegetation patterns have been computed from simulated climatology (Claussen & Esch 1994), and continental vegetation has been incorporated as a dynamic component into a global climate model (Henderson-Sellers *et al.* 1993, Walker 1994). In this study, the vegetation model is a static, equilibrium-response model. Claussen (1993, 1994), in his studies on the "Shift of Biome Patterns due to Simulated Climate Variability and Climate Change" and "On Coupling Global Biome Models With Climate Models" uses Prentice's *et al.* (1992) model of biomes, i.e., potential natural vegetation zones. This model is based on physiological considerations rather than on correlation between climate distribution and biomes as they exist today, with biomes emerging through the interaction of constituent plants. Hence, the biome model can be applied to the assessment of changes in natural vegetation patterns in response to climate changes. Although it does not simulate the transient dynamics of vegetation, it provides constraints within which plant community dynamics should operate. These new biome models are improving the global classification of vegetation, although they only produce maps of potential, not existing vegetation. Running *et al.* (1994) proposed a simple new classification logic , for global vegetation based on remote sensing, to be used in global biogeochemical models. The critical features of this classification are that it is based on simple, observable, unambiguous characteristics of vegetation structure that are important to ecosystem biogeochemistry and can be validated in the field; the structural characteristics can be determined by remote sensing, so that repeatable and efficient global re-classifications of existing vegetation will be possible (Fig. 12).

At present, model simulations for the Amazon rain forest are always based on "all-or-nothing" situations, i.e., forest or grassland/pasture. The consequences of intermediate changes (partial land conversion) have not yet been studied but, given the non-linearity of the coupled vegetation-climate system, may not be intermediate. Furthermore, global climate models do not resolve regional climates, although nested smaller-scale models can be used. These are active areas of research at present (Melillo *et al.* 1990).

DEFORESTATION AND REGIONAL CLIMATE CHANGE

In general, as described above, tropical deforestation directly affects the regional climate parameters of precipitation, surface temperature and cloud cover, due to modifications of the climate-relevant biosphere properties albedo, root depth and surface roughness.

Surface roughness reduction, after conversion of rain forest into pasture, causes a decrease in rainfall due to reduced convergence in low-pressure areas, and modifies the turbulent vertical exchange of air masses.

At the continental scale, several simulations have been carried out on the climatic impact of complete deforestation of the Amazon Basin (Lean & Warrilow 1989, 1990, Nobre *et al.* 1991) (Fig. 13). Reduced evapotranspiration and a general reduction in rainfall, although by variable amounts, was found in most simulations. The studies cited above show reductions of about 20% in rainfall in simulations where forest was replaced by grassland. Therefore surface water flow will also be affected. Lean & Warrilow (1989, 1990) showed that albedo and roughness changes contributed almost equally to the rainfall reduction. Nobre *et al.* (1991) suggest that the switch to a more seasonal rainfall regime, which they obtained, would prevent forest recovery.

According to the model operated by the British Meteorological Office, rainfall in the Amazon region will decline by about 20%, runoff by about 12%, evaporation by 30% and the ground level temperature will increase by about 2.5°C. The reduction in soil moisture will be very large, at almost 60%.

However, the results of mathematical models cannot be regarded as firm forecasts (Rowntree, oral testimony 1989). The assumption made in these models that, for instance, complete tropical deforestation takes place, does not reflect the actual situation at present. Nevertheless, these results can be seen as a clear indication that deforestation leads to a reduction in the volume of rainfall. It is difficult to prove this by means of direct observation. It will probably not be possible to prove that tropical deforestation results in reduced rainfall until it is already too late to act (Fearnside, oral testimony 1989).

The rainfall rates of all models drop by approximately the same amount, namely by around 3 to 4%, in conjunction with a 1% increase in albedo. All these models are based on an assumed increase in albedo. The model operated by the British Meteorological

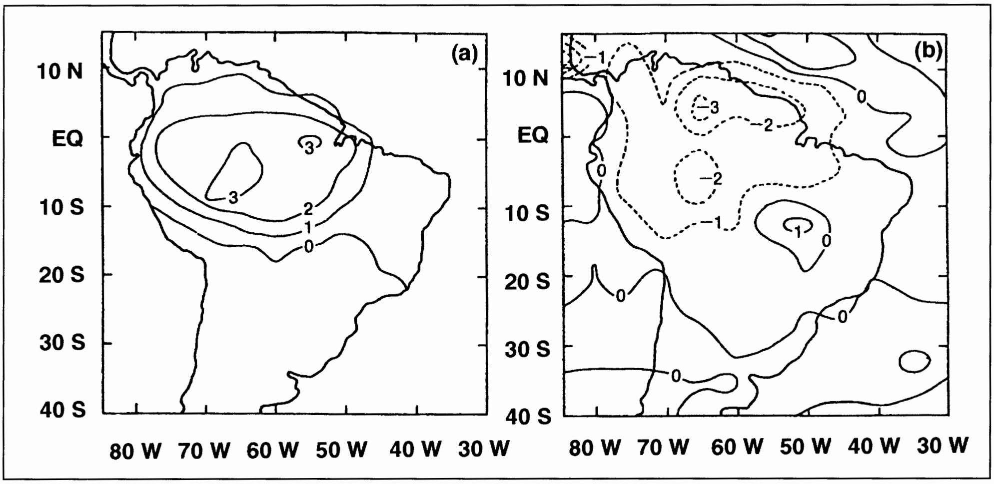

FIG. 13. Changes in annual means due to deforestation of northern South America (Nobre *et al.* 1991): (a) Surface temperature (contours every 1° C), (b) Precipitation (contours every 1 mm day^{-1}, negative contours are dashed).

Office also takes account of changes in cloud cover. Cloud cover in the southern Amazon Basin decreased by 2% and in the northern Amazon Basin by about 3%.

Forecasts for changes to the volume of rainfall vary widely according to region and time. The model of the British Meteorological Office assumes a maximum increase in albedo of 7% in the regions of today's tropical forests. Between September and November, rainfall is heaviest in the northern Amazon Basin, while between December and February it is heaviest in the southern Amazon Basin. According to this model, the volume of rainfall decreases in most regions as a result of tropical deforestation. However, in certain regions rainfall increases. This stems from the change in general atmospheric circulation caused by deforestation, which leads to a subsequent change in the pattern of rain-bearing tropical air flows. The reduction in rainfall is greatest during the season that is drier in any case, thus prolonging its duration.

As a result of deforestation, the average ambient ground-level temperature increases by 2 to 3° C, and soil temperature by 3 to 4°C (see Fig. 13; Nobre 1989). In addition, the daily fluctuation in air temperature becomes more pronounced. This is a result of the Anglo-Brazilian Amazonian Climate Observational Study (ABRACOS), which demonstrated local-scale, mesoscale, and large scale climatic impacts of Amazonian deforestation (Gash *et al.* 1996). In addition, in areas of substantial deforestation, higher sensible heat fluxes from the cleared forest produce deeper convective boundary layers, with differences in cloud cover (Gash & Nobre 1997). Changes in local temperature, indicating difference between radiation and energy balance are the easiest to verify. These temperature increases have been confirmed by observation. Temperature increases close to ground-level, because reduced evaporation leads to reduced rainfall, allowing a larger amount of solar radiation to be converted to heat at the earth's surface.

EDGE EFFECTS

Edge effects between burned-down and pristine forest areas mostly have an impact on microclimate. Increased solar radiation and air flow along the edges causes vegetation to dry out more and faster than under normal conditions, and to become more easily inflammable. A thinning-out of the canopy cover by selective felling has the same effect. Along the edges changes in radiation provide conditions for a different ground vegetation, i.e., the growth of a dense scrub vegetation usually absent in primary forest. Due to these changes, ground fires can intrude into the forest more easily during a new forest fire. If humidity decreases and incoming solar radiation increases, the species composition at the forest edge will change. The role of pioneer species adapted to these climatic conditions, as a sort of umbrella necessary for the regeneration of primary forest, is of greater significance for cleared areas than for edges.

Species composition of the vegetation in those edge areas not affected by forest fires will be influenced by changed radiation conditions. The decrease in humidity causes stomatal resistance to increase, thus reducing gas exchange (Dickinson & Henderson-Sellers 1988). The plants are under constant water stress – living between dying of thirst and starvation – except for the times during and directly after a rain shower. Due to reduced gas exchange, and rates of photosynthesis decrease the plants are in a steady state, thus surviving these changes in environmental conditions, or are being replaced by better-adapted pioneer species. Therefore, the primary forest with its long-lived species fixing CO_2 in the long term will lose precisely these edge areas even after a forest fire. The size of these edge areas has not yet been studied, but it is expected to show considerable variability depending on the different forest types in Amazonia, the size of the deforested area, the distribution of precipitation during the year, the type of soil, etc. The significance of these edges must not be underestimated, because the forest is being continually fragmented by felling activities (e.g., in Rondonia), and the remaining "stripes" of forest are usually only a few kilometers wide.

The initial effects of forest fires, for example, are seen in an increased tree turnover time. Turnover, measured in terms of tree mortality and recruitment, has increased since the 1950s, with an apparent pantropical acceleration since 1980. These trends in forest dynamics may have profound effects on biological diversity (Phillips & Gentry 1994).

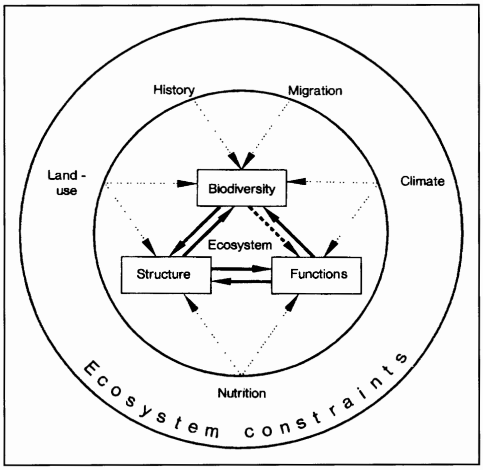

FIG. 14. Schematic representation of the interactions of biodiversity, community structure, and function within an ecosystem, and effects of external ecosystem constraints on these parameters. Solid arrows indicate strong feedbacks, while the dashed arrow between biodiversity and function indicates a weak feedback. Dotted arrows indicate major effects by external factors on ecosystem internal processes (Schulze & Mooney 1994).

BIODIVERSITY AND THE STABILITY OF ECOSYSTEMS

Documentation of the role of biodiversity for ecosystem function will be critical for developing an integrated understanding of the role of natural ecosystems in the global climate system and for preparing to deal with any climate change that does occur (Pitelka 1994).

The relationship between biodiversity and ecosystem structure as well as function is presented in Fig. 14, together with the impacts of climate, nutrient supply, land use and colonization. It is unclear exactly how and why a change in biodiversity might alter the functioning of ecosystems, i.e., the transfer of carbon, water, and nutrients, and the maintenance of ecosystem stability. However, the stability of ecosystems is directly dependent on the natural biodiversity of the biotope, because the species influence each other, the vegetational cover, the soil, and vice versa. Ulrich (1987) developed a theory according to which ecosystem stability depends on the element cycle, i.e., soil chemistry controls the type of vegetation and hence biodiversity. Susceptibility to diseas as well as natural catastrophe (fire, storms, etc.) is much greater in biotopes with low biodiversity. The significance of parasites in natural plant communities is essentially unknown. They can be important at two levels: as a selection factor at the genetic level, and for fragmentation of populations, i.e., they limit the density of single-species populations (Burdon 1994).

It is difficult to determine the minimum level of biodiversity necessary to stabilize an ecosystem because different biotopes are highly variable. A general conclusion is that it is not normally possible to quantify the number of species comprising a functional ecosystem. However, if the species complement is not at the maximum which the ecosystem can include, then the dynamic richness is reduced. This reduced richness then lays the ecosystem open to invasion and disruption (Woodward 1994).

There are various theories regarding the limiting factors for the natural colonization of new ecosystems such as the degraded areas of Amazonia. Nutrient supply and light are highly significant. At first the availability of nutrients in the soil is the limiting factor for new colonization. During succession the amount of light becomes a selection factor, while later on light and nitrogen become limiting, with each species having its own optimum combination of these factors.

Global anthropogenic changes in climate, rates of habitat disturbance, nutrient loading rates and other environmental constraints will have a major impact on the successional dynamics and maintenance of biodiversity. The greatest effect may be to increase vastly the importance of colonization limitation. Species adapted to the unique combinations of environmental constraints that result from global change may not exist or may be so distant from a site that they are unable to colonize it. One of the major effects of global change on natural ecosystems is that successions will occur without the benefit of a pool of species adapted to the current climatic conditions. This limitation on colonization will exceed any that has occurred during recent glacial recessions. The resulting communities would be species-poor and highly susceptible to dramatic invasion by plants, herbivores, pathogens, and predators (Tilman 1994).

In summary, global change will certainly cause species diversity to decline in the short term. If the rate of global change eventually declines to the point that equilibrium models (e.g. Tilman 1994) become useful, we predict that species diversity in the long term could (1) decrease in temperate mesic and tropical regions due to atmospheric nitrogen deposition, (2) increase in dry mid-continental, temperate and tropical regions due to increased drought and overgrazing, and (3) decrease in Arctic regions in response to increased nutrient availability (Hobbie *et al.* 1994).

GLOBAL EFFECTS OF AMAZONIAN DEFORESTATION

Global impacts of deforestation have not yet been resolved for Amazonia, but are always presented for all tropical rain forest regions worldwide.

TABLE 1. Global annual emissions of greenhouse gases (modified from Houghton 1990).

	Annual emissions	Percentage of total emissions	Radiative forcing relative to CO_2 molecule	Contribution to greenhouse effect in 1980s (%) Total deforestation
CO_2			1	50
Industrial	5.6 Pg C			
Biotic	2.0–2.8 Pg C			
Tropical deforestation	2.0–2.8 Pg C	26–33		13–16
CH_4			25	20
Industrial	50–100 Tg C			
Biotic	320–785 Tg C			
Tropical deforestation	155–340 Tg C	38–42		8
N_2O			250	5
Industrial	1 ? Tg N			
Biotic	3–9 ? Tg N			
Tropical deforestation	1–3 ? Tg N	25–30		1–2
CFCs			1000s	20
Industrial	700 Gg			
Biotic	0 Gg	0		0

Assessment of the contribution of tropical deforestation to global climate change shows that trace gas emissions presently account for only about one fifth of the atmospheric CO_2 increase. The resulting additional greenhouse warming is about one tenth, because CO_2 contributes only half of this effect. Major sources are still the burning of fossil fuels (coal, oil and gas; Grassl 1990). However, while global climate changes in the context of an additional greenhouse effect are caused primarily by the industrial nations, the impacts can feed back on the tropical forests in a way that cannot yet be foreseen. Thus, the net effect of tropical deforestation is to warm climate: the removal of all tropical forests could warm the climate by about 0.3° C.

Current global emissions of greenhouse gases have been compiled by Houghton (1990) (Tab.1). The net flux of carbon from tropical deforestation has not been measured directly, because deforestation is too widely dispersed in time and space. Instead, carbon flux is calculated from two kinds of data: estimated rates of deforestation and reforestation, and changes in the amount of carbon held in vegetation and soils per unit area of land. Houghton calculated that the release of C to the atmosphere from land-use change in 1990 was 1.1–3.6 Gt C, while the IPCC (1990) estimate of net average annual emissions for the decade 1980–1989 was 1.6 Gt C, which is consistent with Houghton's figure within the limits of uncertainty.

CONCLUSION

The Amazon Basin contains almost one-half of the world's undisturbed tropical evergreen forest as well as large areas of tropical savannah (Melillo *et al.* 1996). Large areas have been deforested, with the forest being replace by pasture that is mainly used for cattle ranching. Therefore an appropriate balance is required between the need to maintain human economic development and our need to understand the mechanisms involved in global processes, since these are prone to change as a consequence of the scale and effectiveness of our interventions in natural systems (Shuttleworth 1994). This paper presents and reviews the past progress and previews future large-scale observational and modeling studies which address the need for better understanding rain forest degradation and its relation to development.

With regard to future research, emphasis should be on the following:

Accurate measurements of climate-relevant factors in primary forest, secondary forest and agricultural areas. This includes the following parameters: albedo, precipitation, temperature, wind velocities, relative humidity, and evaporation rate.

Investigation of edge effects. At present, little is known about the impact of deforestation on forest areas bordering open fields. It is essential to measure these edge effects in order to reach conclusions on the stability of the remaining forest. The research literature contains no data on, for example, changes in humidity at different levels in the forest and the impact of these changes on vegetation. For extensively deforested areas, the effects of changes in microclimatic factors (wind, precipitation, albedo) at higher levels in the adjacent forest can be traced for several kilometers. These kinds of investigations have to be carried out in the various forest formations of Amazonia with great accuracy, because changes in humidity and radiation conditions (not only quantitative, but also the modified spectral composition of light) affect photosynthesis and biomass production as well as, eventually, species composition. Subsequently, output and fixation of climate-relevant gases (CO_2, CH_4, NxO, etc.) will also change.

The carbon dioxide fertilizer effect, i.e., balancing the carbon budget under future CO_2 conditions (2 x CO_2). To assess the potential carbon-fixation capacity of the vegetation in different forest formations of Amazonia, investigations have to be carried out at various sites employing semi-enclosed systems with 2 x CO_2 supply over at least several months using representative species from each individual forest type. Since there are hardly any analyses available on the carbon budget of individual species at different age stages over at least one vegetational period (ca. 1 year), the respective objects of investigation also have to be measured under present CO_2 conditions for standardization purposes. So far, investigations of Amazonian rain forest vegetation involving the exposure of plants or their organs to increased carbon-dioxide concentrations for only a few hours or days were unable to produce any conclusions about the behavior of plants under continuous stress.

The extent, richness and diversity of Amazonia, and its importance and relevance to the future of our planet, mean that it is the ideal open-air laboratory for interdisciplanary global research. Therefore the Large-scale Atmospheric Micrometeorological and Biospheric Amazonian Data Acquisition (LAMBADA) project was established. The study will address the need for better understanding of hydrometeorological interactions with emphasis on the measurement and modeling of regional and mesoscale advection processes (Shuttleworth 1994).

Emissions of greenhouse gases (NxO, NH_4, O_3) by Amazonia's flora and fauna. This objective, requiring a grid of monitoring sites, relates to the key question of the LBA project, namely "What are the fluxes of trace gases into the atmosphere over Amazonia, what are the controls on these fluxes, and how will these be altered with changing land use?" LBA is designed to create the new knowledge needed to understand the climatological, ecological, biogeochemical, and hydrological functioning of Amazonia, the impact of land use change on these functions, and the interactions between Amazonia and the earth system (Nobre *et al.* 1996).

Changes in hydrological cycles due to climate change, clear-felling, and subsequent land use. As far as the significance of the Amazonian forest cover for global climate is concerned, what is required are further studies on rainfall distribution, as well as modified cloud cover and cloud formation, due to deforestation. Investigations to date have lacked the area coverage of the sampling grid. Therefore basing an assessment on the extrapolation of those results to the entire Amazonian basin cannot yield sound answers.

The improvement of understanding of how processes in the soil, plants and atmosphere change as forest is replaced by pasture means that we can have more confidence in the interpretation of model predictions. However, it is still absolutely imperative that more knowledge is gained on Amazonia's tropical rain forest in various fields of basic research. In the words of Wendell Berry "We cannot know what we do as long as we do not know what Nature would do if we did nothing."

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