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Understanding biosphere—precipitation relationships : Theory, model simulations and logical inferences

Y. C. SUD, K. M. LAU, G. K. WALKER* and J. H. KIM**

Laboratory for Atmospheres, NASA/Goddard Space Flight Center, Greenbelt, MD-20771

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सार — भूमि-वायुमंडल के मध्य पारस्परिक क्रिया को नियंत्रित करने वाले वनस्पति के चार मुख्य जीव-भौतिकीय नियंत्रक वनस्पति के निम्न गुणों से उत्पन्न होते हैं : (क) वाष्पोत्सर्जन, (ख) पत्तों में ही सौर विकिरण को समाहित करना, (ग) रंध्री (स्टोमी) नियंत्रण द्वारा वाष्पोत्सर्जन को व्यवस्थित करना, और (घ) प्रक्षुब्ध आवर्तों के अनुसार सतह की कठोरता में संशोधन (सामान्यत: वृद्धि) करने की क्षमता। कुछ प्रेक्षणात्मक विश्लेषणों व सामान्य परिसंचरण मॉडलों के साथ किए गए प्रतिरूपण अध्ययनों से वनस्पति एवं वर्षण के सुस्पष्ट पारस्परिक प्रभाव की जानकारी मिली। वनस्पति संबंधी प्रक्रिया को कृत्रिम ढंग से उभार कर किए गए अध्ययनों से वर्षा की वनस्पति पर सुस्पष्ट निर्भरता का पता चला। सुहेल और अन्य उष्णकटिबंधीय मरुस्थल के सीमावर्ती क्षेत्रों में, जहां वाष्पोत्सर्जन कम होता है, सतह-ऐल्बिडो (मरुस्थल बनने की प्रक्रिया) के बढ़ने से वर्षा कम होती जाती है। वनस्पति के संभावित प्रभाव से कुछ चुने हुए क्षेत्रों में वाष्पोत्सर्जन अथवा भूमि-सतह की कठोरता बढ़ने पर स्थानीय वर्षा भी बढ़ जाती है। उपर्युक्त प्रभावों के निजी अथवा संयुक्त मेल से भारतीय उप-महाद्वीप में अधिकाधिक मानसूनी वर्षा का प्रतिरूपण हुआ।

उष्णकटिबंधीय वनों और वर्षा के बीच संबंधों की जानकारी प्राप्त करने के लिए जीवमंडल के यथार्थपरक मॉडलों के साथ किए गए प्रतिरूपण अध्ययनों से पता चला कि अमेज़ान के वनों के नाश से जलवायु गरम और शुष्क हो गई है। चूंकि वन अधिक सौर ऊर्जा को समाहित करने व अधिक वाष्पोत्सर्जन उत्पन्न करने के साथ-साथ सतह-कठोरता के प्रभाव के कारण आर्द्रता को एकस्य करते हैं, अत: स्वाभाविक तौर पर यह माना जा सकता है कि वर्षण के लिए वन अनुकूल स्थिति उत्पन्न करते हैं। हमारे नए प्रतिरूपण प्रयोगों से न केवल उपर्युक्त निष्कर्षों की पुन: पुष्टि हुई है, बल्कि अब होने वाली वनों की कटाई के कारण विश्व पर पड़ने वाले दुष्प्रभावों के परिणामों का भी पता चला है। पिछले दशक के प्रतिरूपण परिणामों की सूक्ष्म व्याख्या से यह अनुमान लगाया जाता है कि जीव-वायुमंडल की पारस्परिक क्रियाओं में होने वाले परिवर्तन विभिन्न जीव-पारिस्थितिकियों, जैसे, वन, चारागाह, कृषि-भूमि और मरुस्थलों का अस्तित्व बनाए रखने तथा उनकी वृद्धि के लिए अनुकूल आवश्यकताओं की पूर्ति में महत्वपूर्ण भूमिका निभाते हैं।

ABSTRACT. The four major biophysical controls of vegetation, which govern land-atmosphere interactions, emanate from the ability of vegetation to: (a) evapotranspire, (b) trap solar radiation within leaf organizations, (c) regulate evapotranspiration by stomatal control, and (d) modify (generally increase) the surface roughness on the scale of turbulent eddies. Simulation studies with General Circulation Models together with a few observational analyses have provided a rational understanding of vegetation-precipitation interaction. In studies with artificially enhanced vegetation-related processes, a strong dependence of rainfall on vegetation has been inferred. For Sahelian and other tropical desert-border regions, where evapotranspiration is small, increasing the surface-albedo (desertification) decreases rainfall. When evapotranspiration and/or land-surface roughness are increased in some selected regions — a potential effect of vegetation, an increase in local rainfall is produced. The above effects, both individually and jointly, have simulated increased monsoon rainfall over the Indian subcontinent.

Modelling studies directed at understanding the relationship between tropical forests and rainfall with realistic models of the biosphere have simulated a warmer and drier climate in response to Amazonian deforestation. Since forests absorb more solar energy and produce much larger evapotranspiration, as well as moisture convergence through the surface-roughness effect, positive feedback effect of forests on precipitation can be expected naturally. Our new simulation experiments not only reaffirmed the above results but also suggested potential global consequences due to the ongoing deforestation. From a synthesis of modelling results of the last decade, it is further inferred that variations in the biosphere-atmosphere interactions play an important role in redistributing continental precipitation to fulfill the survival and growth requirements of different biomes: forests, pasture, agricultural lands, and deserts.

Key words — Biosphere, Vegetation, Deforestation, Simulation, Flux, Convection, Evapotranspiration, Rainfall, Boundary layer, Roughness.

1. Introduction

The ongoing tropical deforestation in Amazonia and elsewhere has led to an urgency about the need for understanding the influence of the biosphere on the climate and environment. Since the deforestation is primarily anthropogenic, it can certainly be curtailed and/or monitored to circumvent the

* General Science Corporation, 6100 Chevy Chase Drive, Laurel, MD 20707—2959.

** Applied Research Corporation, 8701 Corporate Drive, Suite 1120, Landover, MD 20785.

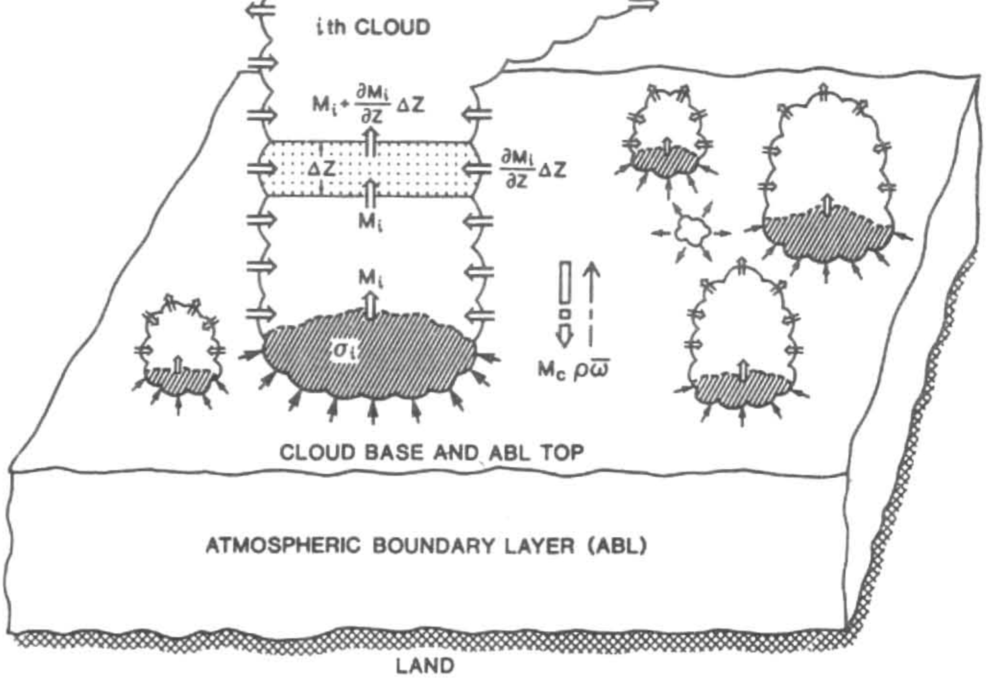


Fig. 1. Schematic representation of cumulus ensembles of different cloud sizes (types) (*Adapted* from Arakawa and Schubert 1974)

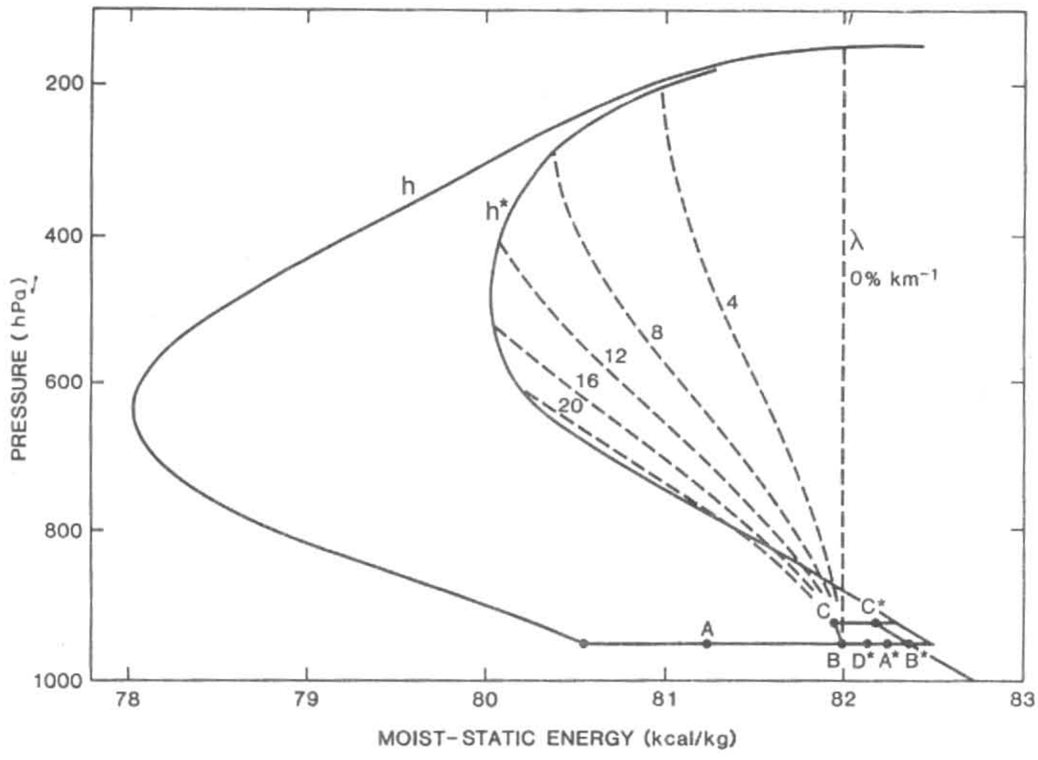


Fig. 2. Profiles of h and h^*. for a typical tropical atmosphere with cloud paths for different λ's

potential catastrophe that may occur in the global climate and/or the local environment. Toward this goal, a number of observational and modelling studies have been performed throughout the world. Our motivation here is to synthesize the results of several earlier and currently ongoing studies so as to better understand the response of the climate system and to provide valuable guidance for the future of global environment.

In his review of the influence of vegetation on rainfall from a historical, as well as pragmatic perspective, Anthes (1984) noted that people in different parts of the world (mostly tropical) generally inferred from observations that forests help to increase the rainfall. Nevertheless, such an inference could be easily flawed if it stems from the existence of forests in the raining areas. A scientific understanding of the influence of land-surface fluxes on atmospheric circulation has emerged only recently and a consensus has begun to develop as a consequence of a number of simulation studies with biophysical models of land/vegetation-atmosphere interactions within the state-of-the-art atmospheric General Circulation Models (GCMs). To grasp the essence of several pioneering discussions and works on the subject, one needs to refer to comprehensive papers by Dickinson (1980 and 1983), Hare (1983), Mintz (1984), Garrett (1993) and a recent report by Dirmeyer and Shukla (1994). On the other hand, the evidences from the analyses of observational data are scanty, often based on indirect inferences from correlations. However, Otterman *et al.* (1990) have shown an observational evidence of a remarkable correspondence between vegetation increase accomplished by reducing land use (protected grazing) and rainfall for southern Israel.

Let us first conceptually analyse the basic influence of the biosphere on land-surface fluxes. We know that vegetation (*also biosphere*) absorbs the photosynthetically active radiation (PAR) more efficiently than the bare soil. In fact, several studies have established that an increase in leaf-area-index (LAI) increases PAR absorption in the biosphere. Such relationships enable one to express land-surface albedo as a function of LAI and other geophysical parameters (Dorman and Sellers 1989). Observations show that a typical moist (dry-arid) soil albedo of 15% (45%) easily reduces to only about 8-10% for lush green vegetation cover. It is also common knowledge that open country (with copious vegetation cover) is significantly cooler than cities. This happens primarily because biosphere converts a large part of solar flux into latent energy flux (LE). Cooler land would naturally produce less outgoing longwave radiation at the surface (OLRS), less surface sensible heat energy flux (HE), as well as less ground heat flux (G). Evidently, the biosphere increases LE while reducing all other land-surface fluxes.

Charney (1975) enunciated and subsequently Charney *et al.* (1977) demonstrated that the surface-albedo increase is a major contributor to the positive feedback effect of a desert on itself: the "so called" biogeophysical component of desertification. Several subsequent simulation studies, including a few by the author and his colleagues at the Goddard Laboratory for Atmospheres (GLA), have shown an equally important influence of evapotranspiration and land-surface roughness, both of which have been shown to increase the rainfall [*e.g.*, Sud and Smith 1985, Sud and Molod 1988 and Sud *et al.*1988]. However, we must recognize that a majority of the earlier simulation studies were conducted for tropical environment and were often limited to summer conditions. Consequently, we must try to understand how summer rainfall, which is largely (80% or more) convective between 30°S & 30°N, responds to vegetation. We will first start with theoretical considerations (section 2), then refer to short forecasts invoking 5-day rainfall anomaly simulations for deforested Amazonia (section 3), and finally discuss the use of several fully interactive biosphere-atmosphere simulations with GCMs (section 4). A logical thesis stemming from these studies will be discussed in the conclusions (section 5). We will also show how the biosphere can affect, as well as adapt to, the rainfall climatology of the earth.

2. Theoretical considerations

Despite absorbing more solar radiation and subsequently releasing it in the form of surface fluxes, vegetated regions are generally cooler than the bare land because a significantly larger fraction of the absorbed net radiation is transformed into evapotranspiration. In this way vegetation increases the moisture content, as well as the moist static energy, of the Atmospheric Boundary Layer (ABL). Heat and moisture fluxes can easily produce rising motion, moist-convection, and rainfall. Nevertheless, it is common observation that an isolated grove of trees in an otherwise bare-land environment does not produce a discernible effect on the local rainfall. Indeed, if heat and moisture fluxes on such small scales could produce noticeable increase in moist-convection, one would find a high incidence of moist convection and rain in the vicinity of cooling towers (used by electric power plants) which spew out a large amount of heat and moisture into the atmosphere. On such small space scales, the eddy-mixing time-scale is very short (typically a few minutes); and this mixing inundates any convective activity, because the time required for the life cycle of moist-convection is much larger (typically 30 minutes). Consequently, whenever convective clouds tend to form over the cooling towers (or other small regions), the dry ambient air mixes with the clouds and dissipates them. Nevertheless, if the space scales of anomalies in surface-fluxes (and therefore vegetation) were of the order of 10-100 km, the outcome could be quite different; on this scale, the eddy-mixing time becomes much larger and that, in turn, allows the cumulus convection to mature. The cloud-scale turbulent plumes that emerge out of the ABL will have sufficient time to become convective towers. We must remember, however, that these space scales are not so large as to significantly affect the synoptic-scale systems which control the global weather and which are typically 100-10,000 times larger than the individual convective clouds. Thus, convective cells that are large enough to produce rain but are too small to produce any significant effect on the planetary and synoptic scales allow one to infer biosphere-atmosphere interaction from simple theoretical considerations without having to resolve the complexities and non-linearities of the interaction between surface fluxes, moist convection, and atmospheric dynamics.

The above analogue can be used to understand the mechanisms of biosphere-ABL-moist convection interactions referring to the basic principles invoked in the Arakawa and Schubert (1974) cumulus parameterization. According to Arakawa and Schubert, cumulus clouds of different horizontal scales (sizes) emerge from the ABL whenever the environment is moist convectively unstable. Clouds entrain ambient air during ascent, and ultimately detrain at the level of neutral buoyancy which is depicted in Fig. 1 schematically for an assertment of cloud sizes (types). The moist convection occurs when the moist static energy at the ABL top, h_{ABL} ($= c_pT + gz + \lambda q$, which is roughly equal to the moist-static energy, h_{cloud}, at the cloud base), exceeds the saturation moist static energy h^* ($= c_pT + gz + \lambda q^*$) of the ambient air. Here h (h^*) have three terms : enthalpy, potential energy, and latent enegy of water vapour. The variables represent specific heat of air, c_p; temperature, T; acceleration due to gravity, g; height above sea level, z; latent heat of evaporation, λ; and the ambient (saturation) specific humidity, q (q^*). The cloudy air parcels popping out of the convectively unstable ABL grow into convective cloud masses of different shapes and sizes. These clouds entrain ambient air during ascent and eventually detrain where h_{cloud} path curve (drawn dotted because of an irreversible process) again intersects the h^* curve. Fig. 2 shows that clouds emerge with $h_{cloud} = h_B$ and then rise to the detrainment level where $h_{cloud} = h^*$. Different λ's represent different fractions of ambient air mass entrainment. Larger λ's cause the cloud-path to turn leftward due to mixing of ambiocnt air with smaller h into the rising cloud with larger h. Due to vegetation processes, the moist static and saturation moist static energies that are at B & B* might get readjusted to : (i) A & A* for increased surface albedo scenario; (ii) C & C* for increased roughness-length scenario; and (iii) B (representing no change) & D* for increased evapotranspiration scenario. The difference between B* and h^* at the same height is explained by the temperature jump across the boundary layer inversion. Clouds that entrain more (less) are small (large) and consequently detrain at a lower (higher) level. However, the bases of all convective clouds must appear above the ABL top.

In this example, the ambient h is roughly 81.5 k cal/kg; it must increase to a value of about 83.0 k cal/kg, which is close enough to h^* value for the emerging ABL parcel to become buoyant by condensation heating. This represents a rise in temperature of roughly 6°C. It contains both sensible and latent heat, and it amounts to a thermal energy of about 1500 k cal m^{-2} for a 1 km deep boundary layer that can be acquired in 3-4h via solar heating during a typical summer day. Evidently, a substantial part of increase in h_{ABL} is due to land-surface fluxes particularly in summer. Thus, if land-surface albedo were to increase, more solar energy would be reflected and, therefore, less would be available for latent and sensible heat fluxes; consequently, the contribution of the land-surface fluxes to h_{ABL} (as well as the convective cloud base) would shift h_B leftward to h_A (Fig. 2). If the h-z paths for different cloud types are reworked from the new starting point, A, the resulting convective clouds either (a) may not emerge because of the higher relative humidity requirements for the lifting condensation to occur, or (b) may emerge and be relatively shallower and produce less rain. Hence, we infer that surface-albedo increase is bound to reduce moist convection.

An increase in evapotranspiration may not substantially alter the magnitude of total surface fluxes : net OLRS, sensible heat, and evapotranspiration. However, by making the ABL cooler, the evapotranspiration vitally alters the saturation

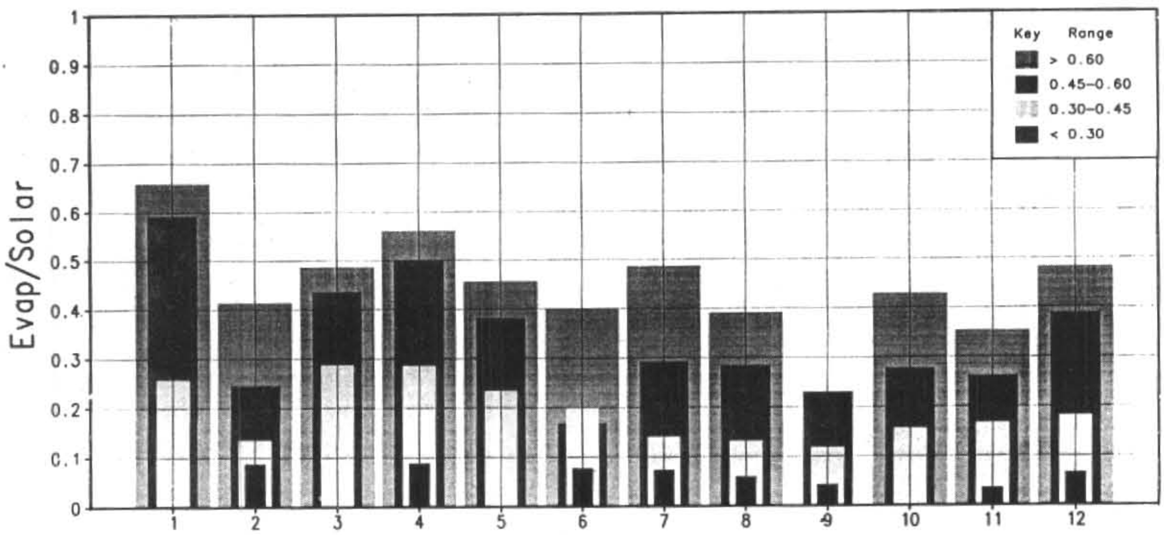


Fig. 3. Evapotranspiration binned by soil moisture as a fraction of solar income for different SiB-biomes: (1) Tropical rainforest; (2) Broadleaf deciduous forests; (3) Mixed broadleaf deciduous an needle leaf evergreen forests: (4) Needle leaf evergreen forests: (5) Needle leaf deciduous forests: (6) Savannah (grass. deciduous trees and shrubs): (7) Grassland/pasture: (8) Broadleaf deciduous shrubs singly or in patches: (9) Broad-leaf deciduous shrubs single or bare-land: (10) Tundra. (11) Desert or Barren land and (12) Cropland typically wheat

moist-static energy at the cloud base. *i.e.*, it moves h^*_B to h^*_D, which is much closer to h_B (Fig. 2). Thus evapotranspiration enables the cloudy parcels of ABL origin to reach the Convective Condensation Level (CCL) and thereby promote the onset of cumulus convection. Conversely, reduced evapotranspiration increases the sensible heat flux which raises the ABL temperature while reducing the specific as well as the relative humidity. If the warm and dry ABL-air is unable to meet the CCL criterion. only dry convection can ensue which is relatively shallow and further dries out the ABL by mixing it with the warmer and drier air aloft. Sud and Molod [1988 (a)] isolated the moisture removal process by dry convection as a key mechanism that keeps the Sahara desert dry during the summer season. *Indeed, the stomatal control mechanism for evapotranspiration which works through the foliage leaves is yet another remarkable feature of evapotranspiration. In the rainy (wet) periods, the increased stomatal opening enables aggressive growth, producing increased evapotranspiration; while, during the dry spells, the stomates close to conserve water and the evapotranspiration reduces significantly.* In this way, biosphere helps to promote moist convection that may occur under humid conditions by injecting small but critically valuable amount of water vapour into the ABL.

An increase in surface roughness, such as produced by uneven forest canopies, helps to increase the kinetic energy of the typical turbulent eddies (~50 m size) and deepens the ABL. Physically, it represents the influence of mechanically-generated turbulence on the ABL growth. This growth would reduce h, because generally the humid ABL air mixes with the drier air from aloft. For a stable lapse rate, the mixing usually decreases the specific humidity and increases the temperature within the ABL. However, a drier ABL promotes increased evapotranspiration to partly mitigate the drying, while the deeper ABL-top is naturally cooler and its saturation specific humidity is lesser; consequently, both h_C and h^*_C, reduce at the cloud base (Fig. 2). Evidently, the two competing mechanisms have the following effects: one tends to reduce h_C; while the other undoubtedly reduces h^*_C; therefore, it is not imperative that the boundary layer growth must always enhance moist-convection. However, by testing this effect on an ensemble of more than 100,000 atmospheric profiles, Sud *et al.* (1993) have demonstrated that ABL growth increased the convective rain in roughly 90% of the cases. Thus, we have identified three modes of modification of surface fluxes by the biosphere, all of which affect h and h^* so as to promote moist-convection. We argue that if changes in the biosphere are on a 10-100 km scale, which can only affect the local ABL but without significantly altering the global-scale circulation, the foregoing inferences provide a useful answer to the fundamental question of the response of rainfall to the presence of biosphere.

With regard to the actual magnitude of modification of surface fluxes by the biosphere, there is no simple answer. The answer depends on a

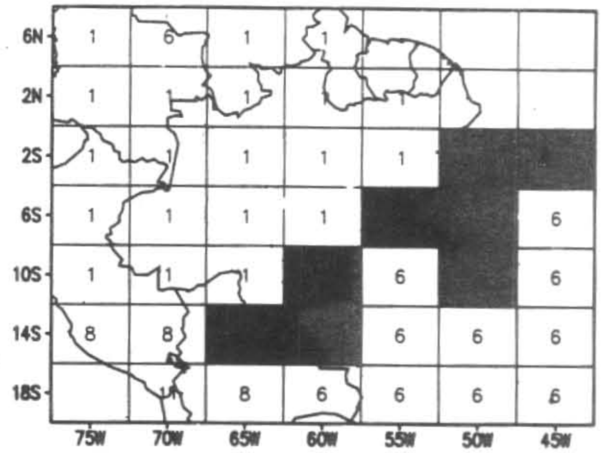


Fig. 4. Discretized deforested regions of Amazon basin from *in situ* data analysis of Skole and Tucker (1993) represented on a 4 × 5 resolution for the GLA GCM

number of vegetation and land-hydrology parameters: soil moisture, soil type, ambient conditions, solar income, and, in fact, a host of factors and biophysical constants that are used in the mathematical model of a biosphere. A broad estimate of the magnitude of changes in surface fluxes was obtained by binning the fraction of absorbed solar energy as a function of biome type which shows the joint effects of vegetation and soil (Fig. 3). The graph is based on a 3-year (1978-1981) integration with the GLA GCM and, therefore, includes the response of the biosphere to the annual cycle. According to this figure, there is a large variation in the evapotranspiration yields (as a fraction of solar income) among different vegetation types. Consequently, one contends that the variation in surface fluxes caused by the biosphere can be quite large. Indeed, this also demonstrates that the biosphere plays a key role in maintaining as well as modulating the rainfall climatology of the earth.

3. Short forecasts

Until recently, the potential climatic influence of ongoing deforestation of tropical rainforests in Amazonia, equatorial Africa, and other tropical regions has been an unanswered question. Simulation studies with GCMs often invoking extreme scenarios, in which the entire tropical forest regions/basins were deforested and replaced by pasture, have suggested a local reduction in rainfall with general warming and drying of the region. Nevertheless, a scientifically based understanding of the influence of biosphere remains elusive, because of the varying degree of complexity of GCMs as well as significant variations in the design of simulation experiments (*see* section 4). As a logical advance from the theoretical inferences of the previous section, let us attempt to delineate the influence of Amazon deforestation from short forecasts.

By making several short forecasts and using the ensemble averages to infer the influence of biosphere on local rainfall, Walker *et al.* (1994), in essence, disallow adjustment of the large-scale circulation to the vegetation anomaly, but, on the positive side they circumvented the situation where systematic simulation errors can accumulate to create spurious feedbacks. For specifying the locations of deforested areas, the study used Skole and Tucker's (1993) *in situ* observations of deforestation. Since deforestation is generally partial allowing for roads and alleys and the planned anthropogenic activity, the deforested regions were more appropriately replaced by savanna in this investigation. Fig. 4 shows the shaded deforested areas that were converted into savanna (Biome type 6 in SiB nomenclature). The five 5-day vegetation anomaly simulations were drawn from a 3-year-long integration with the GLA GCM which was assumed to represent the "nature" run. The rainy period of Amazonia: January through March, was chosen for this experiment. This period has virtually no soil-moisture stress and, hence, the biosphere evapotranspires at the near-potential rate. Consequently, the vegetation differences are the key determinants of the outcome. These differences are primarily related to the biophysical differences between the two biomes: forests and savanna. The five forecasts were repeated with savanna conditions, and the ensemble averages of the forested and savanna simulations were compared. The comparisons show that the evapotranspiration anomaly is primarily limited to the deforested regions [Fig. 5(a)]: it shows a mean reduction of 0.79 mm/day, which is 18.75% of the evapotranspiration in the control. The rainfall reduction is also largely limited to the deforested regions, but it begins to spread with time by advection and the dynamical consequences of changes in the regional diabatic heating [Fig. 5(b)]. Such rainfall reductions cause local moisture divergence and accompanying moisture convergences in the neighbouring regions where the simulated rainfall anomalies are positive. The near-surface wind adjustment to the lowered surface roughness in deforestation occurs in a short time (within about a day), while the rainfall anomalies persist even in the 2 and 5-day averages. The overall rainfall reduction is roughly 1.2 mm/day in deforested regions, which is roughly 8% of the total rainfall in the control. The study shows quite vividly that the current level of Amazonian deforestation simulates reduced local

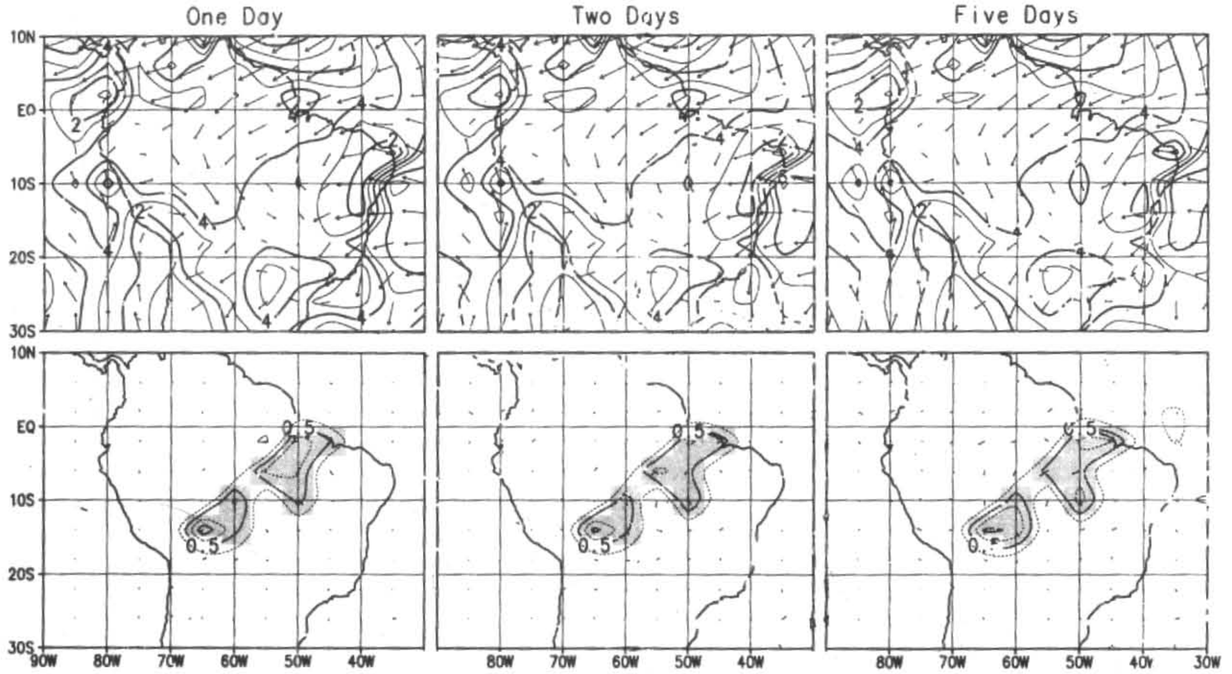


Fig. 5 (a). Ensemble-mean evapotranspiration fields (in mm/day) of all the five cases from left to right for 1, 2 and 5-day average fields for the control case (top panels) and deforested *minus* control case (bottom panels). Wind vectors are for the control case

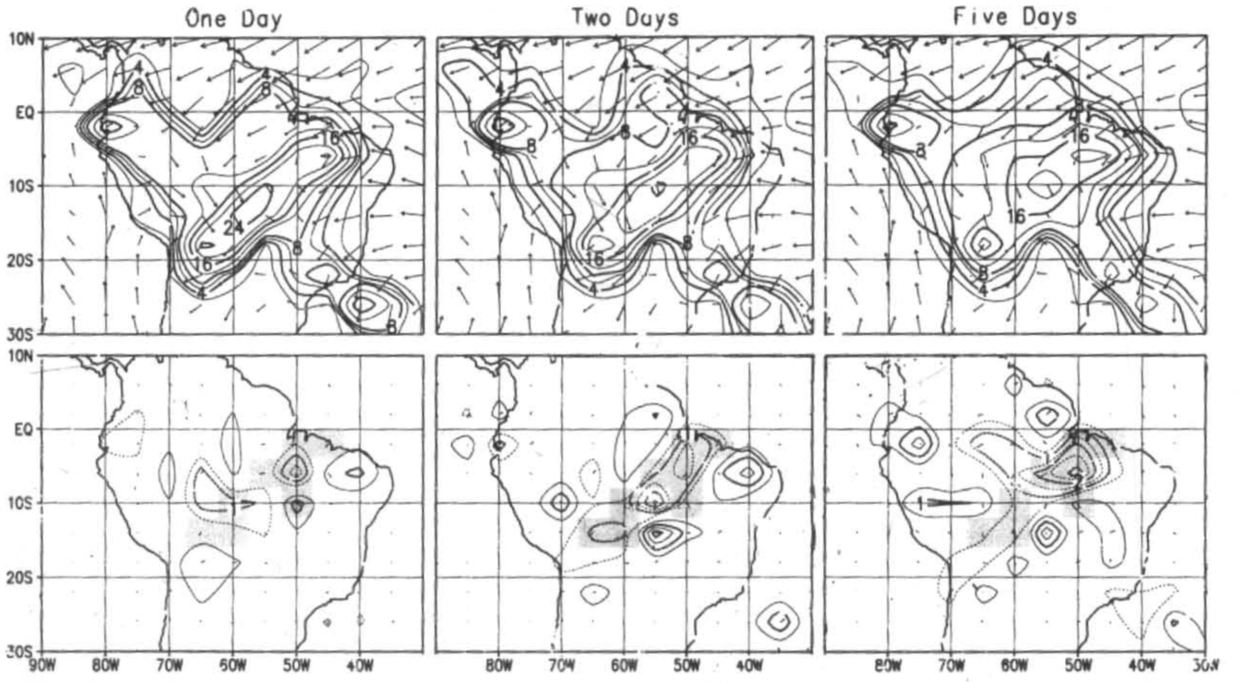


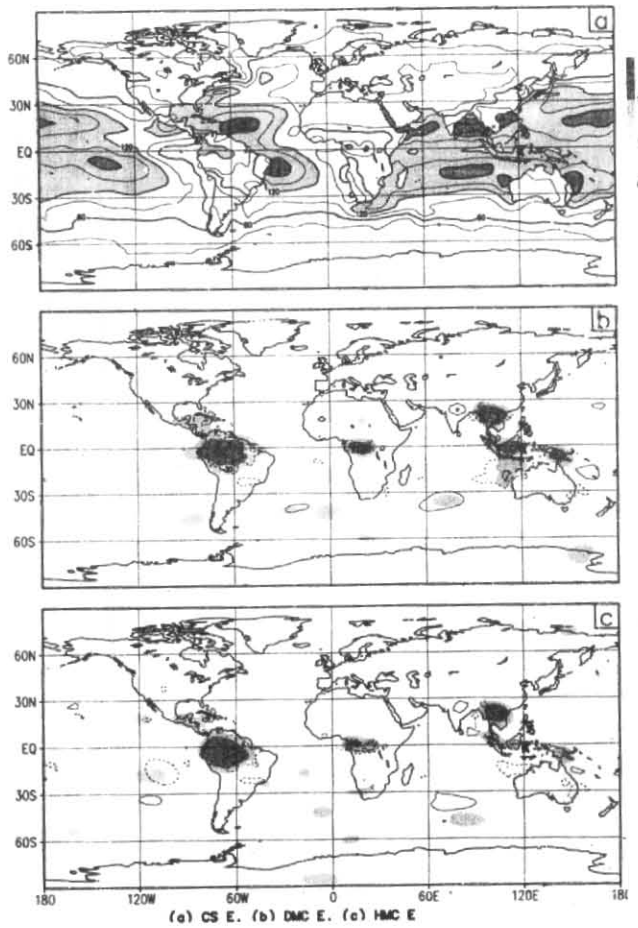
Fig. 5 (b). Ensemble-mean precipitation fields (in mm/day) of all the five cases from left to right for 1, 2 and 5-day average fields for the control case (top panels) and deforested *minus* control case (bottom panels) rainfall, but this is much less than that simulated in full deforestation scenarios (section 4).

4. Deforestation simulations

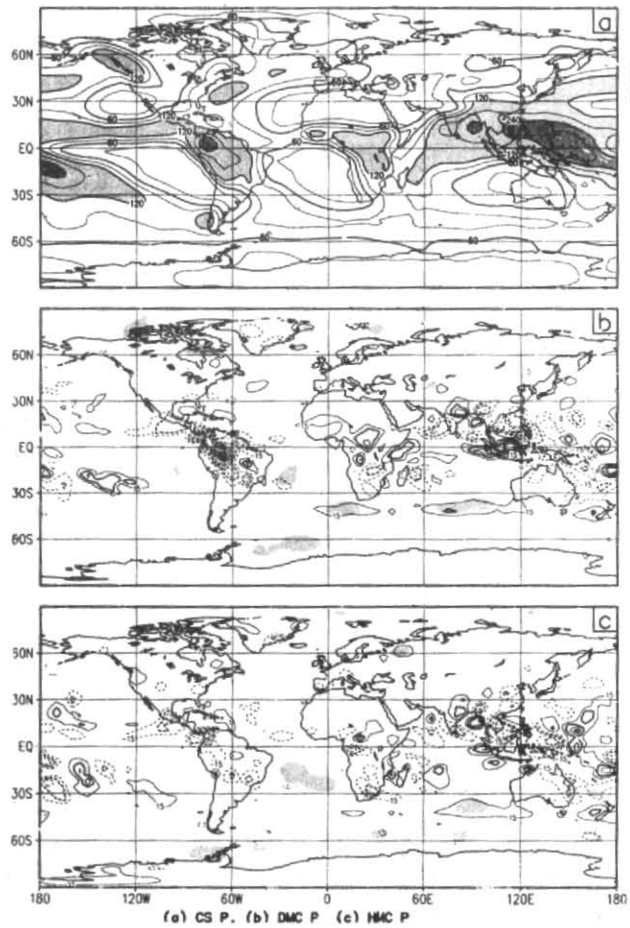
Although some simulation studies were conducted to determine the influence of tropical deforestation on rainfall prior to the use of biospheric models within GCMs (*e.g., see*, Henderson-Sellers and Gornitz 1984), the concerted effort to understand the relationship followed the advent of biospheric models such as SiB and BATS. In these studies, the tropical deforestation was extended to either all tropical or merely Amazonian rainforests. The deforested regions were generally replaced by pasture even though the biophysical properties of the two biomes do vary somewhat among these simulations. Dickinson and Henderson-Sellers's (1988) results showed a significant warming but with virtually no change in rainfall while Nobre *et al.* (1991), Handerson-Sellers *et al.* (1993), and Lean and Warrilow (1989) and Lean and Rowntree (1993) showed a significant reduction in rainfall, alongwith some warming of the region. The Lean and Rowntree (1993) study identifies net radiation, rainfall interception, and runoff parameterizations to be the key determinants of the overall outcome. Henderson-Sellers *et al.* (1993) have comprehensively summarized various simulation studies and pointed out the common and differential features among them. Eltahir and Bras (1993) have successfully explained the differences between Nobre *et al.* (1991) and Dickinson and Henderson-Sellers (1988)'s results by considering the combined influence of boundary-layer heating and upper-level cooling. The potential confusion in the earlier simulation results led the author and his colleagues at GLA to perform two new studies with the GLA GCM [Sud *et al.* 1994 (a & b)]. Presumably, these deforestation-scenario studies can enable us to develop a better understanding of the dependence of rainfall on vegetation for the tropical environment. The ultimate goal is to understand and abate the undesirable effects of tropical deforestation on the climate system. However, the influence of tropical deforestation on global climate must be examined *vis-a-vis* other mechanisms of global climate change so that the simulation results are not confused with other climate change phenomena. Often, the interannual climate variability is much larger than that of the slowly emerging effects of the ongoing tropical deforestation. The primary cause for the climate variability in the tropics has been determined to be the Sea Surface Temperature (SST) anomalies, particularly the *El Nino* and *La Nina* episodes. Nevertheless, the interactive land-hydrology, and hence its accurate parameterization, is essential for the maintenance of the climate variability as demonstrated by Koster and Suarez (1994). Even for Sahelian droughts, the role of surface-albedo increase due to desertification may ultimately turn out to be secondary to the role evolving global-scale SST anomalies (Folland *et al.*, 1986).

Recently, Sud *et al.* [1994 (a)] made two 3-year (1979-1982) integrations using the GLA GCM with the Simple Biosphere (SiB) model for land-atmosphere interactions. The control case used the usual SiB vegetation cover, while the deforestation case imposed a scenario in which all tropical rainforests were entirely replaced by pasture. This study duplicated a scenario simulated by several earlier studies described before. The results show that tropical deforestation leads to the following: (a) It decreases evapotranspiration by a small but statistically significant fraction as compared to the control [Figs. 6 (a & b)] in all the deforested regions. This is accompanied by increased land-surface outgoing longwave radiation, and sensible heat flux, thereby warming and drying the planetary boundary layer [*see* Table 1 (a) for area-averages]; (b) It produces significant and robust local, as well as global, climate changes. As compared to the control case, the local effect includes large changes (mostly reductions) in rainfall [Figs. 7 (a & b)] and diabatic heating [Table 1 (a)], while the large-scale effect is to weaken the Hadley and Ferrel cells. The latter causes an imbalance in the condensation heating and radiative cooling, thereby causing a much larger airmass to be drawn from the indirect polar cells [Figs. 8 (a) & 9 (a)]; (c) It decreases the surface stress (drag force) over the deforested land, which, in turn, intensifies winds in the planetary boundary layer thereby affecting the dynamical structure of moisture convergence. The simulated surface winds are about 70% stronger and are accompanied by significant changes in the power spectrum of the annual cycle of surface & PBL winds and rainfall [Table 1 (a)].

Deforestation scenario simulations also produce a large decrease in rainfall in the northeastern coastal region of Amazonia where the northeast trades invade the continent and experience an increase in the surface drag force caused by the roughness of the forests. Deforestation reduces this roughness and its dependent surface drag and moisture convergence, thereby making a large impact on the local structure and magnitudes of rainfall anomaly.



Figs. 6 (a-c). Simulated evaporation and its anomalies (in mm/month) for the 1095-day (3-calendar-year) integration period : (a) Control (CS): (b) DS *minus* CS (DMC): (c) HS *minus* CS (HMC). Light (dark) shading in the difference fields in the middle and bottom panels (b and c) represent statistical significance at 90% (95%) levels. A contour interval of 15 Wm^{-2} is used with alternate thick and thin lines



Figs. 7 (a-c). Same as Fig. 6 but for simulated rainfall and its anomalies

Indeed, the change in surface roughness-length (z_0) also introduces a large complexity into the land-PBL-atmospheric interactions, which could potentially produce increased local drying and reduced moist convection. The influence of z_0 reduction in deforestation is assessed by comparing the above two integrations with a third, in which deforested pasture was assumed to have a z_0 of 265 cm (same as rainforests) instead of 7 cm (typical value for pasture). The differential analyses of this simulation as compared to the control and earlier deforestation simulations shows that the z_0 change marginally affects the apportionment of available net surface-radiation energy into latent and sensible fluxes: consequently, the average increase in near-surface temperature and decrease in evapotranspiration are not too different between the two cases [Figs. 6 (b & c) and Table 1(b)]. Indeed, the corresponding rain fall reduction over Amazonia are much smaller as compared to the deforestation case with surface roughness change [Figs. 7(b & c)].

The simulated area averaged rainfall (evapotranspiration) differences are −0.73 (−0.96) mm/day for DMC (deforestation *minus* control) and −0.39 (0.83) mm/day for HMC (hypothetical *minus* control). It represents a small fraction of the corresponding annual mean values : 5.99 (3.84) mm/day obtained in the control case. Nevertheless, the corresponding anomalies for Amazonia are −1.48 (−1.22) mm/day for DMC and −0.09 (−0.88) mm/day for HMC simulations respectively [Tables 1 (a & b)]. However, the differences in surface winds, boundary-layer moisture convergences, as well as the rainfall and diabatic heating fields, particularly in DMC are quite large and statistically significant. This happens because z_0 change directly affects the surface stress and its dependent momentum flux.

The most intriguing results of the HMC *vis-a-vis* DMC comparisons are in the Mean Meridional Circulation (MMC) change. The weaker Hadley

TABLE 1 (a)

Rainforests and grassland simulations with and without z_0 change

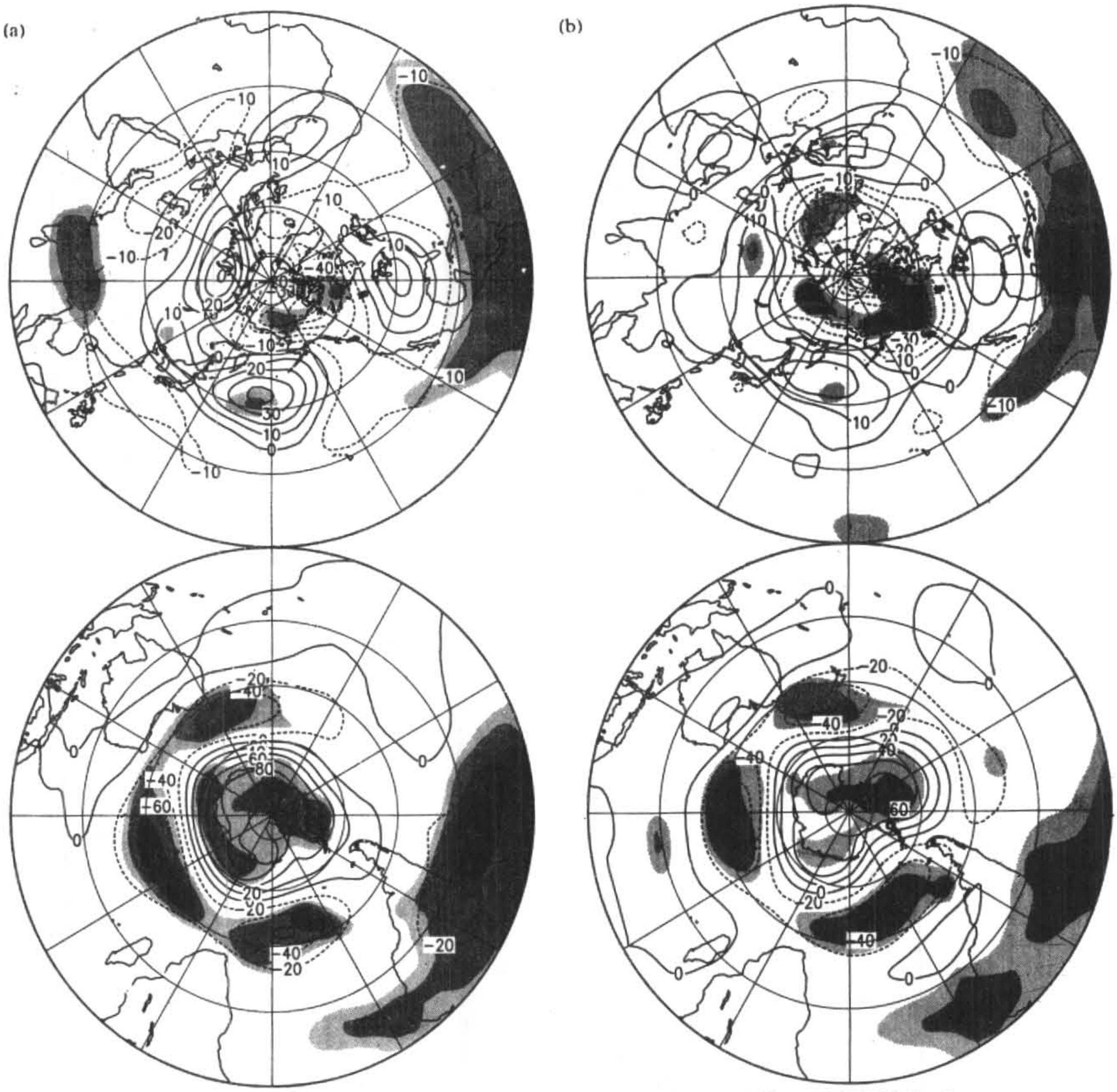
Parameter/or field	Control tropical forests		Deforested Changed z_0		Deforested Unchanged z_0	
	All	Amaz.	All	Amaz.	All	Amaz.
Surface albedo (%)	9.7	9.2	14.6	14.2	14.6	14.2
Shortwave-surface (Wm^{-2})	204.3	209.3	196.8	205.3	194.5	197.7
Longwave-surface (Wm^{-2})	50.3	49.1	65.1	69.9	55.8	54.9
Net rad.-surface (Wm^{-2})	154.0	160.2	131.7	135.4	138.7	142.8
OLR (Wm^{-2})	232.8	239.1	237.2	247.8	233.4	236.5
Latent flux (Wm^{-2})	111.1	110.6	83.3	75.2	87.0	85.1
Sensible flux (Wm^{-2})	42.8	49.8	48.4	60.2	51.7	57.8
Temp.-surface (K)	297.1	299.2	298.4	301.2	297.9	300.2
Temp.-ground (K)	296.2	298.5	299.4	302.4	297.7	300.0
Temp.-vegetation (K)	296.7	298.9	298.9	301.9	297.6	299.9
Wind-magnitude (ms^{-1})	0.98	1.93	2.02	4.54	1.05	2.28
RH-surface (%)	73.7	71.4	64.6	58.7	68.7	65.8
Total ppt. (mm/day)	5.99	5.40	5.26	3.92	5.60	5.31
Conv. ppt. (mm/day)	5.18	4.82	4.59	3.56	4.87	4.74
Evapotranspiration (mm/day)	3.84	3.82	2.88	2.60	3.01	2.94
Moist. diverg. (PBL) (mm/day)	−2.05	−2.75	−1.82	−2.01	−2.10	−2.07
Moist. diverg. (column) (mm/day)	−2.15	−1.58	−2.38	−1.32	−2.59	−2.37
Water input to soil (P-E) (mm/day)	2.15	1.58	2.38	1.32	2.59	2.37
Runoff amount (mm/day)	2.25	1.60	2.39	1.34	2.61	2.37
Change in moisture (mm/day) Storage in soil-layers	−0.10	−0.02	−0.01	−0.02	−0.02	0.00

cell, in association with increased surface pressure (not shown) and 200 hPa height anomalies [Figs. 8 (a & b)] over Antarctic and correspondingly negative surface pressure and 200 hPa height anomalies at 40-50°S. appears in both simulations. The MMC anomalies are also remarkably similar in both simulations and yield statistically significant anomalies virtually at the same places [Figs. 9 (a & b)]. Sud *et al.* [1994(b)] analysed the local energy budget and found that the weaker Hadley cells for deforested scenarios upset the delicate energy balance between radiative cooling and condensation heating of the Ferrel cell, which is thereby forced to draw air mass from the indirect polar cell to maintain the condensation heating necessary to mitigate the local radiative cooling. On the other hand, large differences in rainfall over the deforested Amazonia and the neighbouring oceans are largely caused by the differences in local moisture convergence and divergence distributions brought about by the change in land-surface roughness: these local differences are virtually inconsequential for the MMC which responds to zonal mean diabatic-heating anomalies: consequently, the surface-albedo change, which was identical in both deforestation scenarios, seems to dominate the effect on MMC.

5. Summary and conclusions

Following the Arakawa and Schubert (1974) physically based model of cumulus parameterization in which cumulus cells originate in and emerge out of the atmospheric boundary layer, it is easy to infer how the convective rainfall yield must depend upon the water vapour content and moist static energy of the boundary layer. In summer, the moist static energy at the ABL top crucially relies on the incremental gains from land-surface fluxes. Vegetation helps to increase both the water vapour content and the moist static energy of the ABL and thereby promotes moist convection.

The ensemble mean of five 5-day forecasts also shows that Amazonian deforestation-produced changes can reduce the evaporation by about 0.8 mm/day and the resultant rainfall yield by about 1.2 mm/day. Even though the evapotranspiration anomalies are limited to the deforested areas, the rainfall anomalies emerge all around the deforested



Figs. 8 (a & b). Annual mean 200 hPa geopotential height anomalies (m) for : (a) DMC, and (b) HMC. Shading represents statistically significant differences. The light, medium, and dark shades contain regions having 90%, 95% and 99% statistical significance respectively

areas and slowly evolve to engulf the nearby regions into which the winds advect the land-surface effects. The overall outcome of the 5-day averages appears to be affected by local, as well as continental, scale changes in diabatic heating and circulation.

An intercomparison of the three 3-year simulations: one control, two deforestation scenarios (one with and one without surface roughness-length change) reveals that removal of tropical rainforests changes the temperature, relative humidity, winds, and rainfall. The results show that even the large differences in rainfall (evapotranspiration) are only a small fraction of the corresponding annual mean values; the anomalies for Amazonia are −1.48 (−1.22) mm/day for deforestation *minus* control (DMC) and −0.09 (-0.88) mm/day for hypothetical *minus* control (HMC) simulations. Amazonia being the single largest deforested region, its response can be viewed as a continental scale tropical response.

A comparison of the local rainfall anomalies produced in the two deforestation scenarios shows that rainfall reductions in Amazonia are much larger (smaller) for the case with (without) the surface roughness-length reduction. On the other hand, the MMC leading to a surface high over the

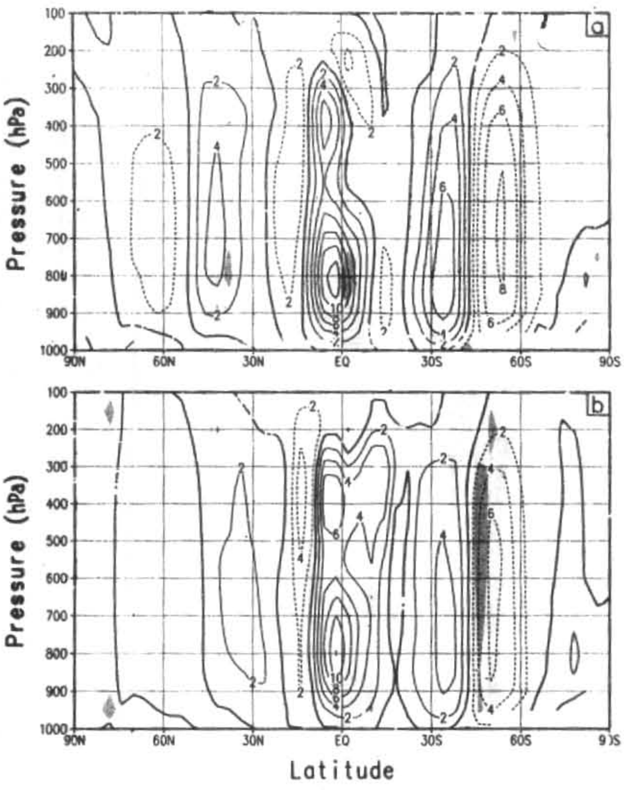
TABLE 1 (b)

Rainforests and grassland simulations with and without z_0 change

Parameter/or field	Anomalies caused by deforestation			
	With z_0 change		Without z_0 change	
	All	Amaz.	All	Amaz.
Surface albedo (%)	4.9	5.0	4.9	5.0
Shortwave-surface (Wm^{-2})	−7.5	−4.0	−9.8	−11.6
Longwave-surface (Wm^{-2})	14.8	20.8	5.5	5.8
Net rad.-surface (Wm^{-2})	−22.3	−24.8	−15.3	−17.4
OLR (Wm^{-2})	4.4	8.7	0.6	−2.6
Latent flux (Wm^{-2})	−27.8	−35.4	−24.1	−25.5
Sensible flux (Wm^{-2})	5.6	10.4	8.9	8.0
Temp.-surface (K)	1.3	2.0	0.80	1.0
Temp.-ground (K)	3.2	3.9	1.5	1.5
Temp.-vegetation (K)	2.2	3.0	0.9	1.0
Wind-magnitude (ms^{-2})	1.04	2.61	0.06	0.35
RH-surface (%)	−9.1	−12.70	−5.0	−5.6
Total ppt. (mm/day)	−0.73	−1.48	−0.39	−0.09
Conv. ppt. (mm/day)	−0.59	−1.26	−0.33	−0.06
Evapotranspiration (mm/day)	−0.96	−1.22	−0.83	−0.88
Moist. div. (PBL) (mm/day)	−0.70	−0.19	−0.05	−0.68
Moist. div. (column) (mm/day)	−0.23	0.26	−0.44	−0.79
Water input to soil (P-E) (mm/day)	0.23	−0.26	0.44	0.79
Runoff amount (mm/day)	0.14	−0.26	0.36	0.77
Change in moisture storage in soil-layers (mm/day)	0.09	0.00	0.08	0.02

South Pole and surface low in the region of 40-50°S are quite similar in both simulations. It is argued that the MMC response to tropical diabatic heating which drives the Hadley cell, is primarily affected by surface albedo anomaly; alternatively, the large regional differences between the two deforestation scenarios largely reveal the influence of z_0 change on the near-surface cross-isobaric moisture convergence or divergence. On the basis of these results, the surface-roughness reduction in deforestation was demonstrated to be an important determinant of the local biogeophysical consequences of tropical deforestation.

A synthesis of the above results leads to a plausible hypothesis of the influence of a rainforest on the local rainfall. A healthy forest canopy with copious foliage is an efficient trap for the solar radiation; a large part of this energy is used to evapotranspire and thereby provide incremental moisture and moist-static energy to the boundary layer, which is essential for moist convection. In a prolonged drought, however, the foliage wilts, as well as dies back, thereby decreasing evapotranspiration, as well as intercepting less solar radiation. This enables more solar radiation to reach the dark tree barks and the forest floor, both of which absorb and disburse the energy as sensible heat and OLRS fluxes. These fluxes warm the boundary layer, which consequently induces a thermally-forced local moisture convergence (Eltahir and Bras 1993). Such a thermal convergence attempts to mitigate the effect of reduced evapotranspiration in a drought. Nevertheless, in a deforestation that produces denuded land, both the surface roughness and evapotranspiration would reduce while the surface albedo increases; this would produce a much larger reduction in rainfall. Indeed, reduced rainfall over denuded land is better suited for *(i)* maintaining the ground water table, and *(ii)* mitigating undesirable consequences of land erosion and flooding. For agricultural land or pasture, the influence of the biosphere is roughly mid-way between forests and denuded deserts as indicated by the evapotranspiration analysis differences; consequently, they exert only a mid-way influence on the local rainfall. These considerations enable us to infer that forests, grasslands, and



Figs. 9 (a & b). Annual mean meridional circulation anomalies : (a) DMC, and (b) HMC in 10^9 kg/sec with shading for regions showing statistically significant differences. The regions having 90%, 95% and 99% statistical significance are shaded light, medium and dark respectively

deserts have the remarkable ability to share the available continental rainfall to better fulfill the needs of the local biota and land surface hydrology without any external help.

Several simulation results have demonstrated that the biosphere has the inherent ability to increase convective rainfall. Since such inferences are difficult to verify experimentally, one must rely on the simulation models to understand the biosphere-atmosphere interactions. The primary qualification for the reliability of predictions of global-scale dynamical consequences of deforestation is the ability of the model to respond to observed surface-flux anomalies of similar magnitudes elsewhere in the tropics, *e.g.*, the response to *El Nino* and *La Nina* SST episodes. In a 10-year (1979-1988) integration with the GLA GCM for Atmospheric Model Intercomparison Project organized by Gates (1992), Kim and Sud (1993) show only partial success in simulating the observed circulation and rainfall anomalies with the GLA GCM even though the model did fairly well as compared to other GCMs. This fact does warrant some caution on the value of prediction of global consequences of tropical deforestation.

Finally, we argue that even if the ongoing simulation studies with General Circulation Models cause some confusion about the influence of tropical deforestation on future rainfall, particularly because of the sheer multiplicity of models and compexities of the physical parameterizations in them, the positive feedback effect of the biosphere on the rainfall, that is inferred from fundamental considerations and logical deductions, is likely to remain valid. We sincerely hope this result is useful for ecosystem managers.

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