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Role of vegetation in sustaining large-scale atmospheric circulations in the tropics

Elfatih A. B. Eltahir

Parsons Laboratory, Massachusetts Institute of Technology, Cambridge

Abstract. The focus of this paper is the role of rain forests in large-scale atmospheric circulations. The significance of this role is investigated by studying the response of the tropical atmosphere to a perturbation in the state of vegetation (deforestation) over three regions: the Amazon, Congo, and Indonesia. A theory is developed to relate tropical deforestation and the resulting changes in the large-scale atmospheric circulation. Field observations and numerical simulations support the argument that tropical deforestation reduces the total net surface radiation, including terrestrial and solar forms. However, the energy balance at the land-atmosphere boundary dictates that for equilibrium conditions, any reduction in net surface radiation has to be balanced by a similar reduction in the total flux of heat, including sensible and latent forms. Since these fluxes supply heat as well as entropy from the forest into the atmospheric boundary layer, a reduction in the total flux of heat reduces the boundary layer entropy. In a moist atmosphere, that satisfies a quasi equilibrium between moist convection and radiative forcing, the equilibrium temperature profile is uniquely related to the boundary layer entropy. Under such conditions, large-scale deforestation reduces boundary layer entropy relative to the surroundings, cools the upper troposphere, and results in subsidence, divergent flow in the boundary layer, and weakening of the large-scale circulation. These changes are simulated using a simple linear model of atmospheric flow. The comparison of the model predictions with observations of atmospheric circulations over the Amazon, Congo, and Indonesia suggests a significant role for vegetation in maintaining large-scale atmospheric circulations in the tropics.

1. Introduction

Rain forests cover large regions in the tropics, including the Amazon basin, the Congo basin, and Indonesia. The same three regions are the locations of three large-scale atmospheric circulations with ascending branches over the forests. These circulations bring moisture from the oceans into the continental regions and generate rainfall over large areas. Vegetation and large-scale atmospheric circulations constitute a biogeophysical system in equilibrium. On the one hand, the climate conditions created by these circulations represent ideal environments for rain forests. On the other hand, latent and sensible heat fluxes from the forest into the atmosphere provide important boundary conditions for the dynamics of these circulations.

Recently, the intensity of human activity in several tropical regions reached a scale that may significantly change land cover and vegetation over large areas through intense deforestation practices. The question that motivates this study is how these changes in land cover may affect the large-scale circulations in the tropics. The impact of tropical deforestation on atmospheric circulations is the immediate focus of this paper. But, at a more fundamental level, this study is about the role of vegetation in driving large-scale atmospheric circulations in the tropics. Deforestation introduces a perturbation in the state of vegetation; hence by studying the response of the atmosphere to large-scale deforestation, we will try to define the role of vegetation in sustaining circulations in the tropics.

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Tropical deforestation is the most significant ongoing activity that modifies land surface characteristics over large areas of planet Earth. The United Nations Food and Agriculture Organization estimates that during the 1980s the tropical rain forests were degraded at a rate of about 46,000 km^2 per year [*Aldhous*, 1993]. Almost half of this damage occurred in the Amazon region [*Skole and Tucker*, 1993]. The annual deforestation rate in South America is 0.6%. Continental Southeast Asia and Central America are loosing larger proportions of their forests: 1.6 and 1.5% per year, respectively. Overall, the area of the tropical rain forests is getting smaller at a rate of 0.6% per year. The study of *Myers* [1991] predicts that most of the forest in the regions of southern and southeastern Asia, Central America, and East and West Africa will be lost by the year 2000. Tropical deforestation accelerated in recent decades due to the increase in population and the urgent need for economic development. These two factors are likely to continue in the near future.

Recent field observations from the Amazon region confirmed that conversion of land cover from rain forests to ranch land increases surface albedo and reduces evaporation rates [*Bastable et al.*, 1993; *Wright et al.*, 1992]. These local impacts modify the energy balance at the land-atmosphere boundary and potentially result in regional climate change. Several studies focused on the impact of tropical deforestation on land surface processes and climate. *Dickinson and Henderson-Sellers* [1988], *Lean and Warrilow* [1989], *Nobre et al.* [1991], *Dickinson and Kennedy* [1992], *Henderson-Sellers et al.* [1993], *Lean and*

Rowntree [1993], *Diremeyer and Shukla* [1994], and *Walker et al.* [1995] performed numerical experiments using general circulation models (GCMs) to simulate the impact on climate due to deforestation in the Amazon basin. Most of these studies predict that large-scale deforestation will result in the following regional climatic changes: warmer surface temperature, less evaporation, and less rainfall. The average change in temperature is of the order of 2°C. The average changes in rainfall and evaporation are of the order of -20 and -30%, respectively. The hypothetical deforestation scenario considered by these GCM studies assumes replacement of the total area of the Amazon rain forest ($\sim 10^6$ km^2) by short grass. These conditions represent the worst possible scenario that could occur if the current rates of deforestation continue into the next century.

The results of the modeling studies seem to agree in the predictions about the impact of deforestation on rainfall, surface temperature, and evaporation. However, although most of the recent studies agree on the sign of the change in the large-scale circulation following deforestation, there is no generally accepted theory on how deforestation affects the large-scale atmospheric circulations. Any change in the atmospheric circulation dictates a similar change in runoff from a large area. Runoff is the convergence of atmospheric water vapor, equivalent to the difference between precipitation and evaporation (neglecting changes in storage at the surface and the subsurface). In addressing this problem, *Eltahir and Bras* [1993] use a linear model to describe the response of the tropical atmosphere to large-scale deforestation. Two competing responses, or mechanisms, are suggested: (1) a thermally direct converging circulation in the boundary layer driven by the increase in surface temperature and (2) a diverging circulation in the boundary layer due to the corresponding decrease in rainfall and latent heating. The first mechanism is similar to that of *Lindzen and Nigam* [1987] which describes the response of the dry tropical atmosphere to gradients in sea surface temperature. The second mechanism is based on representing the effects of water vapor, in an otherwise dry atmosphere, by including the latent heating due condensation in the upper atmosphere [*Gill*, 1980]. The study of *Eltahir and Bras* [1993] explains the evident sensitivity of GCMs results regarding the changes in runoff and circulation as due to the competition between these two different responses. However, the same study does not address the question regarding the sign of the overall change in the total atmospheric circulation following large-scale deforestation. Instead, the final result is left for the competition between the two mechanisms. The origin for the concept that deforestation excites two different and competing responses lies in the separation between the thermodynamics of atmospheric water vapor from the dynamics of the dry atmosphere. In contrast, this paper will consider the unified dynamics of the moist tropical atmosphere. By taking this step, we will address the question of what happens to the total atmospheric circulation following deforestation. The immediate objective of this study is to develop some basic understanding regarding the mechanisms linking deforestation and the change in the large-scale atmospheric circulations. The three largest rain forests are located in the Amazon, Congo, and Indonesia. This paper focuses on these tropical regions and addresses several important questions. For example, will deforestation in the Amazon weaken or strengthen the circulation with the Atlantic Ocean? In another region we may ask the question: what is the impact, if any, of deforestation in Indonesia on the circulation in that region?

This paper describes a theoretical approach to the relation between vegetation and large-scale circulations. The main argument in this paper suggests that vegetation sustains and enhances large-scale atmospheric circulations. In other words, we argue that large-scale deforestation weakens atmospheric circulations. The relative reduction in boundary layer entropy compared to the surroundings causes the weakening of the circulation. In section 2 this reduction in entropy is traced back to the fact that deforestation reduces net surface radiation and the total flux of latent and sensible heat. Observational evidence is discussed to support the fact that deforestation reduces net surface radiation. Similar evidence is drawn from numerical experiments regarding the impact of deforestation on climate. In section 3, a quasi-equilibrium balance is assumed between moist convection and radiative forcing. This equilibrium is similar to the concepts suggested by *Arakawa and Schubert* [1974] and *Emanuel et al.* [1994] in treating the dynamics of convection in the tropical atmosphere. The main advantage of the equilibrium assumption is that it provides a unique relation between the surface entropy and the vertical distribution of temperature. In section 4 the response of the tropical atmosphere to large-scale deforestation is reduced to the problem of studying the impact on the circulation due to changes in the surface entropy. The linear response of the tropical circulation to the change in surface entropy is simulated using a simple model of atmospheric flow. Section 5 describes a general mechanism for relating deforestation and the change in the circulation. In section 6, deforestation scenarios that assume total removal of the rain forests in the regions of the Amazon, Congo, and Indonesia are considered. The response of the tropical atmosphere to large-scale deforestation in these regions is studied with the purpose of inferring the role of vegetation in sustaining large-scale atmospheric circulations in the tropics.

2. Tropical Deforestation, Net Surface Radiation and Boundary Layer Entropy

The boundary layer entropy over any region is increased by surface fluxes of latent and sensible heat and decreased by a combination of three processes: entrainment at the top of the boundary layer, convective downdrafts, and radiative cooling of the boundary layer air. These processes are described schematically in Figure 1. Deforestation affects the boundary layer entropy primarily through the change in the total flux of latent and sensible heat from the surface into the atmosphere. The sign and magnitude of the change in boundary layer entropy depends on how the total flux of heat, including latent and sensible forms, may change after deforestation.

Deforestation modifies the surface energy balance. The change from forest to short grass or bare soil will increase the relative magnitude of sensible heat flux compared to the latent heat flux resulting in a larger Bowen ratio. The reduction in evaporation follows mainly from the smaller root-zone depth associated with short grass in comparison to forest. However, the less obvious question is how deforestation changes the magnitude of the total flux of latent and sensible heat. In the following, this question will be addressed by considering the energy balance at the boundary between the land surface and the atmosphere. Before deforestation the equilibrium state of

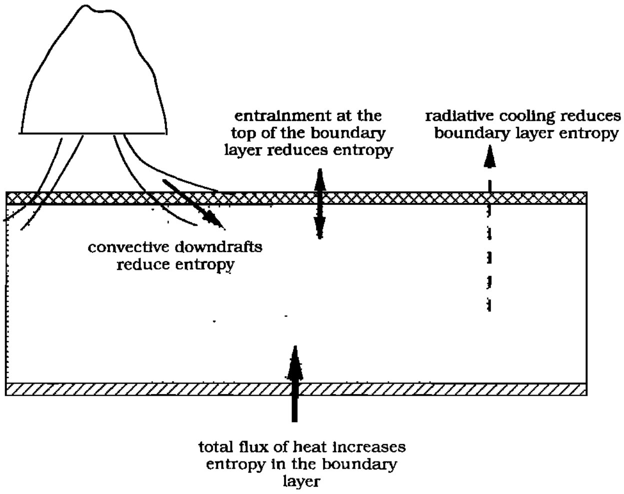

Figure 1. Fluxes of boundary layer entropy.

the energy balance at any point in the boundary between the land surface and the atmosphere is described by

$$N - F = 0 \tag{1}$$

where N is net surface radiation and F is the total flux of heat from the surface, including both the latent and the sensible forms. Each of the two terms N and F has two components,

$$N = N_s + N_t$$
$$F = \lambda E + H \tag{2}$$

where N_s is net solar radiation equivalent to incident solar radiation minus reflected solar radiation; N_t is net terrestrial radiation defined as downward flux minus upward flux of terrestrial radiation; E is evaporation; λ is the latent heat of vaporization; and H is sensible heat flux.

The new equilibrium of the surface energy balance that follows deforestation is described by

$$\lambda(E + \delta E) + (H + \delta H) = (N_s + \delta N_s) + (N_t + \delta N_t) \tag{3}$$

where δ preceding any of the terms denotes a small change in that variable due to deforestation. Subtraction of (2) from (3) results in

$$\delta F = \lambda \delta E + \delta H = \delta N_s + \delta N_t = \delta N \tag{4}$$

This equation suggests that the change in the total flux of heat F is exactly the same as the change in net surface radiation, including both components: solar and terrestrial radiation.

The recent results from the Anglo-Brazilian Amazonian Climate Observation Study (ABRACOS) provide field observations regarding the impact of deforestation on surface radiation. *Bastable et al.* [1993] presented the first comparative observations of the micrometeorology over cleared and undisturbed forest in the Amazon region. The undisturbed forest site is Reserve Ducke near Manaus. The cleared forest site is Fazenda Dimona, a cattle ranch located 100 km north of Manaus. The observations cover a continuous 60-day period from the middle of October 1990 to the middle of December 1990. The data were collected at the two sites simultaneously. The net surface radiation measured at the two sites and averaged for the 60-day period is shown in Figure 2. The net surface radiation at the cleared site is less than net radiation above the

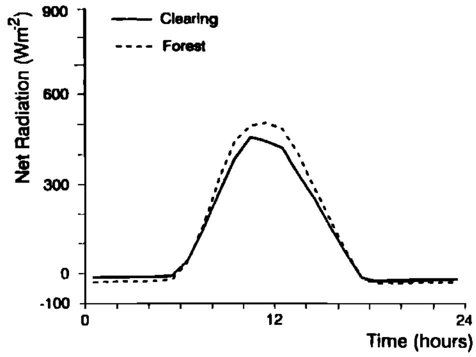

Figure 2. Observations of net radiation over clearing and forest in the Amazon region, from *Bastable et al.* [1993].

forest. Figure 3 shows the observations on the two components of net surface radiation: solar and terrestrial radiation. This figure shows the differences in net solar (terrestrial) radiation between the forest and the clearing. The field observations suggest that the difference in net surface radiation between the forest and the cleared site is contributed, almost equally, by both components: solar radiation and terrestrial radiation. The field observations provide the most compelling evidence on how deforestation modifies surface radiation. But since the scale of the clearing at Fazenda Dimona is small (~10 km^2), some of the atmospheric feedbacks which follow large-scale deforestation are not yet observed. In particular, no changes were detected in the patterns of rainfall and cloudiness. Nevertheless, the observations include other direct impacts of deforestation such as those involving changes in surface temperature, atmospheric water vapor, and atmospheric emissivity.

The impact of deforestation on surface radiation can be investigated using a numerical modeling approach. The numerical simulations of *Eltahir and Bras* [1994] focused on the impact on climate due to deforestation of small areas in the middle of the Amazon rain forest. This recent study used a mesoscale climate model, that is driven by solar radiation and

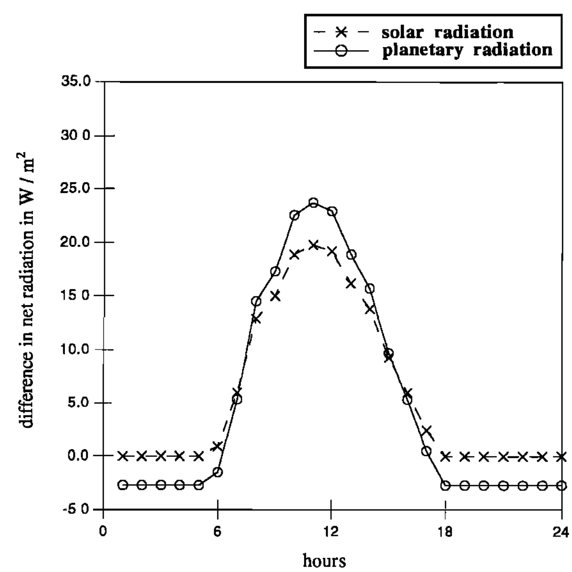

Figure 3. Observations of the reduction in net solar and terrestrial radiation between forest and clearing, reproduced on the basis of information from *Bastable et al.* [1993].

Table 1. Changes in Net Radiation, Net Solar Radiation, Net Terrestrial Radiation Based on Observations and Numerical Simulations

Study	Scale of Deforestation, km^2	Change in Net Surface Radiation, W/m^2	Change in Net Solar Radiation, W/m^2	Change in Net Terrestrial Radiation, W/m^2
Field Observations				
Bastable et al. [1993]	10	−11	−6	−5
Numerical Simulations				
Lean and Warrilow [1989]	10^6	−21	···	···
Nobre et al. [1991]	10^6	−26	−18	−8
Dickinson and Kennedy [1992]	10^6	−18	−3	−15
Lean and Rowntree [1993]	10^6	−19	−4	−15
Eltahir and Bras [1994], January	10^5	−15	−8	−7
Eltahir and Bras [1994], July	10^5	−12	−5	−7

observed boundary conditions, to perform several experiments on the deforestation problem. The numerical experiments consisted of control and deforestation runs and were performed over a domain with a total area of about 2.6×10^6 km^2. The size of the cleared area is of the order of 0.6×10^5 km^2, located around the center of the simulation domain. The results of these numerical simulations agree with the observations of *Bastable et al.* [1993] about the changes in the two components of net surface radiation between cleared and undisturbed forest. Net surface radiation over cleared areas is less than that over undisturbed sites. Several other studies used different GCMs to investigate the deforestation problem in the Amazon region. Table 1 summarizes the results of several of these numerical experiments. Most of these studies consider the impact of deforestation of large areas ($\sim 10^6$ km^2). However, the study of *Eltahir and Bras* [1994] considers deforestation of a relatively small area compared to the other studies resulting in relatively small changes in surface temperature and net radiation. In general, the results of these different numerical experiments seem to agree in the prediction that large-scale deforestation reduces net surface radiation and hence the total flux of heat by about 20 W/m^2.

The change in net surface radiation has two components: a change in net solar radiation and a change in net terrestrial radiation. The change in net solar radiation has three main potential causes: (1) a change in surface albedo, (2) a change in cloud albedo, and (3) changes in atmospheric absorption by water vapor and clouds. Deforestation increases surface albedo, but the resulting change in rainfall and cloudiness decreases cloud albedo and atmospheric absorption. The net result is near cancelation of the two effects. The cloudiness feedback can be appreciated by comparing the results of the study by *Nobre et al.* [1991], which uses a fixed and prescribed cloudiness climatology, with those of *Dickinson and Kennedy* [1992] and *Lean and Rowntree* [1993], which use interactive cloud schemes; see Table 1. The change in net solar radiation in the former study is significantly larger compared to those of the latter studies. This difference can be attributed to the clouds feedback. On the other hand, the results of *Eltahir and Bras* [1994] and *Bastable et al.* [1993] involved small to very small scales of deforestation resulting in small or almost no change in cloudiness. The corresponding relatively large changes in net solar radiation that are reported in Table 1 are consistent with the small changes in cloudiness. Based on the results of all these studies, we conclude that the negative feedback due to the changes in cloudiness and atmospheric absorption of solar radiation is indeed significant. The change in net solar radiation following large-scale deforestation is likely to be of a small magnitude due to this negative feedback.

The corresponding change in net surface terrestrial radiation has three main potential causes: (1) changes in the temperature of the surface and the boundary layer, (2) changes in atmospheric emissivity and absorptivity in the boundary layer which are caused by changes in concentration of water vapor, and (3) changes in reflection and absorption by clouds. Deforestation increases temperature in the lower layers of the atmosphere. This effect tends to enhance upward terrestrial radiation and hence reduces net terrestrial radiation. At the same time, deforestation results in drying of the boundary layer which reduces the greenhouse effect of water vapor resulting in less net terrestrial radiation at the surface. The reduction in cloudiness following deforestation is a positive feedback which reinforces the reduction in net terrestrial radiation. In summary, these three processes tend to excite and reinforce a reduction in net terrestrial radiation at the surface. The change in net terrestrial radiation following large-scale deforestation is the dominant process in modifying the radiation balance.

The net impact of deforestation on the radiation balance at the land-atmosphere boundary is to reduce net surface radiation. When deforestation occurs over a large area, most of the reduction in net radiation comes from the change in terrestrial radiation. But when deforestation covers medium to small scales, the changes in solar and terrestrial radiation seem to contribute similar magnitudes to the overall change in net surface radiation. The energy balance at the land-atmosphere boundary requires that any reduction in net surface radiation has to be balanced exactly by a similar reduction in the total flux of heat from the surface, including latent and sensible forms. Hence large-scale deforestation should result in smaller total flux of heat from the surface into the boundary layer.

The change in surface entropy following deforestation is primarily caused by the change in the total flux of heat. Equation (4) suggests that the change in the total flux of heat following deforestation is equivalent to the change in net surface radiation. Given that is true, then the change in boundary layer entropy will follow the change in net radiation at the surface. The potential feedbacks due to the changes in the three other processes (see Figure 1) that control boundary layer entropy are relatively small compared to the effects on entropy due to the change in total flux of heat. However, these feedbacks are likely to introduce additional reduction in

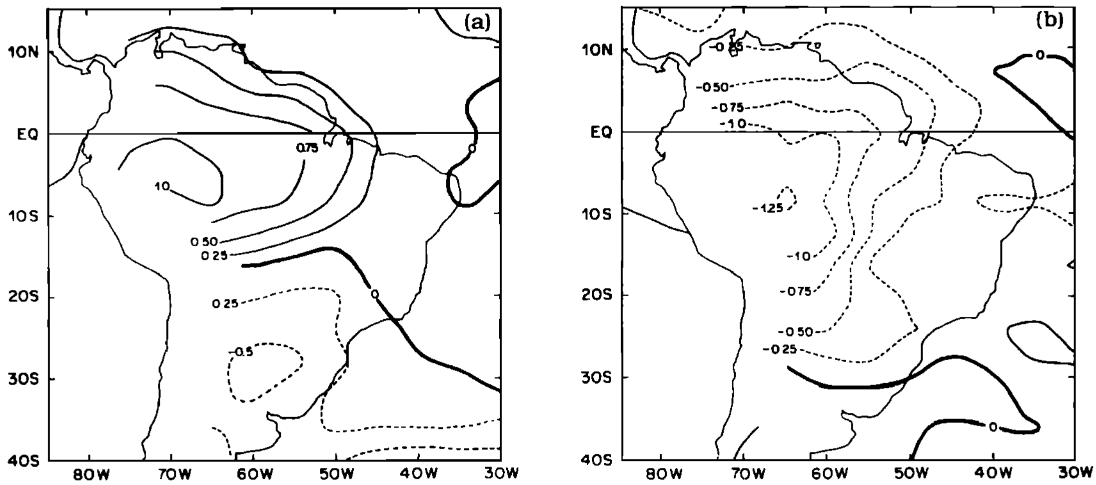

Figure 4. Changes in (a) specific humidity at 850 mbar in g/kg and (b) temperature at 850 mbar in degrees Celsius, from *Nobre et al.* [1991].

boundary layer entropy. If, initially, deforestation reduces boundary layer entropy, then that initial effect is likely to result, through the mixing induced by moist convection, in lower entropy at all levels of the atmosphere. Hence entrainment at the top of the boundary layer and convective downdrafts are likely to mix into the boundary layer, air of an even lower entropy resulting in additional lowering of boundary layer entropy. Since deforestation increases surface temperature, the intensity of entrainment at the top of the boundary layer is likely to increase resulting in further reduction of boundary layer entropy and causing a positive feedback. On the other hand, the same increase in boundary layer temperature enhances radiative cooling and results in an additional sink of entropy. One potential negative feedback that may increase boundary layer entropy is the reduction in convective downdrafts following deforestation. A reduction in the intensity of convective downdrafts would be consistent with the predicted decrease in rainfall. However, a single negative feedback is likely to have a small effect compared to the combination of several positive feedbacks that are equally important.

Although the impact of deforestation on boundary layer entropy is of primary importance for understanding how deforestation affects atmospheric circulations, almost all the numerical simulations on the deforestation problem are reported without reference to the change in surface entropy or boundary layer entropy. Similarly, there is usually no mention of the change in specific humidity. The study of *Nobre et al.* [1991] is the only exception: Figure 4, which is taken from that study, shows the changes in temperature and specific humidity at 850 mbar. The changes in specific humidity are of the order of -1 g/kg. The corresponding changes in temperature are of the order of 1°C. To estimate the impact of these changes on boundary layer entropy, we note that

$$\frac{\dfrac{\partial s}{\partial q}}{\dfrac{\partial s}{\partial T}} \approx \frac{\lambda}{C_p} \tag{5}$$

where s is entropy, T is temperature, q is specific humidity, λ is latent heat of vaporization, and C_p is specific heat capacity of air at constant pressure. Equation (5) implies that a decrease in specific humidity of 1 g/kg can only be offset by an increase in temperature of about 2.3°C. Hence the results in Figure 4 imply a reduction in entropy equivalent to a thermal cooling of about 1° to 2°C. The simulations of *Eltahir and Bras* [1994] predict a reduction in entropy with a similar order of magnitude. Table 2 shows estimates of the reduction in boundary layer entropy due to deforestation. These estimates are based on the results of *Nobre et al.* [1991] and *Eltahir and Bras* [1994]. The reduction in boundary layer entropy following deforestation ranges between 1 and 10 J/kg/°K.

The observations of *Bastable et al.* [1993] describe the conditions that may result from small-scale deforestation (~10 km). At such small scales, advection has a significant effect on the observed specific humidity and temperature. Although field observations show significant drying over the clearing site relative to the forest conditions, the same observations indicate very little change in entropy. Indeed, the change in entropy should become significant when deforestation occurs at large enough scales. In contrast, the numerical simulations of *Eltahir and Bras* [1994] consider deforestation of a large area with the size of about 10^5 km^2. The other numerical simulations consider deforestation of areas that are even larger (~10^6 km^2). At such large scales, deforestation changes the boundary layer entropy. The analysis in this paper is relevant to deforestation at large scales (~10^6 km^2).

3. Relation Between Boundary Layer Entropy and the Vertical Profile of Temperature

The climate of the rain forests in the Amazon, Congo, and in Indonesia is characterized by some of the highest levels of

Table 2. Changes in Boundary Layer Entropy Estimated on the Basis of Information Provided in Two Studies

Study	Change in Boundary Layer Entropy, J/kg/°K
Nobre et al. [1991]	−8
Eltahir and Bras [1994], January	−7
Eltahir and Bras [1994], July	−3

convection and rainfall. Under these conditions, radiative processes and moist convection are the primary processes in determining the vertical profile of temperature. However, the relation between the surface entropy and the vertical profile of temperature is dependent on how convection interacts with the large-scale environment. Moist convective adjustment was suggested by *Manabe et al.* [1965] as a simple approach to the issue of how convection modifies the thermodynamic properties of the atmosphere at large scales. This scheme suggests that when the atmospheric lapse rate tends to exceed the moist adiabatic lapse rate, the intensity of free convection is strong enough to maintain a neutral lapse rate of the equivalent potential temperature. According to this assumption, the vertical profile of temperature is uniquely determined by the surface entropy. *Manabe et al.* [1965] chose this scheme as a simple approach to a problem which was very little understood at that time.

Arakawa and Schubert [1974] studied the interaction of cumulus cloud ensemble with the large-scale environment. They developed the concept of quasi equilibrium between the cumulus ensemble and the large-scale forcing. This condition is satisfied when the timescale for adjusting atmospheric conditions by convection is significantly shorter than the timescale of the large-scale forcing. This condition is satisfied in the regions of interest for this study. The adjustment timescale is of the order of hours and the timescale for large-scale radiative forcing is of the order of days. *Arakawa and Schubert* [1974] and *Lord and Arakawa* [1980] present observations from the tropics which seem to verify the quasi-equilibrium assumption. Under these conditions the vertical profile of temperature undergoes a series of quasi-neutral states in time. The idea that convection is nearly in statistical equilibrium with the environment was emphasized in the recent work of *Emanuel et al.* [1994]. They suggest that the vertical temperature profile in convecting atmospheres is controlled by convection and is uniquely related to the subcloud layer entropy.

In this paper we use the quasi-equilibrium concept to relate the vertical profile of temperature and surface entropy. The impact of deforestation on the tropical atmosphere is felt first at the land-atmosphere boundary as a reduction in entropy. Then, the postdeforestation temperature profile has to be determined by the postdeforestation surface entropy. Implicit in this argument is the assumption that changes due to deforestation are small and will result in conditions where convection is still efficient enough to control the vertical profile of temperature. The results of the GCM studies summarized in the introduction suggest that the potential reductions in rainfall are of the order of 20% or less. Hence we feel that the stated assumption is reasonable. Under the quasi-equilibrium assumption the reduction in boundary layer entropy following deforestation dictates a relative cooling in the temperature of the upper atmosphere. Such changes in the vertical profile of atmospheric temperature would necessarily result in changes in the atmospheric circulation. The nature of these changes is discussed in section 4.

4. Impact of the Change in Boundary Layer Entropy on Large-Scale Atmospheric Circulations

The objective in this section is to develop a simple linear model which relates the changes in boundary layer entropy caused by deforestation to the changes in the large-scale circulation. This model is developed following some of the concepts introduced by *Matsuno* [1969], *Gill* [1980], *Neelin* [1988], and *Eltahir and Bras* [1993]. However, some important aspects of the model development are different and represent significant modifications: (1) the treatment of energy balance in the upper atmosphere is different from previous studies; (2) the dynamics of the moist atmosphere are considered in this study; in contrast, most of the previous studies with the Gill model assumed a dry atmosphere; (3) the quasi-equilibrium assumption, which is adopted in this study, is quite different from treating latent heating as external forcing, in an otherwise dry atmosphere, as implied by some of the earlier work.

The focus of this section is on the flow in the atmospheric boundary layer at equilibrium conditions. The horizontal momentum equations are approximated by assuming a balance among Coriolis force, pressure gradient force, and momentum dissipation,

$$fv' - \partial_x \Phi' - \varepsilon u' = 0 \tag{6}$$

$$-fu' - \partial_y \Phi' - \varepsilon v' = 0 \tag{7}$$

The variables and parameters in the above equations are defined as follows: u' and v' are perturbations in the horizontal components of mass flux in the boundary layer, Φ' is the mass-weighted integral of the perturbation in geopotential height in the boundary layer, ε is the coefficient of Rayleigh friction, and f is the Coriolis parameter. The two equations are linearized about a basic state of no motion. Continuity of the flow in the boundary layer can be described, in pressure coordinates, by

$$\partial_x u' + \partial_y v' - \frac{w'}{g} = 0 \tag{8}$$

where w' is the perturbation to the vertical velocity field above the boundary layer and g is gravitational acceleration.

In a moist atmosphere that satisfies the quasi-equilibrium hypothesis and is free of any large-scale vertical motion, the latent heating due to moist convection is exactly balanced by the large-scale radiative forcing. The direct effect of moist convection is to mix the tropical atmosphere resulting in an almost constant vertical distribution of saturation entropy. This condition is achieved at short timescales characteristic of the moist convection process. As a result, at the timescales of large-scale motions the magnitude of boundary layer entropy determines the vertical profile of atmospheric temperature. This temperature distribution describes a radiative convective equilibrium condition and will be denoted by T_e.

When a large-scale circulation develops in a moist atmosphere that satisfies the quasi-equilibrium hypothesis, the principle of conservation of energy requires that any resulting adiabatic heating (cooling) due to the large-scale vertical motion is exactly balanced by an additional radiative cooling (heating) induced by the vertical motion itself. In other words, at the timescale of the large-scale circulation the energy balance for the upper atmosphere (above the boundary layer) is between adiabatic heating (cooling) due to vertical motion and additional radiative cooling (heating). Under these conditions a simplified statement of energy conservation for the upper atmosphere is described by the following:

$$\frac{w'}{g} = \frac{\alpha_n (T' - T_e')}{C^2} \tag{9}$$

The variables and parameters in this equation are defined as follows: T' is the perturbation in the temperature field of the upper atmosphere. The additional radiative cooling (heating) is proportional to the difference between the perturbation in actual temperature and the perturbation in equilibrium temperature. This difference is caused by the large-scale vertical motion. The parameter α_n is the coefficient of Newtonian cooling. The wave speed C is defined by the relation $C^2 = SR\Delta p/2$, where S is assumed equivalent to the moist neutral lapse rate, defined as the rate of change of temperature with respect to the pressure coordinate; R is the ideal gas constant for dry air; Δp is about half the depth of the troposphere corresponding to the region where radiative cooling takes place. In discussing the energy balance of the tropical atmosphere, the role of advection is neglected. This assumption is justified by the observations that temperature gradients are very small in the tropics and that wind speed is relatively small; hence the effects of heat advection should be relatively small. The two equations (8) and (9) will be combined into

$$C^2(\partial_x u' + \partial_y v') = \alpha_n(T' - T'_e) \tag{10}$$

In the spirit of the Gill model, the radiative cooling term, which is due to the perturbation in temperature, will be assumed equivalent to a product of the perturbation in geopotential and the coefficient of Rayleigh friction,

$$\alpha_n T' = -\varepsilon\Phi' \tag{11}$$

A similar assumption was made by *Gill* [1980]. The motivation for making this approximation is analytical simplicity. As pointed out by *Neelin* [1988], unless the equations were used to describe the total atmospheric flow, this approximation has negligible impact on the solution. In this study we intend to look at the impact of deforestation on the circulation, which is presumably small in comparison to the total circulation. For such application this crude assumption may have only a minor effect.

The next step in this derivation is to relate the perturbation in equilibrium temperature T'_e to the perturbation in boundary layer entropy s'_b. The equation of state can be expressed in the following form:

$$T'_e)_p = \frac{p}{R}\,\alpha')_p \tag{12}$$

where α is specific volume and p is pressure. The Maxwell's relations [*Emanuel*, 1994] imply that

$$\alpha')_p = s')_p \frac{\partial T}{\partial P} = s')_p S \tag{13}$$

where s' is perturbation in saturation entropy and S is static stability. Under the quasi-equilibrium assumption the perturbation in saturation entropy is constant in the vertical and is equivalent to a perturbation in boundary layer entropy s'_b. Substitution of (11), (12), and (13) into (10) results in

$$\varepsilon\Phi' + C^2(\partial_x u' + \partial_y v') = -\gamma s'_b \tag{14}$$

where γ is a constant parameter that is defined by the following equation:

$$\gamma = \left(\frac{\alpha_n S p}{R}\right) \tag{15}$$

where p is pressure at the tropospheric level where most of the radiative cooling takes place (~300 mbar). This analysis provides a closed set of equations: (6), (7) and (14), on the three variables u', v', and Φ'. The forcing for this set of equations is given by the perturbation in boundary layer entropy.

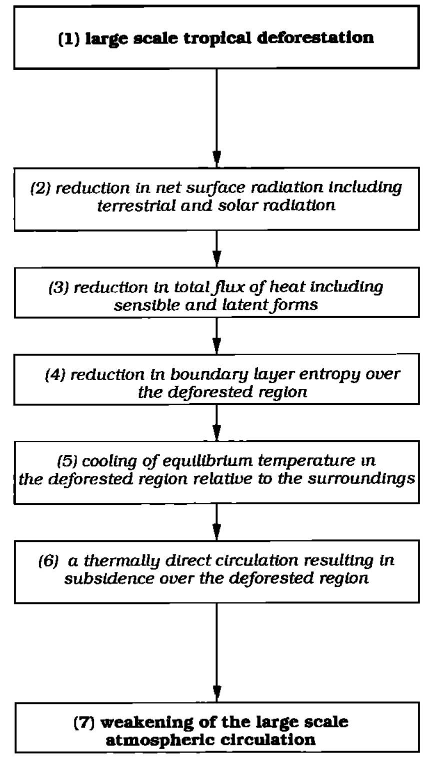

Figure 5. A mechanism relating deforestation and the weakening of large-scale atmospheric circulations.

5. A Mechanism Relating Tropical Deforestation and the Change in Large-Scale Atmospheric Circulation

The following mechanism is proposed for relating deforestation and the change in large-scale atmospheric circulations. The mechanism is illustrated in Figure 5. Deforestation modifies the surface energy balance and reduces the net surface radiation. The surface energy balance dictates that a reduction in net surface radiation should result in reducing the total flux of heat from the surface, including latent and sensible forms. The latter causes a reduction in the boundary layer entropy. Under the quasi-equilibrium assumption a reduction in the boundary layer entropy modifies the vertical profile of equilibrium temperature and causes cooling of the upper atmosphere over the deforested region relative to the surroundings. This relative cooling drives a thermally direct circulation causing a sinking motion over the deforested region. The principle of mass conservation suggests that subsidence over the deforested region is consistent with a converging circulation in the upper atmosphere and a diverging circulation in the boundary layer.

The spatial distribution of equilibrium temperature is the forcing for atmospheric circulations over large rain forests. The observed distribution of net surface radiation has maximum points over the rain forests; see *Budyko* [1986], which favors similar distributions of boundary layer entropy (see Figure 6) and equilibrium temperature of the upper atmosphere. The latter drives thermally direct atmospheric circulations over the rain forests that are convergent in the boundary layer and divergent in the upper atmosphere. Since deforestation results

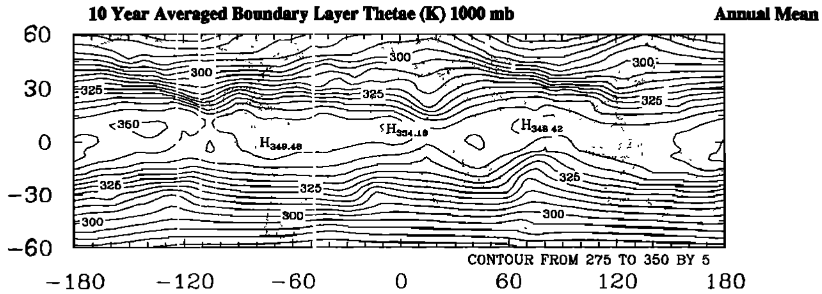

Figure 6. Global distribution of equivalent potential temperature at 1000 mbar. Averages are based on the GFDL data set for the years 1979–1989.

in sinking motion over the deforested region, this anthropogenic change should weaken large-scale atmospheric circulations. These changes in the circulation are discussed in section 6.

In the following, we compare the mechanism of Figure 5 to the theory of *Charney* [1975] who presented a mechanism for explaining the occurrence of droughts in West Africa. Charney's theory is based on the basic concept that deserts work as radiative sinks of energy where the loss of heat due to the emission of terrestrial radiation is balanced by adiabatic warming due to subsidence. Under these conditions, removal of vegetation increases surface albedo, causing an additional radiative cooling. The latter can only be balanced by additional adiabatic warming and hence results in enhancement of sinking motion. The main similarity between the mechanism proposed in this paper and that of Charney is the suggestion that degradation of land cover and vegetation would eventually induce sinking motion. The main important differences between the two mechanisms are the following: (1) the mechanism presented in this paper explicitly considers the physical processes in a moist atmosphere as opposed to the assumption of dry atmosphere that is implicit in the theory of *Charney* [1975]. Indeed, this important difference makes the proposed mechanism more relevant to the dynamics of the tropical atmosphere over the rain forests. (2) The proposed mechanism emphasizes the role of changes in both solar and terrestrial radiation through their impact on the total flux of latent and sensible heat. Hence the direct impact of deforestation is not limited to the changes in surface albedo, as assumed in the Charney mechanism. This component of the proposed mechanism is consistent with the recent field observations of *Bastable et al.* [1993] which confirm that deforestation changes both solar and terrestrial radiation at the surface. (3) The proposed mechanism acknowledges the role of boundary layer entropy in driving large-scale atmospheric circulations. This feature is consistent with some of the most recent dynamical theories of large-scale tropical circulations; see *Emanuel et al.* [1994].

6. Impact of Tropical Deforestation on Circulations of the Amazon, Congo, and Indonesia

This section focuses on the relation of deforestation to the large-scale atmospheric circulations in three regions: the Amazon, Congo, and Indonesia. The forcing due to deforestation and the resulting change in boundary layer entropy are assumed to follow the form given by

$$s_b' = s_{b0}' \cos(kx)(1-y)\exp\left(-\frac{y^2}{4}\right), \ |x| < L;$$

$$s_b' = 0, \ |x| > L. \tag{16}$$

where x and y are distances in the zonal and meridional directions respectively, both normalized by the equatorial Rossby radius $(C/2\beta)^{1/2}$, $(\beta = (\partial f/\partial y)|_{y=0}$, $\beta = 2.3 \times 10^{-11}$ m^{-1} s^{-1}). For the set of parameters used in this study, the Rossby radius is about 12° latitude. Figure 7 describes the form and locations of the forcing function over the three regions. The magnitudes shown in the figure are dimensionless and do not include the factor s_{b0}' of (16). The variable s_{b0}' denotes the magnitude of the forcing at a center point which is taken at the intersection of the equator with 60°W for the Amazon, 20°E for the Congo, and 110°E for Indonesia. The length scale L is assumed to be about 15° latitude ($k = \pi/2L$). A linear set of equations similar to (6), (7), and (14) have been solved analytically by *Gill* [1980]. These analytic solutions assume a forcing that is described by the mathematical form of (16). The solutions are described in detail by *Gill* [1980] and *Eltahir and Bras* [1993]. The change in entropy s_{b0}' is assumed to be about 10 J/kg/°K.

To assess the significance of the predicted changes in the circulation due to deforestation, we compare these changes with observations from the three regions. Figures 8a, 9a, and 10a describe observations of boundary layer mass flux for the Amazon, Congo, and Indonesia, respectively, from the Geophysical Fluid Dynamics Laboratory (GFDL) global data set. This data set is based on rawinsonde observations. Earlier versions of this data set are described by *Oort* [1983]. The GFDL data set has a resolution of 2.5° along longitudes and 5° along latitudes. The boundary layer mass flux is defined as the average for the following pressure levels: 1000, 950, 900, and 850 mbar. The observations represent the average fluxes for the year 1989. For comparison purpose the Figures 8b, 9b, and 10b describe the change of mass flux in the boundary layer due to deforestation in the Amazon, Congo, and Indonesia. These changes are simulated by the linear model developed in this study. The parameters used in estimating these fluxes are given in Table 3. The main features of the change in boundary layer flow that are simulated by the simple linear model are consistent with the results from the nonlinear GCM simulations of *Nobre et al.* [1991]. This is particularly true for the cyclonic circulation that is induced in the area southwest from the deforested region. The change in mass flux due to deforestation is divergent in the boundary layer and necessarily convergent in the upper atmosphere.

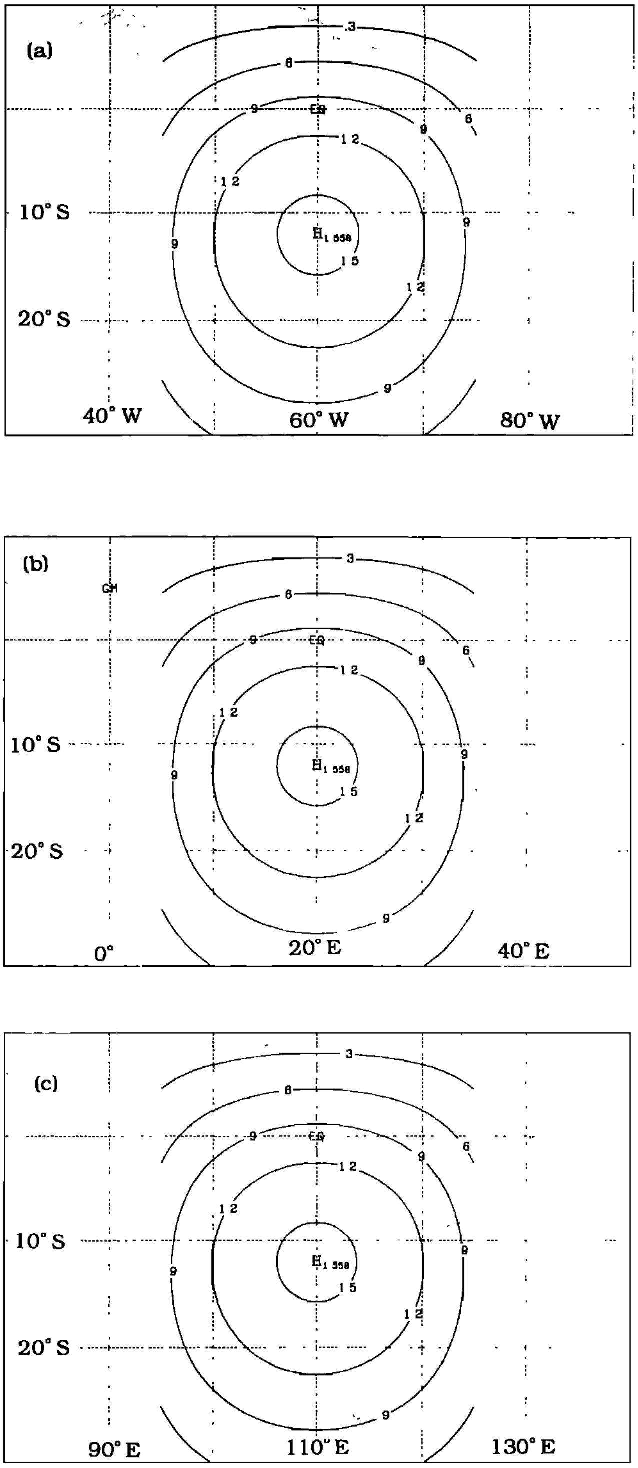

Figure 7. Forcing function for (a) the Amazon, (b) Congo, and (c) Indonesia.

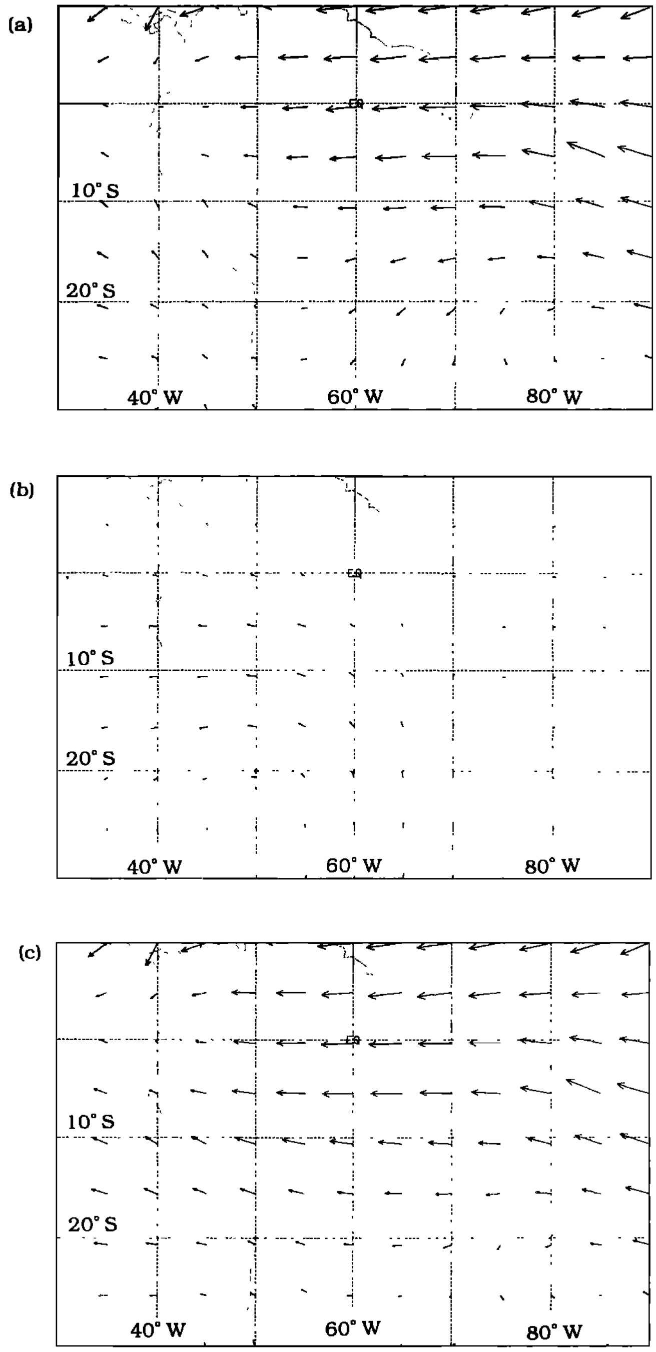

Figure 8. Mass flux in the boundary layer for the Amazon region, the largest vector represents about 1.0×10^4 kg/m/s: (a) observations, (b) predictions of the change due to deforestation, and (c) predictions of post-deforestation fluxes (a plus b).

Figures 8c, 9c, and 10c result from superposition of the previous two sets of figures (a and b), i.e., by adding the observed mass fluxes to the predicted change of mass flux in the boundary layer. The predictions of the changes in mass flux in the boundary layer are small compared to observations of the same fluxes. The difference is about a factor of 5 between the observations and the change in flux. The change in boundary layer circulation tends to strengthen the easterly flow west of the deforested region. The observations are dominated by strong global easterly flow at most of latitudes and longitudes and for the three regions. The boundary layer mass flux that contributes to any of the regional atmospheric circulations is imbedded within the global easterly flow. Hence these regional circulations are better defined by the spatial derivatives of the horizontal boundary layer mass flux.

The spatial derivatives of the horizontal boundary layer fluxes are related to the vertical velocity by (8). Figures 11a, 12a, and 13a describe the vertical velocity over the boundary

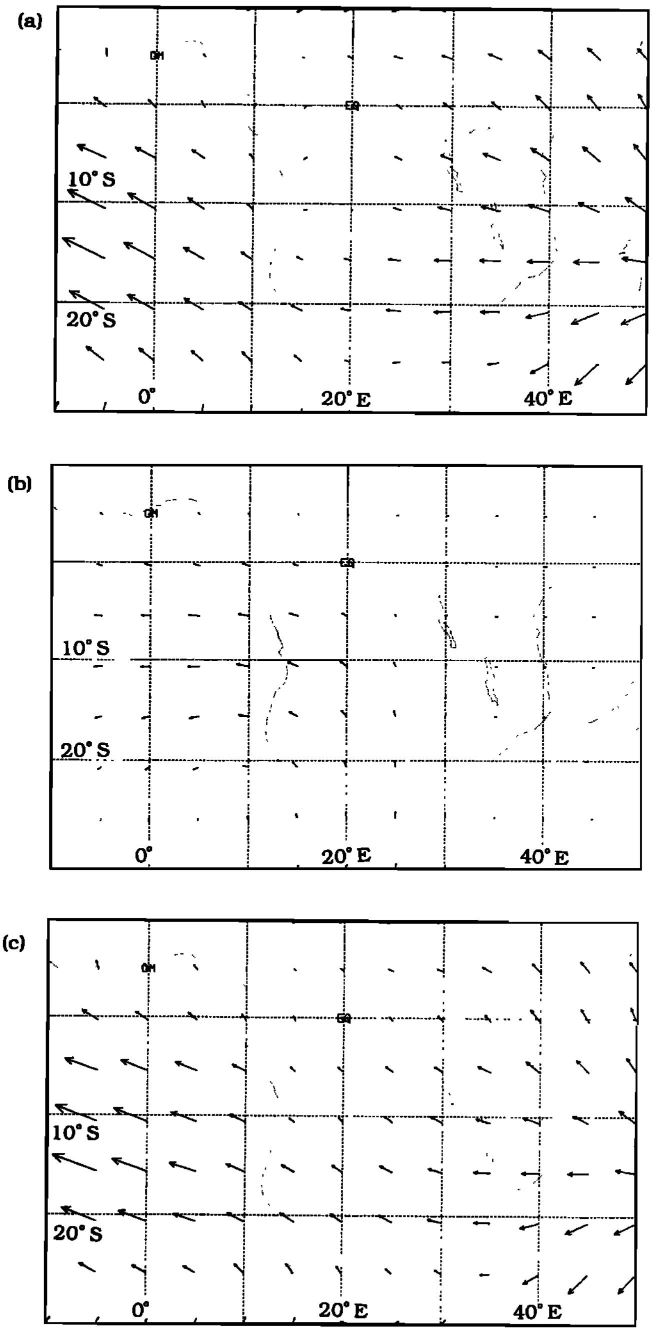

Figure 9. Same as Figure 7 but for Congo.

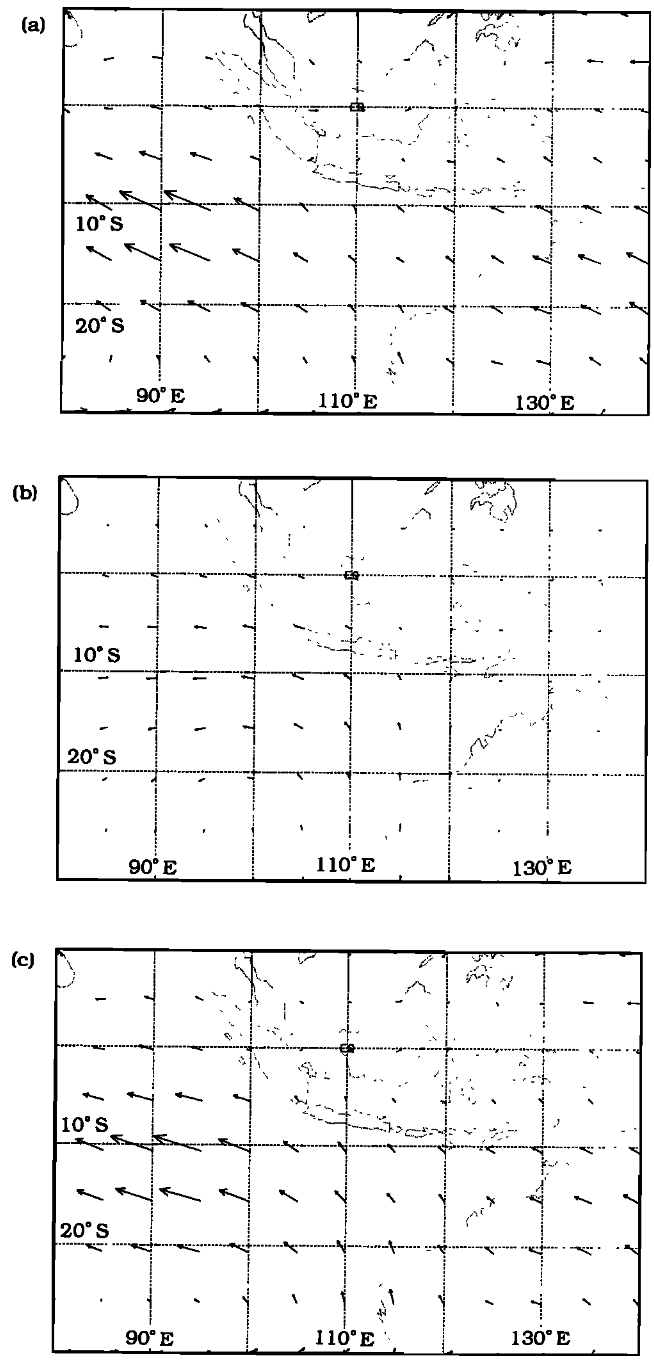

Figure 10. Same as Figure 7 but for Indonesia.

layer for the three regions from the GFDL observations. Since pressure coordinates are used in this analysis, a positive velocity indicates subsidence, and negative velocity indicates rising motion. The three regions of the Amazon, Congo, and Indonesia are characterized by rising motion over large areas. Figures 11b, 12b, and 13b describe the change in vertical velocity following deforestation. A positive velocity field, which indicates subsidence, covers each of the three deforested regions. The superposition of these two sets of figures (a and b) results in the predictions of vertical velocity in the postdeforestation climate as described by the set of Figures 11c, 12c, and 13c. The magnitudes of the changes in vertical velocity are indeed significant over the three regions in comparison to observations. However, the vertical velocity field over the Amazon is controlled to a large degree by the orographic lifting to the west. Hence although removal of vegetation changes the circulation significantly, orographic lifting would still maintain some of the large-scale rising motion over the area to the west and northwest of the Amazon. In the Indonesian region, removal of vegetation may change the circulation significantly over a large area. But the large-scale ascent remains the dominant feature to the east and west of Indonesia. This rising motion is maintained largely by the warm ocean temperature around Indonesia. In the absence of the other large-scale forcings such as orography or warm sea surface temperature the rain forests in the Congo region play a dominant role in sus-

Table 3. Parameters and Variables of the Linear Model

Parameter, Variable	Value	Unit
ε	0.5	day
$\dfrac{\partial T}{\partial z}$	−6	°K/km
$S\left(=\dfrac{\partial T}{\partial p}\right)$	1.0E−3	°K/pascal
Δp	500	mbar
$C\ (=(SR\Delta p/2)^{1/2})$	85	m/s
p	300	mbar
α_n	4.2	(W/m^2)/°K
$\gamma\left(=\left(\dfrac{\alpha_n Sp}{R}\right)\right)$	0.44	$\dfrac{(\mathrm{W/m^2})}{\mathrm{J/kg/°K)}}$
s'_{bo}	−10	(J/kg/°K)
Equatorial Rossby radius $(=(c/2\beta)^{1/2})$	1350	km
Timescale $(=(2\beta c)^{-1/2})$	0.2	day

Read 1.0E−3 as 1.0×10^{-3}.

taining the large-scale circulation. Hence in comparison to the other two regions the region of Congo should have the largest potential sensitivity to deforestation.

In developing the relation between the deforestation and the change in the circulation, we assume that the changes in vertical velocity are relatively small such that moist convection remains efficient enough to control the post-deforestation profile of atmospheric temperature. This level of approximation is consistent with the assumption of linearity in the atmospheric flow model. However, if the changes in vertical velocity are significant enough such that vertical mixing is inefficient, the quasi-equilibrium assumption breaks down and the response of the tropical atmosphere to any further forcing cannot be described by the theory of this paper. Under those conditions the response of the tropical atmosphere is described by the theory of *Eltahir and Bras* [1993]: the atmosphere responds to two forcings, the increase in surface temperature, and the reduction in the latent heating associated with condensation. In other words, the mechanism of section 5 and the mechanism proposed by *Eltahir and Bras* [1993] describe the response of the tropical atmosphere to deforestation at two different stages: the early stage, while the atmosphere satisfies the quasi-equilibrium assumption, and the late stage, following the breakdown of that assumption.

7. Conclusions

1. A mechanism has been proposed to describe how large-scale deforestation modifies atmospheric circulations in the tropics. This mechanism suggests that the change in moist entropy of the boundary layer, which results from changes in surface temperature and humidity, is the main agent in communicating the impact of deforestation on surface conditions to the large-scale circulation. However, most of the numerical studies on the deforestation problem do not report the changes in humidity or entropy. For the purpose of understanding the impact of deforestation on climate, a discussion on the nature of the changes in these two fields is useful and informative.

2. The changes in vertical velocity and boundary layer mass flux over the Amazon, Congo, and Indonesia are simulated using a linear model of the boundary layer flow. The comparison of these predictions with observations suggests that although the changes in boundary layer mass flux induced by deforestation are small compared to the observations, their spatial derivatives are not negligible in comparison to those of observations. Since vertical velocity is a linear function of these derivatives, a comparison of the predictions of vertical velocity with observations indicates that the changes in vertical velocity following deforestation are significant. However, orography and warm ocean temperature are important forcings for the large-scale circulations in the Amazon and Indonesia, respectively. Unlike these two regions, the circulation over the Congo

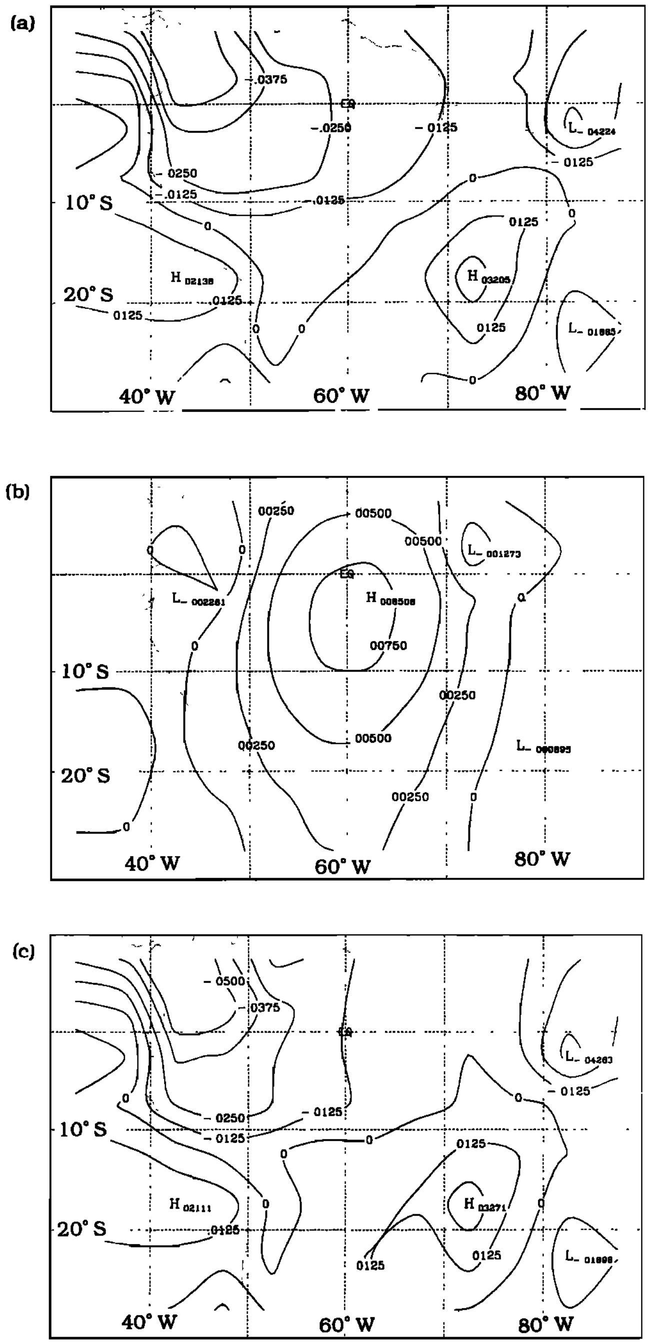

Figure 11. Vertical velocity over the boundary layer for the Amazon region, in kg/m/s^3 (= 0.01 mbar/s): (a) observations, (b) predictions of the change due to deforestation, and (c) predictions of post-deforestation velocity (a plus b).

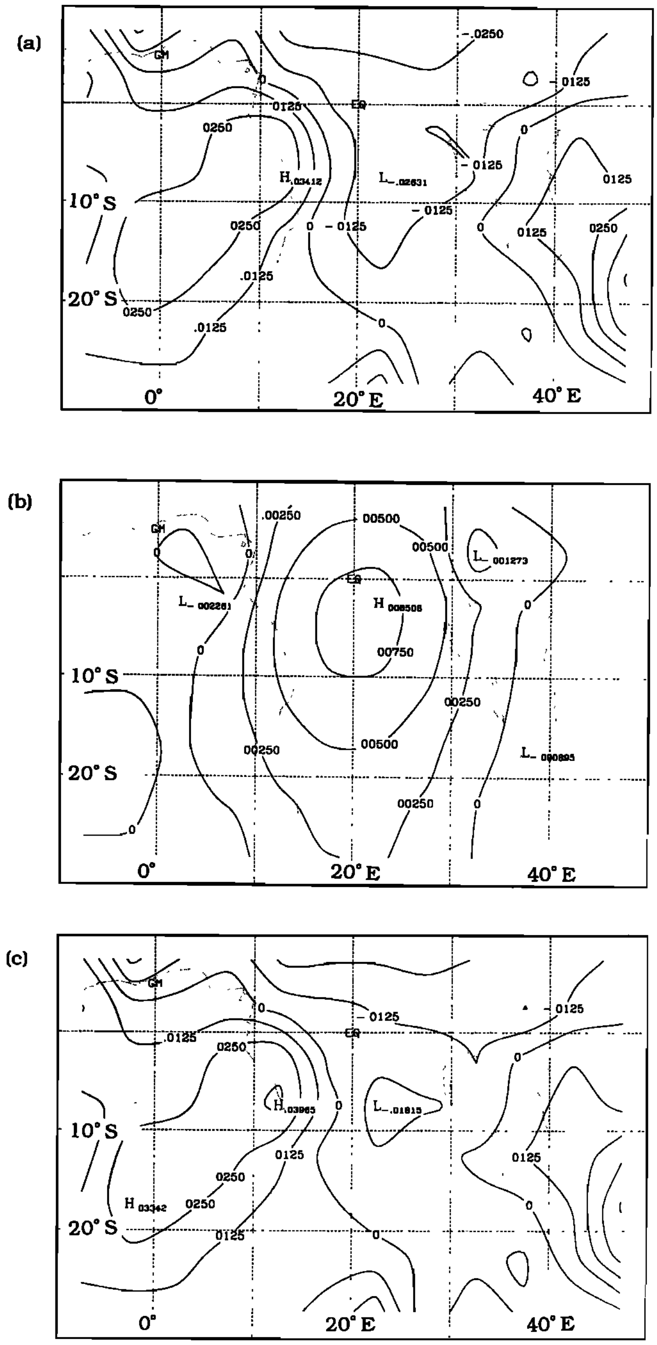

Figure 12. Same as Figure 10 but for Congo.

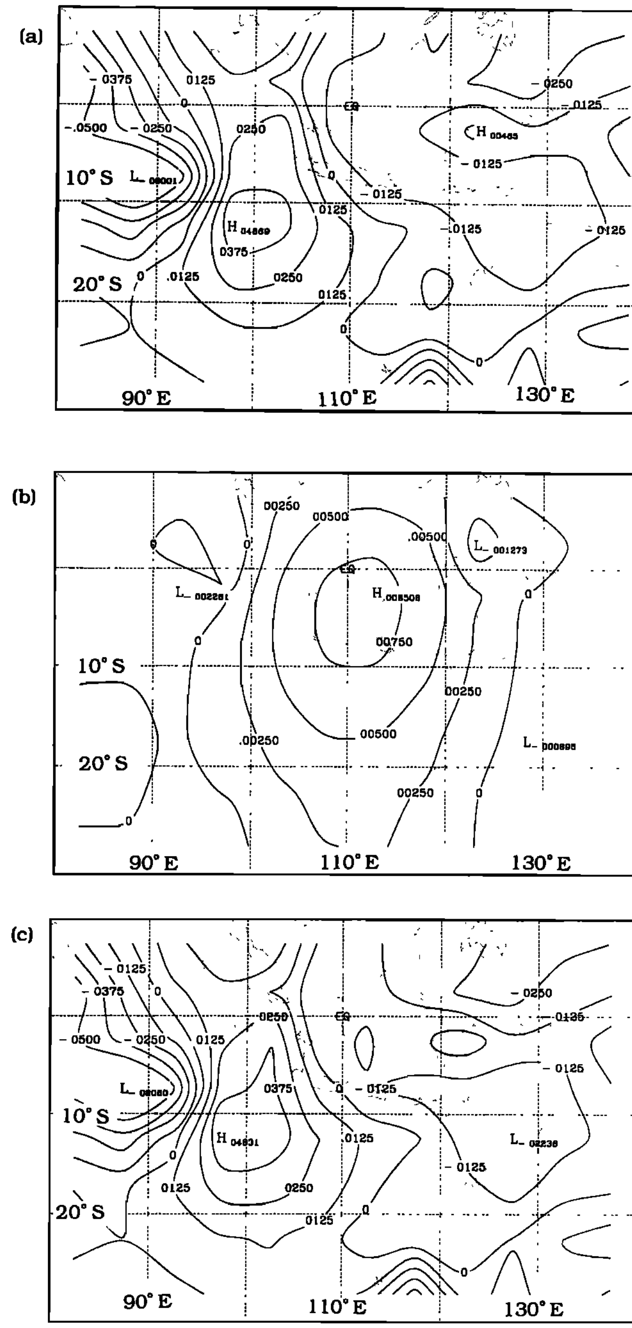

Figure 13. Same as Figure 10 but for Indonesia.

region is largely dependent on the role of vegetation and rain forests. Hence the circulation over Congo may be more vulnerable to the impact of anthropogenic changes in vegetation cover.

3. The potential changes in large-scale atmospheric circulations following deforestation in the tropics are significant enough to suggest an important role for vegetation in sustaining the total observed circulations. Hence the main conclusion of this paper states that rain forests enhance large-scale atmospheric circulations. However, the same circulations are responsible for maintaining the environmental conditions (rainfall, temperature, and humidity) that sustain vegetation and rain forests. Hence this conclusion implies a complex and nonlinear relation between vegetation and climate. This climate-vegetation interaction will be a subject for our future research. (The term "sustain" is used in this paper to describe a significant contribution to the forcing and maintaining of atmospheric circulations; it is not meant to indicate any exclusive role in the forcing of these circulations.)

4. The concept that vegetation and rain forests help sustain large-scale atmospheric circulations in the tropics underlines a set of important land-atmosphere-ocean interactions. These interactions are illustrated schematically by Figure 14. Depending on the state of vegetative cover of any region, a strong, moderate, or weak circulation may develop. At the extreme end of possible conditions a desert may not sustain any circu-

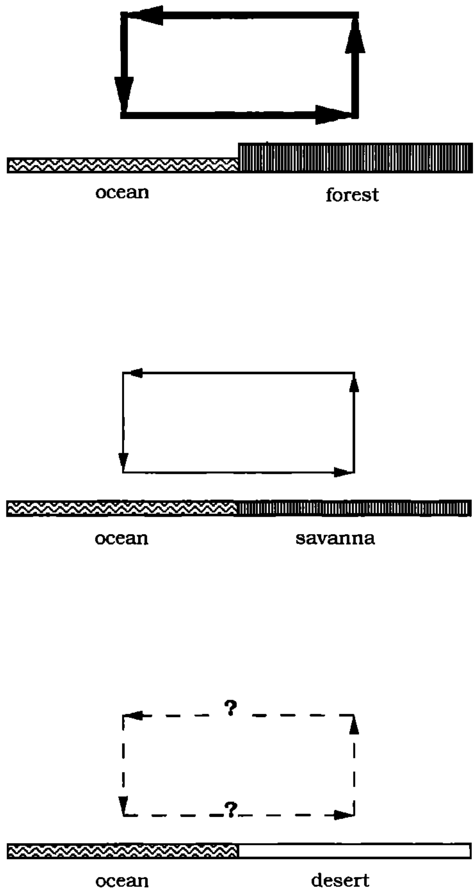

Figure 14. Land-atmosphere-ocean interactions at large scales.

lation with an adjacent ocean. The multiple equilibria that are described schematically in Figure 14 may be relevant to some of the paleoclimatic observations from several tropical regions. For example, *McClure* [1976] notes the occurrence of a period with wet climate in the Arabian peninsula 25,000 years ago. Today, this region is a desert with virtually no interaction with the adjacent ocean. In another tropical region, *Street and Grove* [1979] present data on the status of the water levels in the lakes of intertropical Africa during the period between 25,000 and 20,000 years ago. For several African lakes, water levels were higher than today's levels. The variability in solar forcing, sea surface temperature of the tropical ocean, and the type of vegetation cover are among the factors that may ultimately explain these observations. The relevance of this theory to paleoclimatic observations remains speculative at this point. However, the role of vegetation in the paleoclimate of tropical regions is an interesting subject that deserves further investigation.

Notation

C wave speed, m/s.
C_p specific heat capacity of air at constant pressure, J/kg/°K.
E evaporation, kg/m^2.
f Coriolis parameter, s.
F total flux of heat from the surface, including latent and sensible forms, W/m^2.
g gravitational acceleration, m/s^2.
H sensible heat flux, W/m^2.
N net surface radiation, W/m^2.
N_s net solar radiation, W/m^2.
N_t net terrestrial radiation, W/m^2.
p pressure, kg/m/s^2.
q specific humidity, kg/kg.
R ideal gas constant for dry air, J/kg/°K.
s entropy, J/kg/°K.
s_b boundary layer entropy, J/kg/°K.
S rate of change of temperature with respect to pressure, K m s^2/kg.
T temperature, °K.
T_e temperature at radiative convective equilibrium, °K.
u, v components of mass flux in the boundary layer, kg/m/s.
w vertical velocity above the boundary layer (pressure coordinates), kg/m/s^3.
x, y distances in the zonal and meridional directions, respectively, m.
α specific volume, m^3/kg.
α_n coefficient of Newtonian cooling, s.
Δp half the depth of the troposphere corresponding to the region where radiative cooling takes place, kg/m/s^2.
ε coefficient of Rayleigh friction, s.
Φ mass-weighted integral of the geopotential height in the boundary layer, kg/s^2.
γ parameter defined by equation (15).
λ latent heat of vaporization, J/kg.

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E. A. B. Eltahir, Parsons Laboratory, MIT, Room 48-207, Cambridge, MA 02139.

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