Understanding Climate Sensitivity to Tropical Deforestation in a Mechanistic Model

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(Manuscript received 4 June 1997, in final form 7 October 1997)

ABSTRACT

An analysis is provided to explain the sensitivity of regional climate to tropical deforestation based on an intermediate-level mechanistic model for land–atmosphere interaction. The analytical framework is made possible by the fact that feedback to local thermodynamics from large-scale dynamics is negligible, so the determining processes take place in situ. The analytical method accurately reproduces the intermediate-level numerical model results for an albedo change scenario and further provides insight into the mechanisms. A three-way balance among large-scale adiabatic cooling, moist convective heating, and radiative heating allows two positive feedback mechanisms, moisture convergence feedback and evaporation feedback, that give rise to the high sensitivity. The analysis also highlights a deficiency in column energy balance commonly used in tropical simple models, which results in a sensitivity that is likely too high. In light of these findings, some immediate needs for further advancing understanding of the problem are discussed.

1. Introduction

Atmospheric general circulation model (GCM) simulations have suggested a possible change in the regional climate as the result of tropical deforestation, notably in the Amazon Basin [see Dickinson (1995) for a recent review]. Under a hypothesized Amazon basin-wide deforestation scenario, almost all GCM simulations found a significant reduction in precipitation and evapotranspiration [typically ranging from −10% to −30%, see Hahmann and Dickinson (1997, their Table 1) for a summary], and most found a decrease in moisture convergence with the exceptions of, for example, Lean et al. (1996), indicating a positive atmospheric feedback. However, the picture becomes more complicated when one examines the sensitivities to changes in individual land surface properties, such as albedo and roughness length (e.g., Dirmeyer and Shukla 1994; Sud et al. 1996b; Hahmann and Dickinson 1997). Different experiments emphasize the importance of different aspects of the land–atmosphere system and sometimes draw opposite conclusions. Many more issues arise, such as the roles of control climate and moist convection (Polcher and Laval 1994; Polcher 1995), the effects of a warmer ground (Eltahir and Bras 1993), and the effects of a more realistic surface representation (Lean et al. 1996).

Although progress has been made in understanding the individual model behavior (e.g., Nobre et al. 1991; Sud et al. 1993; Zhang et al. 1996), the complexity of GCMs hinders our ability to fully depict the cause–response relationship and to understand the disparities among the GCM experiments. Much attention has been paid to detail land surface representation, leaving the host GCMs to take care of the atmospheric modeling. The key processes determining the deforestation response both in the atmosphere and over the land are not clearly identified nor well understood. For instance, tropical rainfall is mainly convective and the associated clouds play an important role in the energy budget. Current GCMs use a wide range of moist convective and cloud-radiation parameterization schemes. How much of the difference among the deforestation experiments can be attributed to this is not clear. Land surface processes are equally complicated, and to determine which are the important ones and how they interact with boundary layer and moist convection remains a challenge. It is difficult to cross examine simulations with different land surface schemes and host GCMs, but a theoretical understanding can help delineate the key processes and mechanisms.

Eltahir and Bras (1993) developed a simple model to interpret some early GCM results, highlighting the competing feedback effects of a warmer surface and less precipitation, both of which can result from a reduction in evaporation. Attempting an understanding of the interaction between land surface and large-scale atmospheric dynamics and thermodynamics, Zeng et al. (1996, hereafter referred to as ZDZ) developed an intermediate-level mechanistic atmosphere–land–ocean model. They found high sensitivities of the regional

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climate to deforestation as a result of the existence of a positive feedback loop in the tropical climate system. Their analysis of the thermodynamic equation reveals that the deforestation response is largely determined by a three-way balance among large-scale adiabatic cooling, convective heating, and radiation.

By closely examining these crucial interactions, we are now able to provide a theoretical framework. A key assumption that the temperature change can be neglected enables us to obtain analytical results for the local thermodynamics without worrying about the remote large-scale dynamics. These simple analytical results provide insights into the tropical land–atmosphere interaction problem. However, given the complexity of the problem, we do not attempt in this paper to give an accurate depiction of the whole problem, nor to answer all the questions raised by GCM studies. Rather the author restricts the study to the parameterizations used in ZDZ so that comparisons can be made to the numerical model. We will look at climate sensitivity to albedo change, focusing on the moisture and evaporation feedbacks. This approach proves quite useful by stripping off some complicated but less essential dynamic processes and singling out the important processes and key assumptions associated with them. Equally important, the analysis critically exposes possible deficiencies in the parameterizations and assumptions made and thereby provides guidance for future development. This type of ''deforestation minus control'' study, as typically done in GCM studies, compares the current climate system (control) with a hypothetical climate system with a different boundary condition (deforestation) that is in its own equilibrium climate. So a perturbation theory as defined here studies the difference between the two systems.

We will present the theory and its application to the deforestation problem in the next three sections. In the final section we will discuss the implications to GCM studies and the crucial issue of energy closure.

2. Thermodynamics and moisture closure

We start with the thermodynamic equation

$$C_p\frac{\partial T}{\partial t} + C_p\left(\frac{\partial T}{\partial z} + \frac{g}{C_p}\right)w = Q_{\mathrm{R}} + Q_C, \qquad (1)$$

where horizontal advection of T has been neglected since the advection of temperature is small compared to the wind convergence because the temperature field is much smoother than the wind field. To quantify this, the temperature divergence is

$$\nabla\cdot(T\mathbf{v}) = T\nabla\cdot\mathbf{v} + \mathbf{v}\cdot\nabla T.$$

In the Amazon the temperature gradient across a distance of about 2000 km is typically 3 K (e.g., European Centre for Medium-Range Weather Forecasts monthly climatology; see section 3 for further discussion on the temperature homogenization process), and the wind gradient is about 10 m s^{-1}. Assuming a mean temperature of 300 K and mean wind of 10 m s^{-1}, this gives a temperature advection of 1.5×10^{-5} K s^{-1} and the term due to wind convergence is 1.5×10^{-3} K s^{-1} (the second and the first term, respectively), so that the horizontal advection is negligible. This is in line with Mintz's (1984) argument on the advection of moisture for synoptic-scale motion. This type of scaling analysis holds better on the seasonal or longer timescale of our concern. On these timescales, the convective parameterization averages out higher-frequency variability, leading to a quasi-steady state where the time derivative in (1) vanishes. The second term on the left-hand side is the adiabatic cooling due to large-scale upward motion that is balanced by diabatic heating, namely, radiative heating and convective latent heating. Sensible heat and diffusion are neglected. An atmospheric dry stability is defined as $s \equiv \partial T/\partial z + g/C_p = HN^2R^{-1}$ with the buoyancy frequency N, the atmospheric scale height H, and the gas constant R. Then Eq. (1) becomes Eq. (1e) of ZDZ.

We parameterize the radiative heating Q_{R} by Newtonian cooling:

$$Q_{\mathrm{R}} = C_p(T^* - T)/\tau_{\mathrm{R}},$$

where T^* is an atmospheric equilibrium temperature, T is the actual atmospheric temperature, and τ_{R} is the Newtonian relaxation time. Assuming the net moisture converging into a vertical column is precipitated out, the convective heating is

$$Q_{\mathrm{C}} = \eta C_p P,$$

where η vertically distributes the latent heat released by precipitation according to a typical profile of tropical convective heating. The precipitation P is determined by a simple moisture closure,

$$P = E + C, \qquad (2)$$

where E is the evaporation and C is the moisture convergence. At steady state, Eq. (1) is now

$$sw = Q_{\mathrm{R}}/C_p + \eta P,$$

where w is the large-scale vertical velocity. Only moisture convergence from the planetary boundary layer is allowed to contribute to precipitation. This leads to model behavior dictated by the thermodynamic balance at the top of the boundary layer (800 mb here).

It is of interest to compare the ZDZ model with the simpler Gill (1980) model and Lindzen and Nigam (1987) model. The Gill model uses a single baroclinic mode in the vertical assuming moist convective heating excites only that mode. The Lindzen–Nigam model emphasizes the importance of boundary layer processes. In this respect the ZDZ model is a hybrid of the two: it uses energy balance in the boundary layer but some dynamics of the free atmosphere baroclinic mode. In addition, the three-layer ZDZ model has more freedom to allow the effect of the barotropic mode.

Combining Eqs. (1) and (2), the thermodynamic equation at 800 mb is now

$$\frac{s}{\eta}w = \tilde{Q}_R + E + C, \quad (3)$$

where $\tilde{Q}_R \equiv Q_R(\eta C_p)^{-1}$, and w, T, T^*, η, and C denote their values at or below 800 mb for simplicity. Since the moisture distribution is much smoother than the corresponding wind pattern, moisture convergence is dominated by wind convergence (i.e., neglecting moisture advection):

$$C \approx \rho q w,$$

where ρ and q are air density and humidity in the boundary layer, respectively.

The fact that the adiabatic cooling term on the left-hand side of Eq. (3) is directly proportional to vertical velocity w, as is moisture convergence C, leads to the moisture convergence feedback discussed in the next section. Expressing w in terms of C,

$$\frac{s}{\eta}w \equiv \rho q_0 w$$

$$= \frac{q_0}{q}C,$$

where $q_0 \equiv s(\eta\rho)^{-1}$. Defining $\tilde{q} \equiv q/q_0$, Eq. (3) now becomes

$$\frac{C}{\tilde{q}} = \tilde{Q}_R + E + C.$$

Rearranging the equation yields

$$\frac{1 - \tilde{q}}{\tilde{q}}C = \tilde{Q}_R + E. \quad (4)$$

This is similar to the moisture convergence feedback used in Webster (1981), Zebiak (1986), and others. It is also similar to the gross moist stability of Neelin and Held (1987), with our $\tilde{q}$ equivalent to their moisture stratification, and $1 - \tilde{q}$ equivalent to their gross moist stability term ΔM, scaled by the dry stability. Equation (4) is the moist static energy equation as the sum of Eq. (1) and Eq. (2). It is essentially Eq. (7a) of ZDZ.

3. Application to land surface albedo change

Up to this point we have derived an analytical relation linking the energetics and moisture terms. We now proceed to the application over land in an attempt to explain the results found in ZDZ for a continuous deforestation scenario.

Over land, the surface wetness is assumed proportional to precipitation (see ZDZ for more detail):

$$\beta = P/P_0,$$

where P_0 is a constant with a value of 7 mm day^{-1}; above that the soil is saturated. This lumps all the hydrology into one linear relationship. It applies to a steady state as an average representation over a season and reflects the simple fact that the more rain, the wetter. As crude as it looks, this extremely simple parameterization catches the zeroth-order dependence. We take it as a starting point when better simple parameterization is not available to climatological application (see section 5 for further discussion). A further assumption is that under the deforestation scenario, the change in evaporation due to wetness dominates over other effects such as wind and humidity changes so the potential evaporation remains constant. Then a simple relation between evaporation and precipitation can be extracted:

$$E = \beta E_p = \tilde{\beta}P, \quad (5)$$

where E_p is the potential evaporation and $\tilde{\beta} \equiv E_p/P_0$ is a constant parameter under the above assumptions.

Eliminating C and E in Eqs. (2), (4), and (5), we have

$$P = \frac{\tilde{q}}{1 - \tilde{q} - \tilde{\beta}}\tilde{Q}_R,$$

$$= \frac{\tilde{q}}{1 - \tilde{q} - \tilde{\beta}}\tilde{\epsilon}(T^* - T), \quad (6)$$

where $\tilde{\epsilon} \equiv (\tau_R \eta)^{-1}$. This equation ties the hydrology closely with radiative heating and is a powerful relation, as will be demonstrated.

Under a deforestation scenario where surface albedo is increased by a small amount, the change in precipitation is

$$\Delta P = \frac{\tilde{q}}{1 - \tilde{q} - \tilde{\beta}}\tilde{\epsilon}(\Delta T^* - \Delta T), \quad (7)$$

with albedo effect absorbed in ΔT^*. This implicitly assumes other changes in the coefficients are negligible, and the approximations made in the climatology can be carried along. This assumption was made by ZDZ in their experiment A5 (Amazon Basin albedo increased by 0.05). The changes in moisture convergence and evaporation are most easily expressed in terms of ΔP:

$$\Delta C = (1 - \tilde{\beta})\Delta P \quad \text{and}$$

$$\Delta E = \tilde{\beta}\Delta P.$$

A high sensitivity of precipitation to change in radiative forcing is seen in Eq. (7) due to the smallness of the denominator. This is controlled by two mechanisms: moisture convergence feedback and evaporation feedback, as illustrated in Fig. 1. In the first mechanism, an increase in precipitation releases latent heat that drives large-scale upward motion [Eq. (3)], which causes more moisture convergence, leading to more precipitation. In the evaporation feedback, higher precipitation leads to a wetter surface and more evaporation, which in turn contributes to even more precipitation through the moisture closure (2). The overlap of the two positive

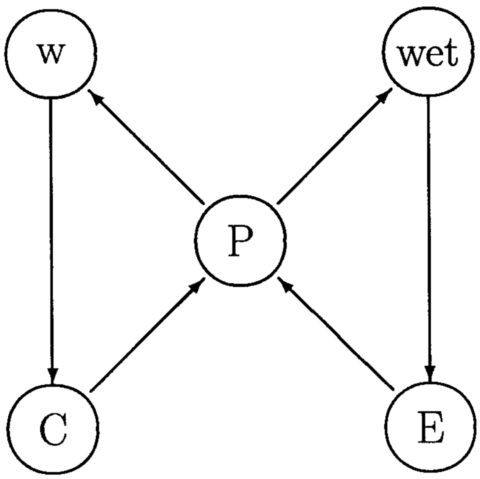

FIG. 1. The two major feedback loops in the perturbed region: moisture convergence feedback and evaporation feedback. Here, P is precipitation, w is large-scale upward motion, C is large-scale moisture convergence, wet is surface wetness, and E is evapotranspiration.

feedback loops at precipitation gives a higher sensitivity than a simple combination of the two does. For instance, a perturbation in moisture convergence is propagated into a change in evaporation through the precipitation connection and vice versa. However, the evaporation feedback maybe partially compensated for by a change in radiation through an energy balance requirement not accounted for here, as will be discussed in section 5.

In addition to the above two feedback mechanisms, another key issue is how the radiation feeds back—that is, what the actual temperature change ΔT is like in our formulation. Here, ΔT is a result of large-scale dynamics as well as local perturbation and is difficult to quantify without solving the full dynamic equations. It turns out ΔT is negligible compared to ΔT^*. It is difficult to prove this in a rigorous mathematical manner, but Gill (1980) gives a well-known analytical example where a concentrated heat source near the equator is propagated out and damped on the way. The basic reason is that localized thermodynamic perturbation is spread out by a Helmholtz-like operator so that the temperature is homogenized within a radius of deformation. This dynamic effect on temperature homogenization is well known among tropical dynamicists (e.g., Held and Hou 1980). If the perturbation is localized relative to the action range of dynamics as measured by the radius of deformation, then ΔT will be small. Based on this criterion, we propose a scaling argument:

$$\frac{\Delta T}{\Delta T^*} = \frac{A_r}{\frac{4}{3}\lambda_x(2\lambda_y)}, \qquad = \frac{3A_r}{8\lambda_x\lambda_y}, \tag{8}$$

where A_r is the area of the perturbed region, and λ_x and λ_y are the radii of deformation along zonal and meridional direction, respectively. There is a factor 2 in front of λ_y because of the extension in both north and south directions. The 4/3 factor before λ_x comes from the fact that the westward propagating Rossby wave has a speed one-third of that of the equatorial Kelvin wave, so the information goes only one-third as far to the west in a steady state with the same damping.

Now a quick numerical estimate is possible. The moist Kelvin wave speed squared is

$$c^2 = (1 - \tilde{q})C_p H N^2 Z_B / R,$$

where Z_B is the boundary layer depth. Given the numerical values (see next section), we have $c \approx 60$ m s^{-1}. Then the equatorial radii of deformation are

$$\lambda_y = \left(\frac{2c}{\beta}\right)^{1/2} \approx 2000 \text{ km} \quad \text{and}$$

$$\lambda_x = \frac{c}{\alpha} \approx 18\,000 \text{ km},$$

where β is the variation of the Coriolis parameter with latitude at equator, and the zonal damping coefficient α is estimated as

$$\alpha = (\alpha_1 \alpha_2)^{1/2} \approx (3 \text{ day})^{-1},$$

with equivalent momentum damping in the boundary layer $\alpha_1 = (1 \text{ day})^{-1}$ (see ZDZ, their Fig. 5) and Newtonian cooling $\alpha_2 = (9 \text{ day})^{-1}$ (a weighted average over land and ocean). The Amazon region is about 2000 km $\times$ 2000 km. Then using Eq. (8), we get

$$\frac{\Delta T}{\Delta T^*} = \frac{3 \times (2000 \text{ km})^2}{8 \times 18\,000 \text{ km} \times 2000 \text{ km}}, \qquad = 0.04,$$

where ΔT is only 4% of ΔT^*. We note that since the zonal radius of deformation is much larger than in the meridional direction, the perturbation is mostly spread out along the zonal direction.

We have also run the ZDZ model, which shows ΔT is about 3% of ΔT^*, which is in support of our scaling argument. The fact that ΔT can be neglected in Eq. (7) greatly simplifies the matter by limiting all the feedbacks to a local region, as long as this region is small compared to the radius of deformation and large enough so the moist convective parameterization holds and the mesoscale effect can be neglected.

The equilibrium temperature T^* is directly related to solar radiation and surface albedo by assuming a dry adiabatic lapse rate. A surface temperature, T^*_s, that serves a similar role as sea surface temperature (SST) over ocean is computed using absorbed solar radiation at ground and the Stephan–Boltzmann law. Combining Eqs. (3b)–(3d) of ZDZ,

$$\epsilon\sigma(T^* + \Gamma_d z)^4 = (1 - A)S,$$

where σ is the Stephan–Boltzmann constant, Γ_d is the dry adiabatic lapse rate, z is height, A is the surface albedo, and S is the downward solar radiation at surface. The cloud effects are neglected. Given a small perturbation in albedo ΔA, one has

$$\frac{4\Delta T^*}{T^* + \Gamma_d z} \approx -\frac{\Delta A}{1 - A},$$

which can be further approximated by neglecting $\Gamma_d z$ and A in the denominators as

$$\Delta T^* \approx -\frac{T^*}{4}\Delta A. \qquad (9)$$

4. Results and comparison with the numerical model

The parameter values are based on ZDZ with $N^2 = 10^{-4}\ \mathrm{s}^{-1}$. The vertical profile of latent heating gives $\eta = 0.0625\ \mathrm{K\ mm}^{-1}$, $\tau_R = 26$ days. This gives $q_0 = s(\eta\rho)^{-1} = 44\ \mathrm{g\ kg}^{-1}$; $\tilde{\epsilon} = (\tau_R \eta)^{-1} = 0.615\ \mathrm{mm\ day}^{-1}\ \mathrm{K}^{-1}$. We then have

$$\tilde{q} = \frac{q}{q_0} = 0.36 \quad \text{and}$$

$$\tilde{\beta} = 0.5.$$

In the simulated January climatology of ZDZ, averaged over Amazon, $T^* - T = 5.5$ K. Using Eq. (6), we get $P = 8.8\ \mathrm{mm\ day}^{-1}$, almost identical to model-simulated $8.9\ \mathrm{mm\ day}^{-1}$ (ZDZ, their Table 5), $E = 4.4\ \mathrm{mm\ day}^{-1}$, and $C = 4.4\ \mathrm{mm\ day}^{-1}$.

In the case of surface albedo increase by 0.05 (ZDZ's experiment A5), neglecting ΔT and using (9), Eq. (7) becomes

$$\Delta P = \frac{\tilde{q}}{1 - \tilde{\beta} - \tilde{q}}\tilde{\epsilon}\Delta T^*. \qquad (10)$$

With $\Delta T^* = -(\Delta A/4)(T^*) = 3.75$ K given $\Delta A = 0.05$ and $T^* \approx 300$ K. This leads to $\Delta P = -6.0\ \mathrm{mm\ day}^{-1}$. This is somewhat larger than ZDZ's result of -5.0 mm day^{-1}, largely due to the smoothing used in the numerical model. In fact, if one takes the smoothed $\Delta T^* \approx 3$ K (Zeng 1994, Fig. 4.1j) one gets $\Delta P = -4.8$ mm day^{-1}, which is a close agreement.

We have rerun ZDZ's model without smoothing over the Amazon deforestation region. Figure 2 shows the decrease in rainfall as surface albedo increases for theory and the ZDZ model. The theory of Eq. (10) predicts a trend of precipitation decrease very close to the model initially. The model precipitation then drops slightly faster and levels off as albedo increases further due to the lower cutoff for surface wetness. This is also approaching the regime where the linear theory does not apply. Although the model control climate without smoothing is probably too wet, the theoretical trend does not depend on it. We note that this good agreement does not nessesarily imply the theory is close to the real world or GCM simulations. Rather, this demonstrates the capability of the analytical approach to pinpoint the major processes and mechanisms in the intermediate-level model because essentially the same parameterizations are used.

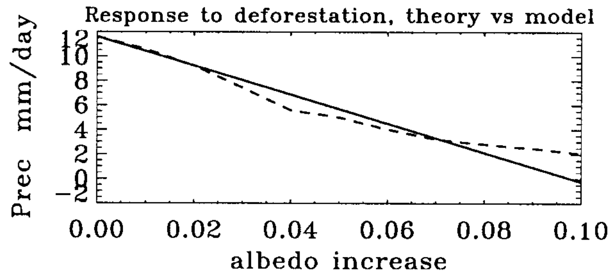

FIG. 2. Perpetual January precipitation decreases as surface albedo increases. The solid line is predicted by the analytical theory with the control climate (where albedo increase is zero) matched to that of the numerical model; the dashed line is the result from the numerical model of ZDZ.

5. Discussion

We have provided an analytical framework to diagnose the sensitivity of regional climate to tropical land surface change as found in the intermediate-level mechanistic model of ZDZ. The theory gives a clear picture of the important processes and mechanisms by stripping off some less essential dynamic processes.

The analysis reveals that the three-way balance of adiabatic cooling, convective heating, and radiative heating is supplemented by a direct proportionality between large-scale upward motion and moisture convergence. Surface hydrology further links evaporation with precipitation and moisture convergence. In this framework the high sensitivity to surface albedo change is dominated by two processes: first, the moisture convergence feedback, where more precipitation drives stronger large-scale upward motion and more moisture convergence; second, the evaporation feedback, where more rainfall makes a wetter ground and more evaporation. Both processes positively feed back into precipitation through the moisture closure. Precipitation acts as the bridge linking the two feedback loops, giving higher sensitivity than a simple combination of the two does. When it comes to quantitative comparison the theory predicts a somewhat higher sensitivity than most GCM experiments do. A closer comparison is difficult because the theory does not resolve the seasonal cycle of surface hydrology, and it has been applied only to the case of albedo change. All these factors should be considered in the future.

It is worthwhile to point out that these two feedbacks are local. The large-scale dynamics come in only through the large-scale upward motion term (with advection neglected). At first thought this would appear somewhat surprising because the large-scale dynamics would also change the temperature, wind, etc. Evaporation would be modified by these changes, while difference in temperature can cause radiative heating to change. In fact, the latter is a negative feedback because more diabatic heating would cause the temperature to rise, hence less radiative heating. It turns out that it is precisely this dynamics that spreads out this influence, so the temperature is homogenized within a radius of deformation. This fact is of key importance to our analytical framework, so there should be less concern about the full large-scale dynamics. From this point of view the large-scale Hadley–Walker circulations response to Amazon deforestation, as depicted in ZDZ's Fig. 2, is indeed a mere ''response'' to local dynamics and physics to the first-order approximation.

This also sheds light on the important question of to what spatial extent the deforestation has to go in order to have a significant impact on the local climate. Our theory holds as long as the perturbed region is significantly smaller than the radius of deformation. On the other hand, the region needs to be large enough so that the ensemble effects of moist convection can establish its quasi equilibrium with large-scale dynamics, that mesoscale effects do not significantly alter the scene, and, needless to say, that advection can be neglected. Based on these considerations, we propose that similar sensitivity can be found for a deforestation area smaller than the whole Amazon Basin. How small the spatial extent needs to be depends greatly on the role of subbasin-scale processes. In a GCM simulation where the Amazon is deforested only to its 1988 level, Sud et al. (1996a) found a significant evaporation decrease but a smaller reduction in precipitation due to an increase in moisture convergence. This is possibly associated with the subbasin-scale circulation induced by a warmer surface. Interestingly, in a similar simulation but only looking at a few individual rainfall events, Walker et al. (1995) found a decrease in moisture convergence as well as decreases in evaporation and precipitation. These intriguing results indicate more complexities at finer spatial and temporal scales.

We note that the analytical theory is a perturbation theory. The theory does not really require the control climate state variables such as precipitation, except for knowing it is a deep convective region in order for the convective parameterization to apply. From this point of view, the simpler approach of comparing the changes directly among GCMs is better than ZDZ's approach, where everything is scaled to have the same control climate precipitation (see their Fig. 15). However, the control climate is important in providing the trend parameters. In our theory the most important trend parameters are the moist stability $(1 - \tilde{q})$ and the surface hydrology factor $(\tilde{\beta})$, representing the moisture convergence feedback and evaporation feedback, respectively. The mutually enhancing feedback loops cause high sensitivity to these basic-state parameters. In a GCM these trend parameters are inseparable from the control climate variables, and they reflect the characteristics of simulated moist convection and surface hydrology. This is probably why deforestation results are found to be sensitive to the control climate in many GCMs (e.g., Polcher and Laval 1994).

We now turn to some deficiencies in parameterizations and assumptions as highlighted by the analysis. As pointed out in section 3, the energy balance is not enforced in a vertical column. The radiative parameterization calculates an equilibrium surface temperature T_s^*. Although absorbed solar radiation including albedo is accounted for, T_s^* is subsequently held fixed, as is SST. This ignores a fundamental difference between ocean and land surface: the ocean can be a substantial energy sink or source, whereas the land surface cannot be a sink or source over a period longer than a day because of its low heat capacity and lack of transportation. For instance, an increase in evaporation would consume more solar energy absorbed, which would decrease sensible and radiative heating by lowering surface temperature, unless additional solar heating is supplied. In the SST-like treatment T_s^* is held constant, requiring heat supply from below. In the ocean this is done by ocean dynamics through lateral and vertical transport. Over land there is simply no such mechanism. So the evaporation feedback should be considered together with radiative heating feedback in a consistent manner. This most likely leads to a lower sensitivity. It is of interest to point out that the land surface is parameterized in similarly energetically inconsistent ways by the most existing simple to intermediate-level tropical atmospheric models. Our analysis here clearly demonstrates the danger of this popular practice.

The hydrological aspect of the simple parameterization for evaporation of Eq. (5) also needs refinement. Field experiments in the Amazon (Shuttleworth 1988; Gash et al. 1996) have shown that the evaporation is more or less constant throughout a year, with dry season transpiration from uptaking of deep soil moisture. This is in apparent contrast with Eq. (5), where evaporation closely follows precipitation that has a strong seasonal cycle. However, this simple comparison can be somewhat misleading because Eq. (5) applies to a perpetual climatological change. In any case (5) cannot account for some important effects. Given the current climate that soil moisture is saturated through much of a year, dry season evaporation would be the first to suffer from less soil moisture. In addition to soil moisture availability, the mere fact that grass has much shorter roots than rainforest vegetation would further reduce evaporation. Our belief is that these processes should and can be modeled by a better but still simple land surface scheme.

ZDZ argued that cloud effects can be neglected because of a near cancellation of cloud longwave and shortwave forcing at the top of the atmosphere, as far as the total column energy is concerned. However, as they also cautioned, the surface energy balance would be significantly different because the shortwave radiation is mainly absorbed at the surface, while the longwave radiation is trapped more in the atmosphere. This difference in surface energy budget can have significant impact by changing the evaporation and sensible heat fluxes (Eltahir 1996) and by changing the partitioning of precipitation into evaporation and runoff when coupled with surface hydrology.

Another unsolved issue is the seemingly negative feedback effect of a higher surface temperature after deforestation (e.g., Eltahir and Bras 1993). This warming is at the expense of less evaporation and it is not clear what its role is in the entire energy budget, though an immediate effect is to warm the boundary layer through sensible heat flux and longwave heating. In addition, the energy closure may have a strong constraint on the surface temperature at first. A further complication is surface roughness change due to its uneven contribution to sensible heat and evaporation, to which the theory has not been applied.

We have demonstrated that theoretical frameworks as exemplified here are possible and useful in providing insights into the complicated processes and mechanisms associated with the climate change induced by tropical land surface disturbance. Our analysis helps to clarify the deficiency of the popular but energetically inconsistent treatment of the land surface in simple models. This lays the foundation for further progress along this direction. As we argue, remote large-scale dynamic response being less important, the crucial issue is a comprehensive understanding of column energy balance from the top of the atmosphere to the ground, including radiative, evaporative, sensible heat fluxes, and surface hydrology.

Acknowledgments. The author is indebted to Xubin Zeng and David Neelin for their constructive comments. This work would not be possible without the continuing encouragement and support from Robert E. Dickinson. We thank Paul Dirmeyer, another anonymous reviewer, and editor Inez Fung for their comments that significantly improved the manuscript. Robin O'Brian is thanked for her careful editing of the manuscript. This work was partially supported by NSF Grant ATM-9521389 and NOAA Grant NA46GP0244 (PI: David Neelin).

REFERENCES

Dickinson, R. E., 1995: Land–atmosphere interaction. *Rev. Geophys.*, **33** (Suppl.), 917–922.

Dirmeyer, P. A., and J. Shukla, 1994: Albedo as a modulator of climate response to tropical deforestation. *J. Geophys. Res.,* **99,** 20 863–20 877.

Eltahir, E. A. B., 1996: Role of vegetation in sustaining large-scale atmospheric circulations in the Tropics. *J. Geophys. Res.,* **101,** 4255–4268.

——, and R. L. Bras, 1993: On the response of the tropical atmosphere to large-scale deforestation. *Quart. J. Roy. Meteor. Soc.,* **119,** 779–793.

Gash, J. H. C., C. A. Nobre, J. M. Roberts, and R. L. Victoria, Eds., 1996: *Amazonian Deforestation and Climate.* John Wiley, 611 pp.

Gill, A. E., 1980: Some simple solutions for heat induced tropical circulation. *Quart. J. Roy. Meteor. Soc.,* **106,** 447–462.

Hahmann, A. N., and R. E. Dickinson, 1997: RCCM2–BATS model over tropical South America: Applications to tropical deforestation. *J. Climate,* **10,** 1944–1964.

Held, I. M., and A. Y. Hou, 1980: Nonlinear axially symmetric circulations in a nearly inviscid atmosphere. *J. Atmos. Sci.,* **37,** 515–533.

Lean, J., C. B. Bunton, C. A. Nobre, and P. R. Rowntree, 1996: The simulated impact of Amazonian deforestation on climate using measured ABRACOS vegetation characteristics. *Amazonian Deforestation and Climate,* J. H. C. Gash, C. A. Nobre, J. M. Roberts, and R. L. Victoria, Eds., John Wiley, 549–576.

Lindzen, R. S., and S. Nigam, 1987: On the role of sea surface temperature gradients in forcing low-level winds and convergence in the Tropics. *J. Atmos. Sci.,* **44,** 2418–2436.

Mintz, Y., 1984: The sensitivity of numerically simulated climates to land-surface boundary conditions. *The Global Climate,* J. T. Houghton, Ed., Cambridge University Press, 79–105.

Neelin, J. D., and I. M. Held, 1987: Modeling tropical convergence based on the moist static energy budget. *Mon. Wea. Rev.,* **115,** 3–12.

Nobre, C., P. J. Sellers, and J. Shukla, 1991: Amazonian deforestation and regional climate change. *J. Climate,* **4,** 957–988.

Polcher, J., 1995: Sensitivity of tropical convection to land surface processes. *J. Atmos. Sci.,* **52,** 3143–3161.

——, and K. Laval, 1994: A statistical study of regional impact of deforestation on climate in the LMD GCM. *Climate Dyn.,* **19,** 205–219.

Shuttleworth, W. J., 1988: Evaporation from Amazonian rain forest. *Proc. Roy. Soc. London,* **233B,** 321–346.

Sud, Y. C., W. C. Chao, and G. K. Walker, 1993: Dependence of rainfall on vegetation: Theoretical considerations, simulation experiments, observations, and inferences from simulated atmospheric soundings. *J. Arid Environ.,* **25,** 5–18.

——, R. Yang, and G. K. Walker, 1996a: Impact of in situ deforestation in Amazonia on the regional climate—General circulation model simulation study. *J. Geophys. Res.,* **101,** 7095–7109.

——, G. K. Walker, J.-H. Kim, G. E. Liston, P. J. Sellers, and W. K.-M. Lau, 1996b: Biogeophysical consequences of a tropical deforestation scenario: A GCM simulation study. *J. Climate,* **9,** 3225–3247.

Walker, G. K., Y. C. Sud, and R. Atlas, 1995: Impact of ongoing Amazonian deforestation on local precipitation: A GCM simulation study. *Bull. Amer. Meteor. Soc.,* **76,** 346–361.

Webster, P. J., 1981: Mechanisms determining the atmospheric response to sea surface temperature anomalies. *J. Atmos. Sci.,* **38,** 554–571.

Zebiak, S. E., 1986: Atmospheric convergence feedback in a simple model for El Niño. *Mon. Wea. Rev.,* **114,** 1263–1271.

Zeng, N., 1994: Climatic impact of Amazon deforestation—A study of underlying mechanisms through simple modeling. Ph.D. thesis, University of Arizona, 114 pp.

——, R. E. Dickinson, and X. Zeng, 1996: Climatic impact of Amazon deforestation—A mechanistic model study. *J. Climate,* **9,** 859–883.

Zhang, H., A. Henderson-Sellers, and K. McGuffie, 1996: Impacts of tropical deforestation. Part I: Process analysis of local climatic change. *J. Climate,* **9,** 1497–1517.