THE SENSITIVITY OF THE TROPICAL CLIMATE TO LAND SURFACE PROCESSES

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1 Introduction

The inter-tropical convergence zone displays different properties over ocean and land-surfaces indicating that surface characteristics play an important role in the evolution of this large scale convergence. Charney has shown in 1975 the existence of feedbacks between the tropical atmosphere and the land-surface processes with a potentially unstable character giving thus another indication that strong interactions link land-surfaces processes and the tropical climate. It has been hypothesized that the destruction of tropical rain-forests may have an impact on climate, but prior to answer this question a better understanding of the role of surface processes for tropical climate is needed.

In the present review, the main properties of the surface processes are recalled prior to the presentation of sensitivity studies to surface processes performed with GCMs. First, simple academic experiments are discussed in order to explore possible responses of the tropical atmosphere to given surface conditions. Results from tropical deforestation experiments are then reviewed to determine where uncertainties remain. Finally, a mechanism which could explain the principal characteristics of the interaction between surface processes and the tropical climate is presented.

2 The balance of atmospheric fluxes at the surface

Continental surfaces are a rapidly evolving lower boundary condition to the atmosphere. Thus it is not sufficient to consider each energy flux individually but their balance is the key to understanding the interactions between these two systems. Furthermore, it determines the evolution of the surface. In this section we will review the fluxes which dominate the surface energy balance and discuss how they interact.

The main source of energy at the surface is the incoming solar radiation (R_S) but only a fraction is absorbed and can thus play a role in the surface energy balance. This fraction is determined by surface albedo (α). In the surface energy balance only the net flux (R_{Sn}) will be retained, given by :

$$R_{Sn} = (1 - \alpha)R_s. \tag{1}$$

The other radiative source of energy is the down-welling long wave radiation (R_T) of which the largest fraction is absorbed at the surface. In this part of the spectrum, the surface has also its largest sink of energy: the emitted long wave flux. Thus the net surface long wave radiation is given by :

$$R_{Tn} = \epsilon(R_T - \sigma T_{sk}^4) \tag{2}$$

where ϵ is the surface emissivity of the surface, σ the Stefan-Boltzman constant and T_{sk} the skin temperature, hereafter also referred to as surface temperature.

The turbulent fluxes at the surface, evaporation and sensible heat flux, are most of the time sinks of energy for the continents. They result from the gradients of temperature and moisture between the surface and the atmosphere and the turbulent mixing in the lowest layers of the atmosphere. Following Fick's law these fluxes are given by :

$$H = c_p \rho C_h U_L (T_L - T_{sk}) \tag{3}$$

$$E = \rho C_h U_L \beta (q_L - q_{sat}(T_{sk})). \tag{4}$$

c_p is the specific heat of the air and ρ its density. C_h is the transfer coefficient for heat and moisture (assumed to be same here for simplicity) which depends on an intrinsic property of the surface, its roughness, and the stability of the atmosphere above. T_L, q_L and U_L are temperature, specific humidity and wind speed at the lowest atmospheric level respectively. One of the problems in the moisture flux equation is that the humidity available at the surface for evaporation is unknown. It can only be approximated by assuming a saturated surface ($q_{sat}(T_{ks})$) and limiting the flux by the aridity coefficient (β). β should tend towards zero for dry soils and be close to one when the surface is wet and evaporation is near its potential value. One of the most important tasks of land-surface schemes is to determine this coefficient according to the moisture available through the vegetation or directly from the bare soil. In the first GCMs, this coefficient was a simple function of soil moisture (Manabe, 1969) but nowadays it is obtained through the combination of transpiration, interception loss and bare soil evaporation for the vegetated surface (Viterbo this volume). As it can be seen in equation 3 and 4 the turbulent fluxes can also be sources of energy when the surface is cooler than the atmosphere (stable state) or when the specific humidity is larger than the specific humidity at the surface.

Just as diffusion transports energy from the surface to the atmosphere it will transfer enthalpy from the surface deeper into the ground. This flux is called ground heat flux (G) and is calculated using an equation similar to Eq. 3. The transfer coefficient used will depend on the type of soil, its moisture content and the presence of snow.

Finally, the surface is also a sink of kinetic energy for the atmosphere. The turbulent mixing within the lower layers of the atmosphere will transport momentum from the atmospheric flow to the ground. This flux is given by :

$$M = \rho C_m U_L U_L. \tag{5}$$

C_m is the transfer coefficient for momentum and like C_h, it will depend on the roughness of the surface and the stability of the atmosphere.

At the surface the balance of the thermodynamic fluxes presented above will determine the evolution of the surface temperature. We may write for an infinitesimal thin layer, representing the surface :

$$C_s \frac{\partial T_{sk}}{\partial t} = R_{Tn} + R_{Sn} + LE + H + G. \tag{6}$$

C_s is the heat capacity per unit area of the surface layer. Its value may become very small as the thickness of the layer tends towards zero. The main characteristic which distinguishes the land surface from the ocean is that the layer interacting with the atmosphere is shallow. Thus the heat capacity is smaller than for ocean and changes in the convergence of thermodynamic fluxes will lead to rapid changes of the surface temperature. One may note that for all fluxes, except net solar radiation, the skin temperature plays a key role, thus a change in surface temperature will directly impact the fluxes and redraw the entire balance of fluxes.

The main difficulty in the analysis of surface processes is that the dependence of the fluxes on surface temperature differs widely. Emitted long wave changes with the fourth power of the skin temperature. In the sensible heat flux and ground heat flux however, the dependence is only linear and for evaporation the temperature determines the saturated humidity. Thus, an increase in surface temperature may lead to different modification of the fluxes depending on the situation. For instance if the surface layer is very stable a temperature increase will enhance long wave radiation and the ground heat flux while in other situations it may lead to increase evaporation and sensible heat flux.

Surface temperature is the key variable of surface processes but also the most difficult parameter to analyze when studying the interactions between the surface and the atmosphere. Help comes from the fact that climate is sensitive to the fluxes from the surface and not to its temperature. Each of these fluxes interacts differently with the atmosphere. Both long wave radiation and sensible heat flux warm the atmosphere above the surface. The former will rely on the presence of absorbing gases (water vapor or CO_2) while the latter requires turbulence. Evaporation, on the other hand, extracts energy from the surface which can be transported by the atmosphere before condensation releases heat. Due to these different effects of the surface fluxes to the atmosphere the partitioning of a change in the surface energy between the fluxes will be essential for climate impact. Depending on the availability of soil moisture, an increase of energy input at the surface may in one case, lead to a larger evaporation, thus moistening the atmosphere, and in another case, lead to a higher sensible heat flux which will heat the lower layers. Both responses will have different impacts on climate. Thus, in the analysis of the sensitivity of climate to surface processes changes, surface fluxes have to be studied. It follows that surface temperature can only be viewed as a result of the modified surface balance.

While discussing the balance of fluxes at the surface, four parameters were noted which characterize the surface : albedo, roughness, ground heat transfer coefficient and the aridity coefficient. When one of these parameters is changed it will not only have an impact on the flux in the formulation of which it is used but also the other components of the surface energy balance. Furthermore, the surface fluxes will also be modified by the induced atmospheric changes. To understand the full sensitivity of surface processes and climate to these parameters it is necessary to use atmospheric general circulation models coupled to land-surface schemes which are able to model the complexity of the surface processes.

In the following section we will discuss the sensitivity of the tropical climate to changes in albedo, evaporation characteristics and surface roughness at the surface. These are the main parameters which are affected when the land-surface changes, either in its natural cycle or through human intervention. As pointed out above, the modification of these surface characteristics will affect the surface fluxes and thus the climate. The sensitivity of the tropical climate to changes in the ground heat transfer coefficient have not yet been studied and will thus not be discussed here.

3 Sensitivity to albedo changes

The study of the sensitivity of climate to albedo changes was pioneered by J. Charney (1975) who formulated the hypothesis that the Sahelian drought could be caused by an increase of albedo due to overgrazing. At that time, he proposed a simple mechanism which relies on the fact that the Sahel is one of the regions where the atmosphere, through long wave radiation, loses energy to space to maintain its radiative equilibrium. This loss is compensated by a descending motion of air which heats the lower layers of the atmosphere. If albedo increases in the Sahel, the surface and lower layers of the atmosphere tend to cool, thus more subsidence is needed to maintain the long wave emissions to space. As the subsidence increases, precipitation in the area is reduced. By continuity, convection in the ascending branch of the Hadley circulation has to increase. Charney (1975) showed this with a simple linear model of the Hadley circulation.

This hypothesis seduces by its simplicity but relies only on the radiation balance of the atmosphere. A few years later, Charney tested his hypothesis with a general circulation model (GCM) (Charney et al., 1977). This experiment confirmed the fact that an albedo reduction in the Sahel reduces precipitation but showed that the driving mechanism is not the radiative equilibrium but the modification of turbulent fluxes at the surface (Figure 1).

Although solar radiation increases, net radiation ($R_n = R_{Tn} + R_{Sn}$) is reduced over the Sahel region because the emitted long wave flux increases. The reduced available energy at the surface leads to lower evaporation values. Thus the impact on atmospheric conditions can not only be attributed to long wave radiation but has to be linked to a change in evaporation. The reduction in the latent heat flux is followed by a even larger reduction of precipitation, weaker convective activity and fewer clouds. The reduction in evaporation slowed down the hydrological cycle by reducing the moisture convergence (In long term averages moisture convergence can be considered to be equal to precipitation minus evaporation.). Sud and Fennessy (1982) complemented this result by showing that the reduced heating of the atmosphere through convection is compensated by an increase in subsidence as postulated by Charney in 1975.

These albedo change experiments where repeated by a number of other GCMs and they confirmed the results of Charney (1975, 1977). Results by Laval and Picon (1986), Mylne and Rowntree (1992) and Xue and Shukla (1993) showed the importance of the hydrological cycle of the surface in amplifying the signal. As precipitation is reduced less soil moisture is available, further reducing evaporation and thus amplifying the effect of the decreased net radiation at the surface.

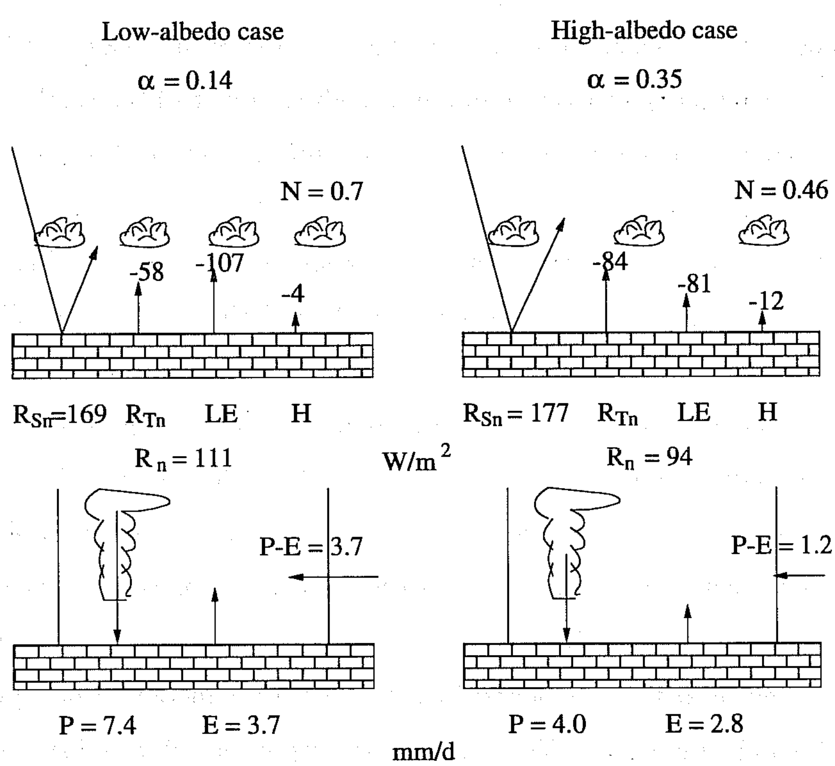

Figure 1: The energy and water budgets over the Sahel region in the experiments of Charney et al. (1977).

Mylne and Rowntree (1982), Dirmeyer and Shukla (1994), and Lean and Rowntree (1997) studied the sensitivity to albedo changes in regions closer to the equator which are under the influence of the inter-tropical convergence zone (ITCZ) for a longer period of the year. These areas, like the sub-tropical region of the Sahel, display a decrease in precipitation when albedo is increased. The driving flux behind this change is again evaporation.

The impact of albedo changes on the tropical climate is probably one of the best studied interactions between the surface and the atmosphere. It appears that an albedo increase will tend to slow down the hydrological cycle and reduce precipitation.

4 Sensitivity to evaporation characteristics

When GCMs are coupled to a simple bucket scheme (Manabe, 1969), it is an easy task to test the sensitivity of the atmosphere to changes in the evaporation characteristics. Modifications can be imposed by choosing values for the aridity coefficient β regardless of soil moisture or other parameters. With current land surface schemes where evaporation is the result of complex processes, such an operation is more difficult as a large number of parameters need to be changed. Despite these difficulties it can still be achieved.

The first experiments of this type were conducted by Shukla and Mintz (1982). In a first integration land-surfaces were supposed to be well irrigated, that is $\beta = 1$ and in a second one land-surfaces were assumed to be totally dry, that is $\beta = 0$. As this was done without altering the surface albedo or other land-surface parameters, the reduction in evaporation has to be compensated by an increase in the sensible heat flux and long wave radiation. In the experiment in which evaporation occurred at the potential rate ($\beta = 1$), surface temperature is reduced and the continental low pressure systems are weakened. On the other hand, when evaporation is suppressed ($\beta = 0$), surface temperature increases and surface pressure over land drops.

The most remarkable feature of these experiments is, without doubt, the fact that in the dry soil simulation, precipitation within the ITCZ is only slightly reduced (Figure 2), indicating an important increase of moisture convergence. This change goes along with a decrease of precipitation over the ocean. On the contrary, in the extra-tropics the cancellation of evaporation is followed by reduced precipitation. These results illustrate very well the fact that precipitation in the tropics has two origins,local evaporation and large scale convergence of moisture. Under certain conditions the lack of local evaporation can be compensated by enhanced convergence

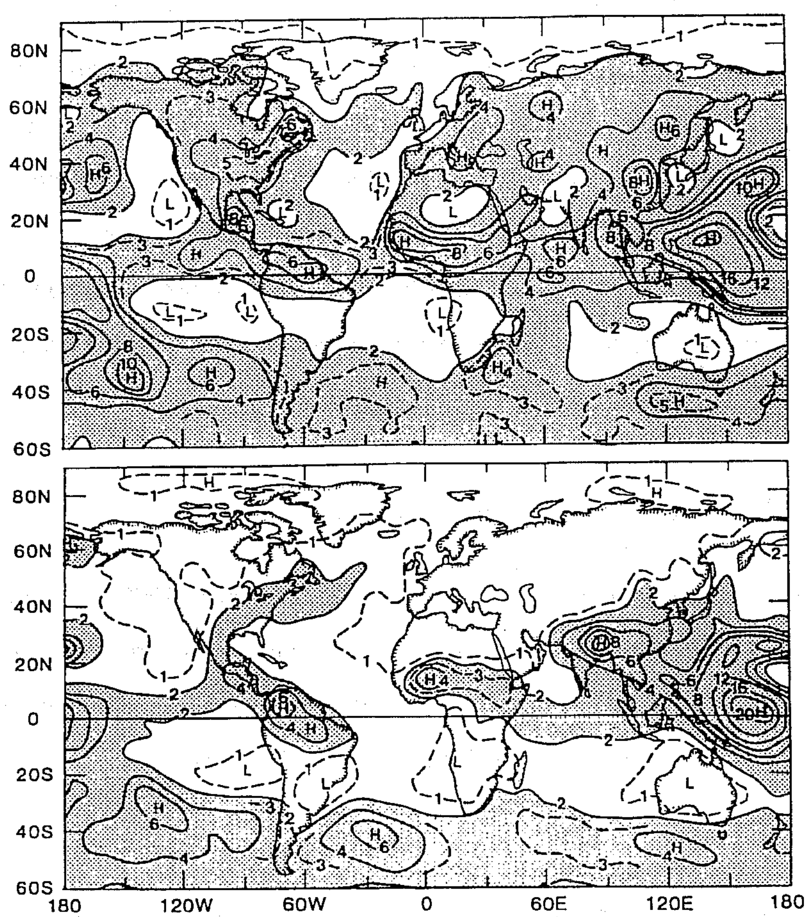

Figure 2: Precipitation (mm/d) for the wet-soil case (top) and dry-soil (bottom) in the experiments of Shukla and Mintz (1982).

of moisture.

Milly and Dunne (1994) changed evaporation by modifying the water holding capacity of soil. The comparison of two extreme cases, one with a deep reservoir allowing high levels of evaporation and the other with a shallow soil providing little evaporation, yields results similar to those of Shukla and Mintz (1982). A detailed analysis of the water transport in the tropics shows that the increase in evaporation affects the mean circulation in such a way that less water is transported from the ocean to the continents.

These results must be taken cautiously as they might depend on characteristics of the GCM used. The convection scheme for instance is one of the critical components in these experiments. Recently Dümenil (personal communication) conducted an experiment similar to Shukla and Mintz (1982) with the ECHAM3 GCM. It was found that in a dry soil case the GCM could not maintain the ITCZ over the continents as moisture convergence was not increased.

These experiments on the sensitivity of the atmosphere to large evaporation changes are simple but very instructive for the study of the hydrological cycle over continents. It would be very helpful if these experiments could be repeated regularly by climate modeling groups and inter-compared in order to improve our understanding of this sensitivity.

5 Sensitivity to roughness length changes

Changes to the roughness of the surface are perturbations of a different kind. They do not only affect the surface energy budget through changes in the turbulent fluxes but also the circulation in the lower layers of the atmosphere as the momentum exchange with the surface is modified, which in turn may impact on the hydrological cycle. Modifications of surface fluxes will be discussed prior to the impacts on the atmosphere.

Sud et al. (1988) performed July integrations in which the surface roughness was reduced on all continents to the value used over oceans (this corresponds to a reduction from 45 cm to 0.02 cm). They found that surface fluxes where not strongly affected by the change in the surface transfer coefficients. The authors explain this result by a compensation due to an increase of the gradient of humidity and temperature between the surface and the lowest atmospheric layer. More recently Lean and Rowntree (1997) performed a five-years experiment in which the surface roughness was reduced only over the Amazonian basin from its value for tropical forests to the one for pasture. This led to a small reduction in turbulent fluxes in this region indicating that the atmosphere did not fully compensate the reduction in the transfer coefficients. In contrast to Sud et al (1988), Lean and Rowntree (1997) used a complex land-surface schemes allowing more complex interactions.

In the case of the sensible heat flux, the relation between the flux, the transfer coefficient and the temperature gradient (Eq 3) is sufficiently simple to determine the atmospheric feedback in any sensitivity experiment that includes a roughness length change. This was done in a deforestation experiment by Polcher and Laval (1994), who showed that the increase in the temperature gradient compensated the reduction in roughness length, yet not by a sufficient amount to leave the sensible heat flux unchanged. It was also shown that the compensation by the atmosphere varies from one region to another. Interesting in this respect are results over Indonesia where the feedback was found to be small because of the proximity of the ocean where surface temperature is imposed.

The most obvious impact of roughness changes is found in the low level circulation and transport in the atmosphere. A decrease of the roughness reduces frictional stress, leads to stronger low level winds and larger transport of moisture over the continents. In the experiments conducted by Sud et al. (1988) this modification lead, in the tropics, to an enhancement of moisture convergence. Over most continents the ITCZ was narrowed and intensified, leading to more rainfall within the convergence zone and a reduction of precipitation in adjacent regions. One notable exception is the south-American continent. It has to be pointed out that these experiments were conducted over the month of July which is the time of the year with the least convective activity over this continent.

For the experiment in which roughness was reduced over the Amazonian basin, Lean and Rowntree (1997) reported a precipitation increase over the area of greatest convergence (Figure 3). In the annual mean, rainfall change covers a broad band from the northwest to the southeast of the continent. On the other hand, the Eastern coast of the continent experiences a reduction of precipitation. In a similar experiment carried out with the LMD-GCM (Brun, 1992), it was also found that the rainy season intensifies due to an increase in moisture convergence. The roughness reduction was applied to areas of tropical forest in Africa and South-America and the analysis was performed on the last 3 years of the simulations.

In a roughness reduction experiments over Amazonia, carried out with the UKMO GCM, Lean and Warrilow (1989) report a reduction in moisture convergence. This experiment, which covered only 8 month starting at the end of July, displayed a rather large change in evaporation. The difference to the results by Lean and Rowntree (1997) was attributed by the authors to deficiencies in the climate simulated by the old version of the model and to a change in the representation of the Andes.

The small number of roughness length change experiments performed with GCMs prevents any definite conclusions on the sensitivity of the atmosphere to this parameter. Nevertheless it appears that the atmosphere has the potential to compensate for the reduction in surface layer transfer coefficients. By reducing frictional stress the inter-tropical convergence zone is enhanced.

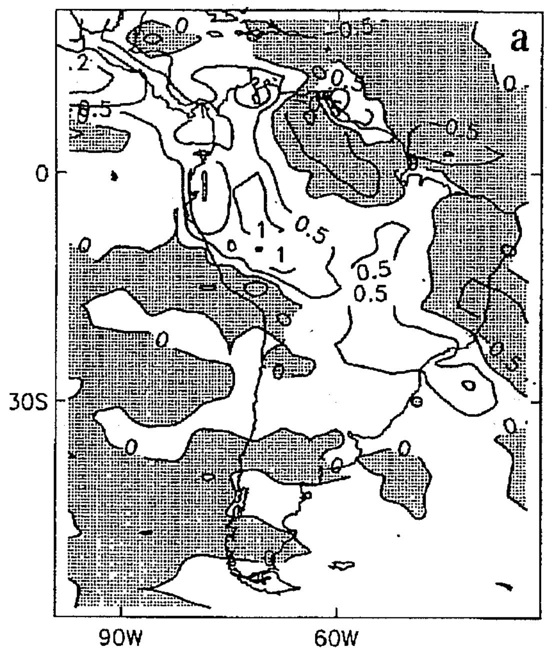

Figure 3: Annual mean precipitation (mm/d) changes induced by the roughness length reduction of the tropical forests (Lean and Rowntree (1997)).

6 Tropical deforestation experiments

Tropical forests in South-America, Africa and Southeast-Asia are being deforested to make land available for agriculture. The destruction of one of the richest ecosystems on Earth not only reduces the biodiversity but through the changes of the physical characteristics of the surface, could also impact on climate. Since complex land-surface schemes are coupled to GCMs one can attempt to model in a realistic manner the changes at the surface and study the impact on climate. The first deforestation experiment was conducted by Dickinson and Henderson-Sellers in 1988. Since then the experiments have been repeated with nearly all GCMs coupled to complex land-surface schemes. Some of the recent experiments are presented in table 1.

Two major field campaigns have been carried out recently in the Amazonian basin in order to understand the surface processes in the untouched forest and how they differ from those over a re-grown pasture. Typical values for the various parameters which are needed to characterize both surface types in complex land-surface schemes used in GCMs were also determined. The first of these campaigns was ARME (Amazon Micro-meteorological Experiment) from 1983 to 1985 (Shuttleworth et al., 1984) which was later followed by ABRACOS (Anglo-BRazilian

Reference	Model	Resolution	Surface scheme	Length
Nobre et al. (1991)	NMC	R40	SiB	1 yr.
Dickinson et al. (1992)	CCM1	R15	BATS	3 yr.
Lean and Rowntree (1993)	UKMO	2.5° × 3.75°	Warrilow et al. (1986)	3 yr.
Polcher and Laval (1994a)	LMD-3	2.0° × 5.6°	SECHIBA	1 yr.
Polcher and Laval (1994b)	LMD-6	2.0° × 5.6°	SECHIBA	11 yr.
Manzi and Planton (1996)	Emeraude	T42	ISBA	3 yr.
Zang et al.(1996)	CCM1	R15	BATS	11 yr.
Lean and Rowntree (1997)	UKMO	2.5° × 3.75°	Revised Warrilow et al. (1986)	10 yr.
Hahmann and Dickinson (1997)	CCM2	T42	BATS	8 yr.

Table 1: Some of the recent tropical deforestation experiments.

Amazonian Climate Observation Study) from 1990 to 1993 (Gash et al., 1996). Only the second one provided data for pasture as well as for forest. When pasture replaces tropical forests the three main properties of the surface are modified in the following way:

albedo: As more bare soil is exposed by pasture, its surface albedo is higher than the one for forests. Observations indicate that albedo for forests varies between 0.12 and 0.14 while for pasture it is comprised between 0.17 and 0.19 (Culf et al., 1995).

Evaporation characteristics: The capacity of tropical forests and pasture to evaporate are different due to a large number of modifications in their physical and bio-physical properties : i) The different sizes of canopy change interception loss, ii) The root systems reach different soil depths and take up water from different horizons of the soil, iii) Difference in photo-activity lead to different stomatal resistances, vi) Pasture exposes a larger fraction of bare soil. As these processes are modeled differently in each land-surface scheme it is not possible to provide a closed set of parameters for both biomes. Furthermore these parameters may have different meaning in each scheme (Polcher et al., 1996).

roughness length: The short pasture produces a smaller frictional stress than the one observed for trees. Thus a roughness length reduction has to be imposed to the model. The roughness of forested areas can be estimated at 2 m (Shuttleworth, 1988) while for pasture the value is close to 0.02 m (Wright et al., 1992).

Most deforestation experiments used the data from the ARME campaign to choose parameters for the tropical forest and validate the model over this region. Parameter values for pasture were deduced from other observations. A number of experiments used parameter modifications from previous studies to enable comparisons. Currently, the ABRACOS data set has only been used to set-up and validate one deforestation experiment (Lean and Rowntree, 1997). In all cases presented in Table 1 the entire Amazonian basin was deforested. In three of these experiments the deforestation was also performed over Africa and Southeast Asia.

The complexity of the surface change renders the analysis of the modifications of the surface fluxes difficult. Indeed, the shift in the balance of fluxes can not be attributed to a single cause as was the case in the academic experiments presented above. Furthermore the combination of the changes in the radiation, turbulent fluxes and evaporative fraction can compensate or amplify each other. Nevertheless, a few observed modifications should be reproduced in GCM experiments. The albedo increase should reduce the net radiative flux at the surface as observed

Reference	ΔR_n $[W/m^2]$	ΔE [mm/yr]	ΔT [K]	$\Delta(P-E)$
Nobre et al. (1991)	-26.0	-500.0	+2.0	Decrease ↘
Dickinson et al. (1992)	-11.0	-25.5	+0.6	Decrease ↘
Lean and Rowntree (1993)	-18.5	-198.0	+2.1	Decrease ↘
Polcher and Laval (1994a)	-12.0	-985.0	+3.8	Increase ↗
Polcher and Laval (1994b)	-14.2	-127.8	+0.1	Decrease ↘
Manzi and Planton (1996)	+1.0	-113.0	+1.3	Increase ↗
Zang et al. (1996)	-16.3	-402.0	+0.3	Decrease ↘
Lean and Rowntree (1997)	-12.7	-157.0	+2.3	Increase ↗
Hahmann and Dickinson (1997)	-10.0	-149.0	+1.0	Decrease ↘

Table 2: The impact of deforestation on the surface fluxes and the hydrological cycle for some of the recent experiments. Results are annual means averaged over an area within the Amazonian basin.

by Culf et al. (1996). This impact can however be amplified by an increase in cloud cover as indicated by two recent studies (Chu et al., 1994; Cutrim et al., 1995). The reduction in net radiation can also be mitigated by a larger long wave radiation caused by a warmer surface. Evaporation will be reduced by the combination of two effects: First, a reduction of interception loss resulting from the smaller canopy of pasture and second, the annual cycle of transpiration for both biomes will be different as pasture has a shallower root distribution than forest. The first effect will be dominant during the rainy season while the second one will prevail as the dry season sets in and the soil dries out. It is believed that the impact on transpiration will be larger than the one on interception (Wright et al., 1992).

In table 2, annually averaged results from a few recent deforestation experiments are given. All models display a reduction in net surface radiation caused by the albedo increase imposed in the deforestation scenario and partly compensated by increase of temperature and thus long wave emission. An interesting fact is that in all GCMs, except in the one which had fixed clouds (Nobre et al., 1991), cloud cover is reduced and counteracts the higher albedo in the deforestation scenario. In Manzi and Planton (1996), the albedo increase was cancelled by the combination of long wave radiation and cloud feedbacks. A reduction in evaporation is a common feature to all experiments, although a wide range of values is covered. The annual evolution of this change varies from a maximum during the dry season (e.g Lean and Rowntree, 1997),to nearly constant reduction throughout the year (e.g. Zang et al., 1996). This result depends on the ability of the land-surface scheme to represent properly the contrast in soil moisture control on transpiration for pasture and forest and the difference in interception loss for both biomes. Surface temperature is warmer in the deforestation experiment except in two cases where no significant changes could be detected in the annual mean. As discussed in section 2, surface temperature is not the variable best suited to evaluate the impact of deforestation. When comparing the first and penultimate experiments presented in table 2, it appears that their temperature changes are comparable but they differ in all other aspects and especially their impact on the hydrological cycle.

The most striking contradictions between these deforestation experiments are certainly the changes in moisture convergence. If for a moment, one is to follow the reasoning exposed by Mintz in his review paper (Mintz, 1984) on the impact of evaporation reductions over extra-tropical continents, one would expect a lower moisture convergence. It was an unexpected result when for the first time a deforestation experiment (Polcher and Laval, 1994b) displayed an increase in moisture convergence. However, if one considers the results presented in section 4 and the fact that the fraction of recycled water in the Amazon basin is estimated to be between

50% (Salati, 1987) and 25% (Eltahir and Bras, 1994), it appears that this outcome is possible. This result demonstrates that the sensitivity to surface processes is different in the tropics and extra-tropics.

The divergences of the model's results at the surface may have different causes. Three main categories can be distinguished.

- The imposed changes in surface parameters are different in all models. This is due to the fact that values are obtained from different sources but also because, by design, the land-surface schemes need different parameters. For instance, a scheme with an explicit representation of the sub-grid variability at the surface will only change albedo on the tropical forest tile which is being deforested, not over the entire grid-box. Most schemes have also parameters which can not be directly measured or for which no observations are available and thus their modification has to be estimated.

- The GCMs used are different and it is not known if they possess the same sensitivities to surface processes. For instance, cloud schemes used in these models vary from fixed clouds to highly complex prognostic ones. Thus the compensation in the net short-wave flux of the albedo increase can result from different processes. Manzi and Planton (1996) have shown that when the convection scheme is changed in the Meteo-France GCM the sensitivity is altered. The same may also be true for the land-surface schemes.

- The climate simulated by the GCMs are different and have therefore different sensitivities. For instance the length of the rainy season will vary from one GCM to the other, and thus the reduction of interception loss will be more or less important.

From the experiments which have been carried out so far it is not clear what will be the impact of tropical deforestation on the atmospheric branch of the hydrological cycle. From the academic land-surface change experiments presented above no dominant process can be identified. The lower net surface radiation should by itself tend to reduce moisture convergence, on the other hand the lower roughness of pasture should favor the converging low level fluxes in the region. Finally, the atmosphere has the potential to compensate for the reduced evaporation with an increase in moisture convergence. At this point it is not clear how these processes combine when the three surface properties are changed simultaneously. Lean and Rowntree (1997) put forward the hypothesis that the linear combination of the roughness length and albedo changes explains the largest fraction of the atmospheric impact. However, no experiment is presented for the sensitivity of the atmosphere to evaporation changes only to prove that a simple linear approximation is valid.

Surface processes and tropical climate are characterized by short time scales and the forcing to the atmosphere is local, thus it is difficult to analyze their interactions with space-tim

7 Sensitivity of convection to land-surface changes

One possible refinement in the analysis of the interactions between land-surfaces and the tropical atmosphere is to separate the different synoptic situations and to study them separately. To discriminate between these situations one may use the intensity of the energy cycle of convection. The method introduced by Polcher (1995) will be described in this section.

The atmospheric column is characterized by its vertically integrated energy balance. In the tropics, this simple model is very useful as it helps to identify convective system and measure their intensity. Convection may be viewed (Figure 4) as a themodynamical engine which takes energy received from the surface, from large-scale convergence of enthalpy and latent energy and from absorbed radiation to transform it into potential energy. On the other hand, subsidence

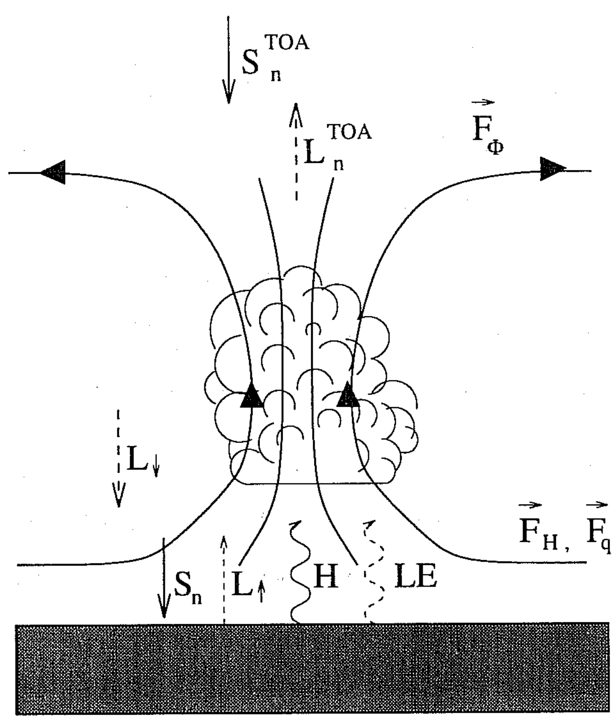

Figure 4: A simple model for the energetics of tropical convection.

is the transformation of potential energy into enthalpy which, combined with the surface fluxes, produces a divergence of enthalpy and latent energy. The equations describing this system are discussed in detail by Polcher (1995). This simple model of convection can be applied to single convective systems as well as to the Hadley-Circulation. To study the surface processes, the measure best suited to characterize the intensity of convection is the divergence of potential energy ($\nabla \cdot \vec{F}_\phi$). Three main reasons can be given :

- Divergence of potential energy is independent of the surface fluxes.
- It is the energy which is exported from the convective regions and transported to the areas of subsidence.
- Potential energy is a very sharp measure of convection as it is the largest single flux in the balance of energy.

This measure of convection is computed with daily averaged data for each grid-box in the region of interest. The data points are then binned according to their divergence of potential energy and the relations to the other energy fluxes are studied. This diagnostic can also be applied at higher frequencies as long as the variations in the storage of energy within the column

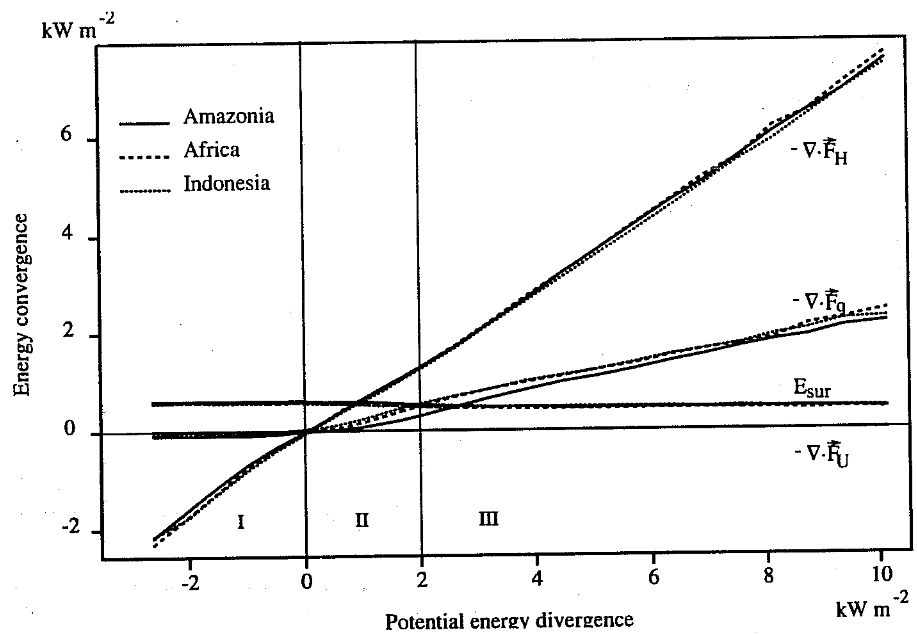

Figure 5: Enthalpy, latent energy, and kinetic energy flux convergence and energy from the surface as a function of the potential energy divergence.

is small compared to the fluxes. Individual peaks in the divergence of potential energy can be tracked if the time evolution of convection needs to be studied.

This binning method is applied to the three main regions of tropical forests (Amazonia, The Congo Basin and Indonesia) in the control integration of the deforestation experiment presented in Polcher and Laval (1994b). Figure 5 shows the relation between the divergence of potential energy, the convergence of enthalpy ($-\nabla \cdot \vec{F_H}$) and latent heat ($-\nabla \cdot \vec{F_q}$) and energy from the surface (E_{sur}). The striking feature in this graph, is that in all three regions, the relationship between the different fluxes is very similar. This indicates that the divergence of potential energy is a general characteristic of convective events in the LMD-GCM. The diagnostic was also applied to the BMRC GCM and a similar figure was obtained.

In Figure 5 three regimes of convection can be distinguished.

I: This class contains the situations of subsidence. The sources of energy for the column are the convergence of potential energy and the fluxes from the surface. In these situations, energy is exported by the divergence of enthalpy and to a lesser extend, latent energy.

II: Within this interval the events will be considered to be weakly convective. The energy received from the surface is of a magnitude similar to the convergence of enthalpy and latent heat. These events produce potential energy.

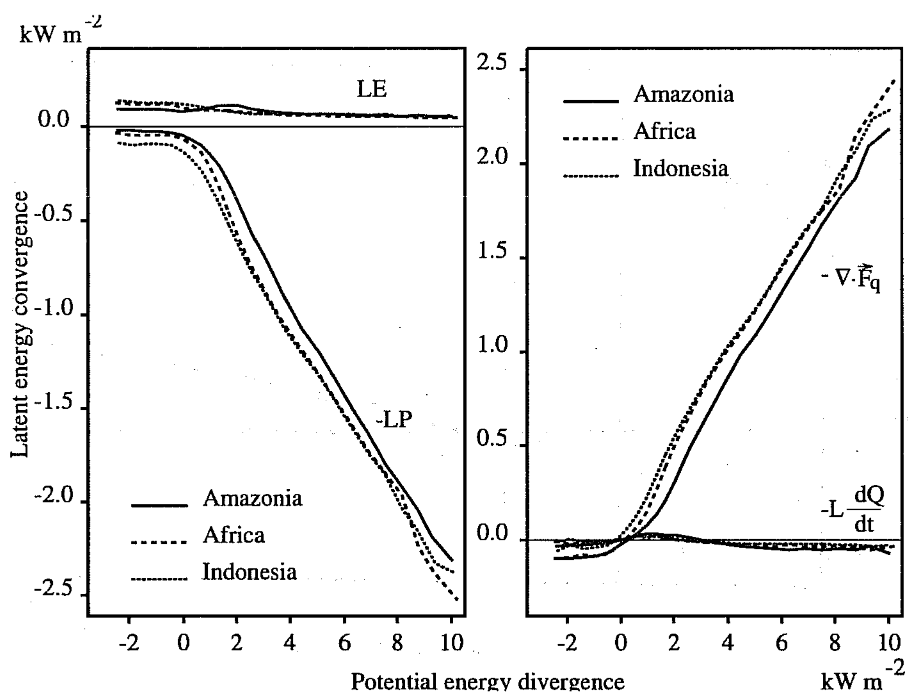

Figure 6: The water cycle of convection as a function of the divergence of potential energy.

III: For large values of divergence of potential energy convection is dominated by the convergence of enthalpy and latent heat. These are the intense convective events.

As convection intensifies, the convergence in the lower levels of the atmosphere increases and the area from which energy is drawn becomes larger. Class III contains the events which make up the ITCZ or the monsoons and interact with the large scale circulation.

The water cycle can be analyzed over the spectrum of values of potential energy divergence in the same way as it was done for the energy cycle of convection. The water balance equation for the atmospheric column may be written in terms of energy as:

$$L\frac{\partial Q}{\partial t} = LE - LP - \nabla \cdot \vec{F_q}. \qquad (7)$$

Where L is the latent heat of evaporation and Q precipitable water. Figure 6 shows the four terms of the water balance as function of the intensity of convection. One may note that, again, all three regions have a similar behavior and that characteristic properties can be defined for each class of convective events. The three classes of events defined above display contrasting water cycles.

I: In this class, local evaporation feeds precipitation and the divergence of latent heat. Depending on the region, the fraction of water recycled or exported will vary.

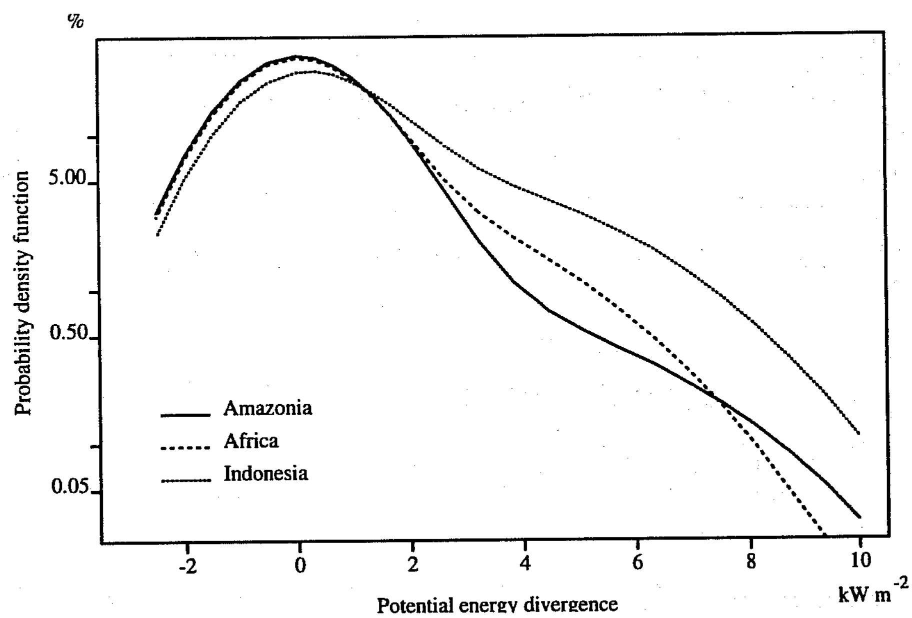

Figure 7: The probability density function for potential energy divergence.

II: Within this interval, the convergence of latent heat changes sign. This means that weak convective events can feed on local evaporation or moisture convergence. The column may also export water for convection at the lower end of the interval.

III: This class of intense convective events is dominated by the convergence of moisture and local evaporation plays a limited role. Precipitation is proportional to $-\nabla \cdot \vec{F_q}$.

Using Figure 6, a first estimation can be made of the sensitivity of precipitation to local evaporation depending on the convective regime. Precipitation of Class I events will be sensitive to local evaporation while in Class III a reduction in evaporation can be compensated by an increase in convergence.

Although the characteristics of convection are similar in all three regions the climate is different because each class of events has a different probability of occurring. Figure 7 shows that in Indonesia Class III events are more frequent and thus more rain and moisture convergence is produced, whereas Amazonia is the region with the smallest number of Class III events in the LMD-GCM. The difference in precipitation between these regions is the reflection of the variation in the number of events. If we compare the annual cycles of monthly mean probabilities for all three classes we note that Class III is predominant during the rainy season while Class II and I are typical situations during the dry season.

Using the tool described above, the deforestation experiment presented in Polcher and Laval

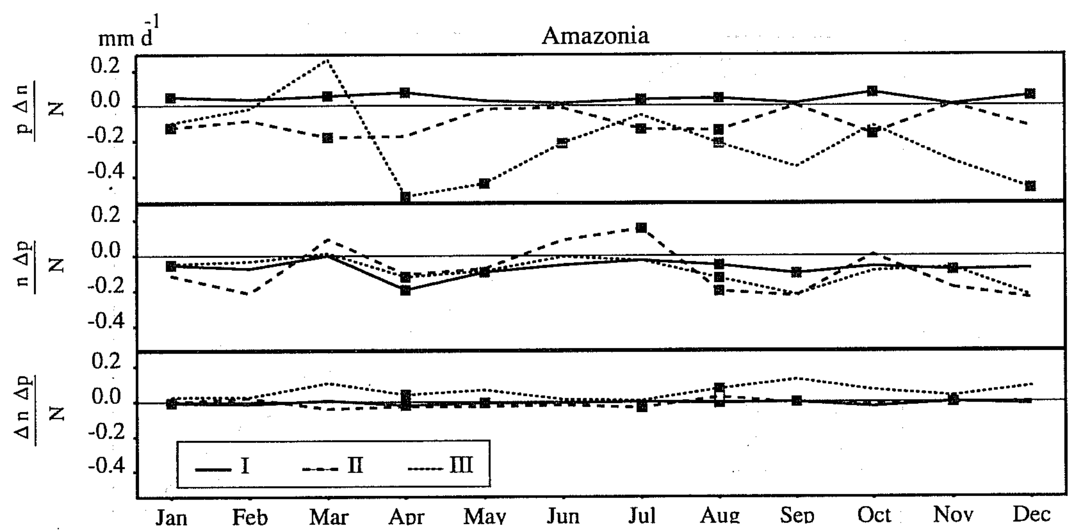

Figure 8: Decomposition for all three classes of the precipitation changes obtained over the Amazonian region in the deforestation experiment presented in Polcher and Laval (1994b). Black boxes indicate changes that are statistically significant within a 95% confidence interval.

(1994b) was analyzed. In the present summary we will only describe the conclusions reached on the precipitation change which was observed in this experiment. As a characteristic precipitation can be defined for each of the convective classes the monthly mean rainfall (P) over a region can be written as :

$$P = \frac{1}{N}\sum_{i=1}^{m} n_i p_i \tag{8}$$

$$\text{with } N = \sum_{i=1}^{m} n_i. \tag{9}$$

Where class i has n_i number of convective events with a characteristic precipitation p_i. The change in precipitation in the deforestation experiments (ΔP) can be decomposed in changes in the number of convective events (Δn_i) and changes in the characteristic precipitation (Δp_i) :

$$\Delta P = \frac{1}{N}\sum_{i=1}^{m}(p_i \Delta n_i + n_i \Delta p_i + \Delta n_i \Delta p_i). \tag{10}$$

The three terms within the sum give for each class the monthly mean change in precipitation which can be attributed to the number of events, the characteristic precipitation and their covariance. Figure 8 displays each of them for the three classes over the Amazonian region.

The first result which can be noted in Figure 8 is the importance of the precipitation changes associated with the reduction of the number of events in class III (upper panel). Furthermore this decrease is statistically significant during the months of May and June, which could not be shown on the regionally averaged precipitation. In Class III, the characteristic precipitation is reduced but this is only a second order change. On the other hand, in Class I a slight increase in precipitation is caused by a higher frequency of events but the reduction in characteristic precipitation is more important. The deforestation caused a reduction in evaporation which in class I reduced the local recycling of water. It also led to a lower frequency of class II events, but the impact on their characteristic precipitation is very variable.

An interesting aspect of this result is that Le Barbé and Lebel (1997) found a similar behavior in the precipitation record in the sub-Saharian region. A statistical method was applied to the daily rainfall observations available over the region to distinguish between the number of precipitating systems and the amount of rainfall they produce. The authors found that the inter-annual variation of precipitation is dominated by the variation of the number of systems and that the change in characteristic rainfall is only minor. Thus it appears that the response of the GCM described above is associated to processes which can also be found in the real world.

To study the sensitivity of the number of convective events ten variations of the deforestation experiment reported in Polcher and Laval (1994b) were conducted. The parameters of pasture in these experiments were varied according to the known uncertainties (Polcher, 1995). These experiments, performed with the same GCM, cover a range of results nearly as large as the one presented in Table 2. Some simulations showed an increase in moisture convergence over Amazonia and others a reduction. Over the three deforested areas (Amazonia, Africa and

Indonesia) the change in the number of convective events and characteristic precipitation for each class was correlated to the variations of the local surface fluxes. Relations were found between evaporation and the characteristic precipitation for Classes I and II while the number of convective events was correlated to the sensible heat flux in Classes II and III. The results are summarized in Figure 9.

The different deforestation scenarios chosen in the LMD-GCM lead to increases as well as reductions of the sensible heat flux. If H is reduced then the frequency of convection is decreased, this tends to diminish regional precipitation. An increase in the frequency of subsident events is also found. A higher sensible heat flux leads to the opposite effect. The lower evaporation associated with deforestation modulates this mechanism as it reduces characteristic precipitation in the classes which include weak convection and subsidence. The upper branch in Figure 9 corresponds to the results described in section 4. In these experiments the reduction in evaporation was imposed and it induced an increase in sensible heat flux which led in the tropics to

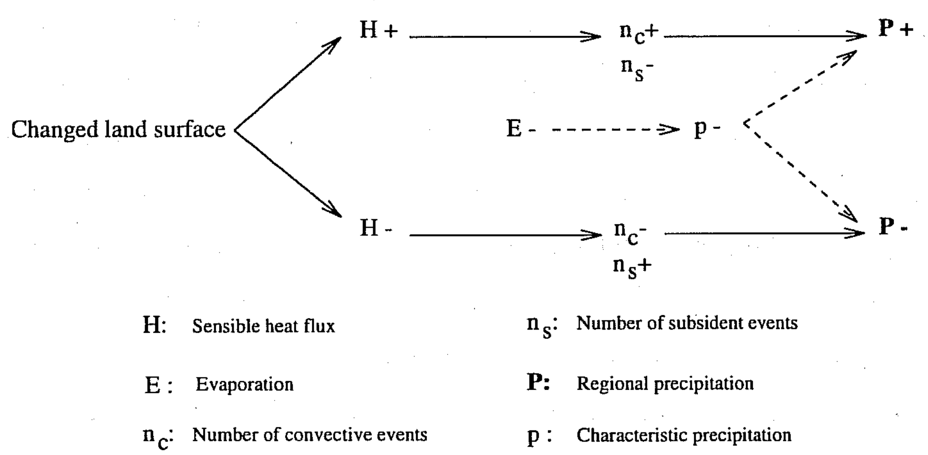

Figure 9: A simple mechanism for the sensitivity of tropical convection to land-surface changes in the LMD-GCM

a larger moisture convergence. On the other hand, the lower branch can be compared to the results presented in section 3. In these simulations a reduction in the turbulent fluxes caused by higher albedo leads to a reduction in convection and precipitation.

An interesting consequence of this mechanism is that the total result on regional precipitation will depend on the distribution of convective and subsident events. The climate of the control experiment will be a determining factor for the outcome of the deforestation experiment. This might explain the diversity of results for the deforestation experiments described in Table 2.

An important caveat in this result is that the correlation was performed with the changes in surface fluxes of the grid-box where convection occurs. In the case of an homogeneous deforestation this is not a major restriction as it can be assumed that the local flux variations are similar to those of the neighbouring grid-boxes. In order to generalize the results three long experiments were carried out with the LMD-GCM where the deforestation was distributed over the Amazonian basin along three different patterns. First results show that the atmospheric feedback can reduce or amplify the changes in surface fluxes and that the impact on precipitation are largely independent of the deforestation pattern used.

8 Conclusions

The sensitivity of the tropical climate to land-surfaces processes is a difficult topic because the atmosphere is able to respond in a multitude of ways, which at times appears contradictory. Because of the short time scale which characterizes the tropical atmosphere traditional analysis procedures may not be sufficient to understand its sensitivities. Averages do not take into account the fact that the atmosphere may change rapidly from one type of sensitivity to another.

A number of tropical deforestation experiments have been conducted over the recent years but divergences persist and it cannot be said what will be the most likely impact of this anthropogenic change. One reason for this uncertainty can be found in the lack of theoretical framework or a well established mechanism which could offer some guidance. To solve this dilemma one could try and impose exactly the same deforestation scenario to all GCMs and compare the results. At present this does not seem feasible as the surface changes are not sufficiently well known to cover the parameters of all land-surface schemes. Another solution would be to perform a number of simple academic experiments, similar to those described in sections 2, 3 and 4. These experiments would have to be simple enough to ensure that the same surface change is performed by all models. The experiments do not need to be long but the data has to be analysed at all time-scales. Comparing the results of a large number of GCMs would help to test the mechanisms currently proposed or to develop new ones.

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