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A model for soil-vegetation-atmosphere interactions in water-limited ecosystems

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[1] We study the interaction between atmosphere, soil moisture, and vegetation in water-limited environments with significant water recycling, and introduce a simple process model including some of the main feedbacks active in the system. In our model, the soil-vegetation-atmosphere dynamics display two stable states for realistic values of the synoptic moisture convergence flux. Starting from low soil moisture and/or low vegetation cover, the system reaches a dry and hot state, whereas it reaches a wet and cool state when starting from higher initial values of soil moisture and of vegetation cover. The role of synoptic perturbations is investigated by inserting a stochastic input of moisture: in this case, a bimodal distribution of soil moisture is obtained. We explore the difference between the dynamics of natural vegetation, capable of adjusting its areal extent to variations in soil moisture, and cultivated vegetation, whose areal extent cannot vary. The model results indicate that the presence of natural vegetation increases the probability of reaching a wet/cool state with respect to the case of cultivated plants.

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1. Introduction

[2] Soil moisture, terrestrial vegetation, and atmospheric flows are parts of a complex interacting system, characterized by the presence of many feedback mechanisms between the various components [*Claussen*, 1998; *Hayden*, 1998; *Hurtt et al.*, 1998; *Pielke et al.*, 1998; *Bonan*, 2002]. For example, soil moisture anomalies, associated with the lack of precipitation, have been suggested to play a crucial role in the variability of the large-scale water balance [*Shukla and Mintz*, 1982; *Rodriguez-Iturbe et al.*, 1991; *Entekhabi et al.*, 1992] and in determining the strength of summer heat waves at continental midlatitudes [*Fischer et al.*, 2007; *Schär et al.*, 1999; *Vautard et al.*, 2007], as well as in determining the rainfall variability associated with the African monsoon [*Douville et al.*, 2001; *Zeng et al.*, 1999; *Nicholson*, 2000; *Savenije*, 1995; *Xue et al.*, 2004].

[3] Analogously, observations and General Circulation Model (GCM) simulations indicate that desertification and deforestation can generate a reduction in precipitation in many regions of the world, such as the Tropics [e.g., *Henderson-Sellers et al.*, 1993; *Chase et al.*, 1996; *Kanae et al.*, 2001], Southern Mongolia and China [*Xue*, 1996; *Zhang et al.*, 2003, 2005]. Deforestation experiments with regional climate models have been performed for the Iberian Peninsula, where the largest precipitation reduction and temperature increase in response to land degradation was observed in the summer season [*Gaertner et al.*, 2001; *Arribas et al.*, 2002]. Reforestation in Southern Israel has been suggested to be responsible for local increase in rainfall [*Otterman et al.*, 1990; *Ben-Gai et al.*, 1998], though in this case the contributions from irrigation and land use change cannot be clearly distinguished. These results indicate that ecosystem dynamics cannot be considered just as a passive boundary condition to atmospheric flows, but rather they are one of the leading actors on stage [see, e.g., *Foley et al.*, 2003; *Zeng et al.*, 1999; *Claussen*, 2004; *Scheffer et al.*, 2005; *Moorcroft*, 2003].

[4] Soil-vegetation-atmosphere feedbacks are particularly important in water-limited ecosystems, where water is the main factor controlling vegetation growth [*Rodriguez-Iturbe and Porporato*, 2004]. Through evapotranspiration, vegetation determines the flux of moisture from the soil to the atmosphere, i.e., the surface moisture flux. This can have significant effect on the local precipitation [*Eltahir and Bras*, 1996b; *Findell and Eltahir*, 1997; *Juang et al.*, 2007b; *Alfieri et al.*, 2008] and on the hydrologic cycle [*Shukla and Mintz*, 1982; *Anthes*, 1984; *Cadet and Nnoli*, 1987; *Xue and Shukla*, 1993; *Cook*, 1994; *Moorcroft*, 2003], especially in regions where local precipitation recycling plays a relevant role [*Eltahir and Bras*, 1996; *Koster et al.*, 2004], such as the Sahara-Sahel region [*Brubaker et al.*, 1993; *Douville et al.*, 2001; *Koster et al.*, 2004] or continental Europe during summer [*Schär et al.*, 1999]. Another well-known land-atmosphere feedback in arid lands is the Charney mechanism [*Charney*, 1975]. Because of lower albedo and longwave emission, heat losses are reduced over vegetated areas as compared to bare soil, possibly resulting in large-scale changes of the monsoonal circulation and enhanced precipitation on vegetated areas, eventually leading to a positive feedback on the vegetation itself.

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[5] Given the complexity of the soil-vegetation-atmosphere system, simplified, process-oriented approaches are a much needed step to shed light on the basic mechanisms at play. In the past ten years, the definition of simplified models of soil moisture dynamics in arid and semi-arid regions has received much attention (see *Rodriguez-Iturbe and Porporato* [2004] and references therein). In some cases, the precipitation field was considered as an external input and the vegetation water stress was parameterized out of soil moisture levels [*Laio et al.*, 2001; *Porporato et al.*, 2001; *Rodriguez-Iturbe et al.*, 1999]. In other contributions vegetation stress was related to carbon assimilation [*Daly et al.*, 2004]. Other models included a feedback of soil moisture on precipitation [*Entekhabi et al.*, 1992; *Brubaker and Entekhabi*, 1995, 1996; *Entekhabi and Brubaker*, 1995; *Porporato et al.*, 2000; *Margulis and Entekhabi*, 2001; *D'Odorico and Porporato*, 2004; *D'Andrea et al.*, 2006; *Abbot and Emanuel*, 2007], attempting at closing the loop between soil and atmosphere. The simplified, mechanistic model of the soil-atmosphere continuum developed by *D'Andrea et al.* [2006] displays multiple equilibria for realistic values of atmospheric moisture flux. This result was interpreted as an indication that the occurrence of summer heat waves at continental midlatitude is strictly associated with the presence of soil moisture anomalies, and it provided an explanation of the observed bimodality of summer soil moisture in continental areas of the United States [*D'Odorico and Porporato*, 2004; *Lee and Hornberger*, 2006].

[6] Other approaches included vegetation dynamics. Some models directly link vegetation cover and precipitation rate, with vegetation represented either with an implicit-space approach [*Brovkin et al.*, 1998; *Scheffer et al.*, 2005] or with an explicit consideration of the spatial aspects, able to include small-scale feedbacks [*Janssen et al.*, 2008]. *Dekker et al.* [2007] introduce both microscale feedbacks (through an explicit-space model for vegetation and soil water) and large-scale feedbacks with a simplified hydrologic cycle as represented by *Entekhabi et al.* [1992]. All of these works indicate the existence of multiple equilibrium states in the atmosphere-vegetation system.

[7] In the following we introduce a simple box model for soil-vegetation-atmosphere dynamics in water-limited ecosystems, specific for regions where local water recycling is important. The model provides a simple physical representation of the moist thermodynamics of the Planetary Boundary Layer (PBL), including lateral moisture fluxes, convection and precipitation. Soil moisture drives vegetation dynamics. Both the transpiration and vegetation-albedo feedbacks are included. The soil-atmosphere continuum is described as in *D'Andrea et al.* [2006] (hereafter referred to as DA), whereas soil-vegetation interactions are modeled as in *Baudena et al.* [2007] (BA in the following). We show that the soil-atmosphere dynamics, known to display multiple stable equilibria, can be significantly affected by the conditions and dynamical response of vegetation cover.

2. Model Description

[8] The model adopted here describes the bulk dynamics of the water cycle in a large continental region and it includes an atmospheric shallow layer and a surface soil layer, as in DA, and an implicit-space representation of vegetation cover, as in BA. The model describes the temporal dynamics of five prognostic variables: average potential temperature (θ_a) and average humidity (q_a) of the atmospheric PBL, average temperature (T_s) and moisture (q_s) of the active soil layer, and the fraction of soil surface occupied by vegetation (b). The dynamics of these quantities is described by the following system of Ordinary Differential Equations (ODEs):

$$\rho c_p h_a \frac{\mathrm{d}\theta_a}{\mathrm{d}t} = \epsilon_a \epsilon_s \sigma T_s^4 + Q_s - \rho c_p h_a \frac{\mathrm{d}\Delta\tilde{\theta}_a}{\mathrm{d}t} + \frac{1}{\tau_a}(\theta_a^* - \theta_a) \qquad (1)$$

$$\rho h_a \frac{\mathrm{d}q_a}{\mathrm{d}t} = E - \rho h_a \frac{\mathrm{d}\Delta\tilde{q}_a}{\mathrm{d}t} + F_q \qquad (2)$$

$$\rho_s c_{ps} h_s \frac{\mathrm{d}T_s}{\mathrm{d}t} = (1-\alpha)F_{rad} - Q_s - \epsilon_s \sigma T_s^4 - L_e E \qquad (3)$$

$$w_0 h_s \frac{\mathrm{d}q_s}{\mathrm{d}t} = I - E - L \qquad (4)$$

$$\frac{\mathrm{d}b}{\mathrm{d}t} = gb(1-b) - \mu b \qquad (5)$$

Details are given by DA for equations (1)–(4) and by BA for equations (4) and (5). Names and values of the different parameters are reported in Table 1. Parameter values correspond to the case of a continental midlatitude summer. Qualitatively analogous results are found using parameter sets appropriate for other water-limited environments, as discussed below.

[9] In equation (1), the atmospheric potential temperature, θ_a, is increased by the infrared radiation, $\epsilon_a\epsilon_s\sigma\, T_s^4$, and by the sensible heat flux, Q_s, from the soil surface. Here, Q_s is given by an aerodynamic bulk formula, i.e., it is proportional to the soil-atmosphere temperature difference. Atmospheric temperature is reduced by the effect of convection (third term on the r.h.s., see section 2.1), and it relaxes to a fixed external temperature θ_a* to represent heat exchanges due to transport and mixing with the upper atmosphere.

[10] In equation (2), air moisture increases because of evapotranspiration E from soil and vegetation (see the discussion below) and it decreases in case of convection because of moisture uplift into the free troposphere. Part of the uplifted water will precipitate back to the soil (see following paragraphs for further details). The third input term is a lateral moisture forcing, F_q, used to represent large-scale mass field convergence, e.g., an influx of moist air of maritime origin. The lateral moisture flux convergence F_q is the main control parameter of the system, and it is discussed in detail below.

[11] Equation (3) represents the soil temperature balance: from left to right on the r.h.s., the terms represent incoming net radiation $(1-\alpha)F_{rad}$, sensible heat flux to the atmosphere

Table 1. List of Parameters of the Soil-Vegetation-Atmosphere Model

Symbol	Meaning	Value	Units
F_{rad}	Net radiation at surface	480	W m^{-2}
α_b	Albedo of vegetated surface	0.14	
α_0	Albedo of bare soil surface	0.35	
L_e	Specific latent heat of evaporation	$2.5 \cdot 10^6$	J kg^{-1}
c_p	Air specific heat	1000	J kg^{-1} K^{-1}
c_{ps}	Soil specific heat	1000	J kg^{-1} K^{-1}
h_a	Thickness of the atmospheric boundary layer	1000	m
h_s	Depth of the active soil layer	0.5	m
w_0	Soil water holding capacity	400	kg m^{-3}
ε_a	Blackbody absorptivity of the PBL	0.25	
ε_s	Blackbody emissivity of the soil	0.85	
ρ	Air density	1	kg m^{-3}
ρ_s	Soil density	1800	kg m^{-3}
C_D	Bulk aerodynamic drag coefficient	0.008	
$E_{\max}$	Maximum potential evapotranspiration at q^*	$1.5 \cdot 10^{-4}$	kg m^{-2} s^{-1}
E_w	Potential evapotranspiration at q_w	$1.04 \cdot 10^{-6}$	kg m^{-2} s^{-1}
q_h	Soil hygroscopic point	0.14	
q_w	Soil wilting point	0.18	
q^*	Minimum soil moisture value with maximum plant fitness	0.46	
q_{fc}	Soil field capacity	0.56	
K_s	Saturated hydraulic conductivity	$3.5 \cdot 10^{-3}$	kg m^{-2} s^{-1}
β	Water retention parameter	14	
θ_a^*	Relaxation temperature for PBL	295	K
τ_a	Relaxation time for PBL	$2.6 \cdot 10^5$	s
θ_e^*	Free troposphere equivalent potential temperature	305	K
$\bar{v}_s$	Mean wind	6	m s^{-1}
f	Precipitation efficiency Weak water updraft (<1 mm d^{-1})	0.2	
	Strong water updraft (>3 mm d^{-1})	0.9	
g_0	Colonization rate of natural vegetation for $\theta_\sigma \gg \theta^*$	0.8	y^{-1}
μ_2	Extinction rate of natural vegetation for $\theta_\sigma \ll \theta_\omega$	0.5	y^{-1}
μ_1	Extinction rate of natural vegetation for $\theta_\sigma \gg \theta_\omega$	0.1	y^{-1}
a	Hyperbolic tangent width	0.005	

Q_s, outgoing infrared radiation $\epsilon_s \sigma T_s^4$, and latent heat flux due to evapotranspiration $L_e E$. The parameter α represents surface albedo, which is largest over bare soil and minimum over vegetated surfaces, i.e., $\alpha = \alpha_b b + \alpha_0(1 - b)$, where b is the vegetated fraction of the soil surface, see equation (5). We choose α_0 = 0.35 and α_b = 0.14 as in *Charney* [1975]. Although Charney probably overestimated albedo variations [*Nicholson et al.*, 1998; *Xue*, 2006], our results do not qualitatively change for other more realistic choices of α_0 and α_b (e.g., with a difference between bare and vegetated soil of about 0.1).

[12] Equation (4) is motivated by the approach of *Rodriguez-Iturbe and Porporato* [2004]. Infiltration from rainfall, I, represents the input term for relative soil moisture, q_s ($0 \leq q_s \leq 1$). Infiltration is equal to precipitation until the soil is saturated (q_s = 1), and above this level the excess water leaves the system as surface runoff. Rainfall occurrence depends on the onset of convection, see below for further details. Output terms are evapotranspiration E, as in (2), and leakage L, due to relatively rapid losses from the bottom of the soil layer by gravity when soil moisture exceeds the field capacity, q_{fc} [*Laio et al.*, 2001].

[13] Vegetation dynamics are represented in equation (5) by an implicit-space equation for the fraction of soil surface covered by vegetation, b ($0 \leq b \leq 1$), similar to *Tilman* [1994]. Vegetation cover grows because of colonization of new sites, represented by the first term on the r.h.s. This term is given by the colonization rate of new sites, gb, a combination of the propagule production rate and survival ability of seedlings, multiplied by the fraction of space available for colonization, $(1 - b)$. The second term on the r.h.s. is the local extinction rate.

[14] Vegetation dynamics are driven by soil moisture, through the dependence of colonization rate, g, and extinction rate, μ, on q_s. In the following, we introduce two different types of vegetation: natural vegetation, that is capable of colonizing new sites if the conditions are favorable, and cultivated vegetation, that cannot spontaneously modify its distribution. For natural vegetation, we follow the approach discussed by BA and take g as a constant, g_0 for soil moisture above the smallest soil moisture level still allowing for fully-open stomata, q^*. The colonization rate abruptly decreases below this threshold (we model it with a steep hyperbolic tangent centered at q^*) and it tends to zero for $q_s < q^*$. The local extinction rate μ assumes two different constant values, μ_1 and μ_2 with $\mu_1 < \mu_2$, respectively above and below the soil moisture threshold corresponding to the plant wilting point, q_w, with $q_w < q^*$. Again, we use a steep hyperbolic tangent shape centered at q_w. Cultivated vegetation is simply assumed to be constant in time, with null colonization and mortality rates. In this case, vegetation becomes a control parameter instead of a dynamic variable.

[15] The above model represents soil-vegetation-atmosphere dynamics in regions where water recycling is important and vegetation is potentially under water stress. These conditions are typically met during the wet season in the Tropics, and during summer (the dry season) at continental midlatitudes [DA; *D'Odorico and Porporato*, 2004]. For example, heat waves in Europe have been shown to affect plants more

through drought than through high temperatures [*Reichstein et al.*, 2007].

2.1. Convection

[16] In the model considered here, as given by DA, convection is parameterized in a simple way, reminiscent of a classical approach of atmospheric and oceanic modeling: convective adjustment [see, e.g., *Glickman*, 2000]. The stability of the air column is assumed to depend only on the difference in moist enthalpy between the PBL and the free atmosphere above, which is taken to have a fixed equivalent potential temperature, θ_e*. Whenever the PBL equivalent potential temperature, $\theta_e = \theta_a\, e^{\frac{L_e q_a}{c_p \theta_a}}$ (where L_e is the latent heat of condensation and c_p is specific heat at constant pressure) is larger than θ_e*, convection occurs, cooling and drying the PBL until $\theta_e \leq \theta_e$* again and static stability is restored. According to its definition, both the cooling and drying terms decrease equivalent potential temperature.

[17] To separately estimate cooling and drying terms, we follow the approach of DA and assume conservation of relative humidity. This choice is consistent with the bulk representation of convection over a large area, as we do here, rather than with the dynamics of a single storm. The above assumptions lead to the formulation of cooling and drying terms $\tilde{\theta}_a$ and $\tilde{q}_a$ in equations (1) and (2). Exact expressions for $\tilde{\theta}_a$ and $\tilde{q}_a$ are given by DA.

2.2. Precipitation

[18] Convection lifts water from the PBL into the free troposphere. A fraction *f* of this uplifted moisture is assumed to precipitate locally, while the rest leaves the box by divergence in the upper troposphere. As DA, we make the hypothesis that the fraction of locally precipitating moisture, called precipitation efficiency, is an increasing function of the intensity of the convective flux. This physical hypothesis is at the basis of the soil moisture-precipitation feedback and it determines the amount of local recycling of precipitation. DA showed that it is in agreement with the results reported by *Schär et al.* [1999] for midlatitude summer conditions. This hypothesis is in agreement, albeit qualitatively, with *Neelin et al.* [2008], that showed a rapid increase of precipitation past a given threshold, as a function of precipitable water in the tropics. *Juang et al.* [2007b] showed that convection takes place only if the relative humidity of air is larger than a minimum threshold value.

2.3. Evapotranspiration

[19] Following BA, we assume that evapotranspiration, E_T, occurs only in the vegetated part of the box. On bare soil, only evaporation, E_0, takes place. The total water flux E is given by the sum of the two contributions, modulated to take into account the saturation deficit of the atmosphere and weighted respectively by the fraction of vegetated surface, b, and of bare soil, $1 - b$:

$$E = \tilde{E}\frac{q_{sat} - q_a}{q_{sat}} \tag{6}$$

$$\tilde{E} = bE_T(q_s) + (1 - b)E_0(q_s)\,, \tag{7}$$

where E_T and E_0 are functions of soil moisture and q_{sat} is a function of PBL temperature. They both vanish below the soil hygroscopic point q_h and increase linearly from zero to E_w at the plant wilting point q_w, above which transpiration from vegetated soil occurs. Above q_w, the forms of E_T and E_0 differ. The evaporation rate from bare soil, E_0, is assumed to continue its linear growth up to soil field capacity, q_{fc}, where it becomes constant. Owing to plant transpiration, E_T increases at a higher rate and it reaches the value of $E_{\max}$ at q*, the plant zero-stress point, above which plants transpire approximately at a constant rate.

[20] In this approach, vegetation dynamics are assumed to depend only on soil moisture, as evapotranspiration, colonization and extinction rates depend only on q_s. This hypothesis is reasonable for water-limited ecosystems [BA; *Rodriguez-Iturbe and Porporato*, 2004], where nutrient availability often is not an independent limiting factor, as it is usually associated with water availability.

[21] Owing to the limit imposed by atmospheric saturation, the actual evapotranspiration, E, is always lower than $\tilde{E}$ (the atmosphere ought to be totally dry in order to achieve $E = \tilde{E}$). For this reason, we fix the maximum evapotranspiration, $E_{\max}$, to a value higher than usually considered (13 mm d^{-1}), as the maximum value is never actually reached because the atmosphere is never completely dry. In this way, the maximum evapotranspiration actually achieved is close to 4 mm d^{-1}, comparable to e.g. *Laio et al.* [2001].

3. Multiple Equilibria in the Soil-Vegetation-Atmosphere System

[22] In the box model of DA without explicit vegetation dynamics, the soil-atmosphere system displays two stable equilibria in a realistic range of lateral moisture flux values. The main factor determining which equilibrium is approached is the initial value of soil moisture: for the same synoptic conditions, summers starting with dry soil have higher chance of ending up in a drought than those starting with moist soil.

[23] Explicitly including vegetation dynamics maintains the two equilibria, respectively corresponding to a "dry" and a "wet" state. In this case, the initial value of vegetation cover plays an essential role, as shown in Figure 1 for the case of natural vegetation. Abscissae represent the initial condition of q_s and ordinates that of b, both varying from 0 to 1. The initial values of the other three variables are kept constant; similarly to the case treated by DA, variations in the initial values of these other variables do not play a relevant role in determining which equilibrium is reached. Starting from low q_s and/or b, the system reaches a "dry" state (white part in Figure 1a), with relatively high air temperatures (~21°C with the chosen set of parameters), soil moisture below q*, and no vegetation ($b \approx 0$, see the complete list of the values of the prognostic variables in Table 2). Soil temperature and sensible heat fluxes are high, while air humidity, precipitation and evaporation are low.

[24] When the initial soil moisture and vegetation cover are large enough, the system reaches the "wet" equilibrium, (shaded area in Figure 1a), with high soil and atmospheric humidity and lower temperatures for the PBL (~14° C) and the soil. Vegetation reaches its carrying capacity (i.e., the maximum fixed point of the system, $b = 1 - \mu_1/g_0$), precipitation and evapotranspiration are much larger than

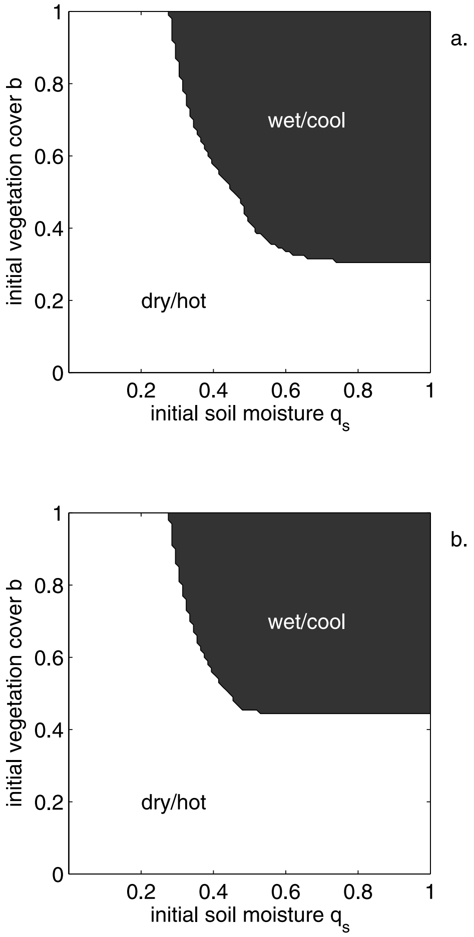

Figure 1. Regime diagram of the soil-vegetation-atmosphere model, for the parameter values mentioned in the text, as a function of the initial condition on q_s (abscissae) and b (ordinates), showing the presence of multiple stable steady states. The dark area represents the wet/cool state, which is reached for high initial values of soil moisture and vegetation cover. The white area represents the dry/hot state, which is reached starting from lower initial values of soil moisture and vegetation cover. Figure 1a is for natural vegetation, and Figure 1b is for cultivated vegetation. For cultivated vegetation, a higher threshold of vegetation cover is needed to reach the wet state.

for the dry state (around 3 mm d^{-1}, see Table 2 for the complete list of values of the prognostic variables). In this model, the absence of vegetation always leads to a dry state, independent of the initial value of soil moisture.

[25] In the case of cultivated vegetation, g and μ are equal to 0, and thus vegetation cover is kept constant at its initial value, because db/dt = 0. In this case, b is considered just as a parameter determining the ratio of vegetated to bare soil surface. This represents the cultivated vegetation as constantly cared for, in such a way that the total cover stays constant in time. Also for cultivated vegetation the atmosphere-soil system displays two stable states similar to those found for natural vegetation. However, in this case larger values of vegetation cover are needed to end up in the wet state (Figure 1b). For any value of b, conditions tend to be drier than in the case with natural vegetation, with higher temperature and lower soil moisture.

[26] The results reported in Figure 1 are obtained for a case where both the evapotranspiration and albedo feedbacks are active. We have also considered a situation where only the evapotranspiration feedback is present, fixing the albedo to a constant value $\alpha = (\alpha_b + \alpha_0)/2$. The results do not display significant changes with respect to those reported in Figure 1, indicating that the albedo feedback plays a secondary role.

[27] The effects of the pure albedo feedback have been studied (not shown) by removing the evapotranspiration difference between bare and vegetated soil [DA; *Laio et al.*, 2001], i.e., $\tilde{E} = E_T(q_s)$. In this case, the albedo feedback can be either positive or negative, depending on the value of the lateral moisture flux. In Figure 1, F_q is fixed at about 0.9 mm d^{-1}. The same behavior can be observed in a range of F_q from about 0.6 to 1.1 mm d^{-1}, as already discussed by DA. When only the albedo feedback is present, one observes a situation similar to that of Figure 1 when F_q is small, in keeping with the classic Charney mechanism. Conversely, when F_q is large, the albedo feedback has a negative effect: for low initial q_s, the system ends up in the wet state when the initial vegetation cover is low, and in the dry state when initial vegetation cover is high. The reason for this behavior is discussed further below.

4. Role of Synoptic Perturbations

[28] As already observed, F_q is an important control parameter of the system, as it represents the role of the synoptic flow. In normal situations, the value of the lateral moisture flux varies in time, ranging from negative (divergence) to positive (convergence) values of a few millimeters per day.

[29] In the following, we simulate synoptic flow variability by stochastically varying the lateral moisture convergence flux. We perform very long integrations of $5 \cdot 10^5$ days, during which F_q assumes a new value every ten days (the time scale of the flow perturbations). For simplicity, F_q has a uniform probability distribution between -1.3 and 3 mm d^{-1}; these values are the same as the ones by DA, where they represented the typical values of the lateral moisture flux at continental midlatitudes in summer. These long integrations do not represent a year-round behavior but rather a sequence of seasons: summers in the continental midlatitude or wet seasons for tropical arid and semi-arid regions. Similar ranges can be chosen for the case of sub-tropical regions during the wet season. The results do not show qualitative dependence on the form of the distribution or on (reasonable) changes in persistence time. Both the albedo and evapotranspiration feedbacks are included.

Table 2. Values Assumed by the Prognostic Variables in the Two Stable States in Case of Natural Vegetation

Variables	"Dry" State	"Wet" State
θ_a (°C)	20.5	13.9
q_a ($\mathrm{kg_{H_2O}/kg_{air}}$)	0.0046	0.0073
T_s (°C)	19.5	13.6
q_s	0.27	0.60
b	0	0.875

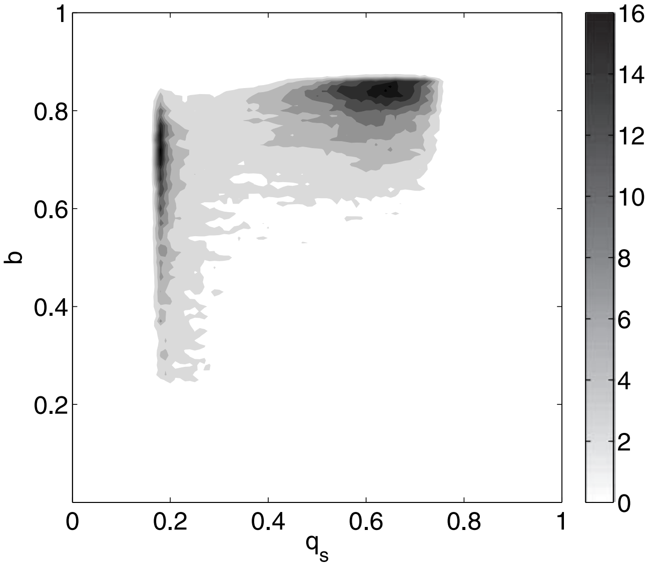

Figure 2. Two-dimensional histogram of soil moisture values (abscissae) and vegetation cover (ordinates) for the case of stochastic synoptic forcing and natural vegetation. The soil moisture distribution is bimodal while vegetation has a unimodal distribution with a maximum close to the system carrying capacity, $1-\mu_1/\gamma_0 = 0.875$. The probability $p(q_s, b)$ is normalized so that $\int p(q_s, b)\mathrm{d}q_s\mathrm{d}b = 1$. Darker colors represent larger probabilities. These results have been obtained with an initial soil moisture value $q_s = 0.5$ and initial vegetation cover values $b = 0.5$; no significant change appears when using different initial conditions on soil moisture or vegetation cover.

[30] With this forcing, the system with natural vegetation displays a bimodal distribution of soil moisture and a single-peaked distribution of vegetation cover, independent of the initial conditions. Figure 2 shows a two-dimensional probability histogram of soil moisture and vegetation cover, obtained from a time series of these two variables. There is no dependence on the initial conditions of the prognostic variables. Together with soil moisture, also temperature, atmospheric moisture, precipitation and evapotranspiration display a bimodal distribution (as in BA). Conversely, the probability histogram of vegetation cover displays a single probability peak just below the carrying capacity, and a tail toward zero. Under dry conditions, the tail stretches out toward lower values of vegetation cover than in the wet state. In this model, natural vegetation does not go extinct, even during dry periods.

[31] As in the case of the constant forcing, we may compare the dynamics of natural vegetation with croplands. The value of b stays constant in time. Figure 3 shows the histograms of soil moisture (on the abscissae) for cultivated plants, for different values of vegetation cover (on the ordinates). To get the same representation shown in Figure 2, we plot in the (q_s, b) plane a series of histograms obtained from time series with different values of vegetation cover. To compare the two figures, the probability $p(q_s, b)$ is normalized so that $\int p(q_s, b)\mathrm{d}q_s\mathrm{d}b = 1$ in both cases. There is no dependence on the initial conditions of the other prognostic variables. Soil moisture (as well as θ_a, q_a and T_s, not shown) displays a bimodal distribution for b larger than about 0.5, with higher probability for the dry than for the wet state. From a comparison with Figure 2, one can see that natural vegetation increases the overall probability of being in a wet/cool state with respect to the case of cultivated plants.

5. Discussion and Conclusions

[32] In this work we have studied a theoretical model of the effects of vegetation on the large-scale hydrologic cycle, focussing on regions where local water recycling is important and vegetation growth is limited by the availability of soil water. Such conditions are often met during the dry season (summer) in continental Europe and the central United States, and during the whole year in semi-arid environments.

[33] This conceptual box model includes the energy balance at the surface and in the Planetary Boundary Layer, the moisture fluxes and an idealized convection scheme. Our approach assumes that vegetation is controlled by the availability of soil moisture and, at the same time, it drives moisture fluxes between soil and atmosphere and affects albedo.

[34] The model system displays two stable states, one "dry" and the other "wet", depending on the initial conditions on soil moisture and vegetation cover. Figure 1, in particular, indicates the importance of vegetation as an active component of the climate system: if the vegetation cover is too small, only the "dry" state is stable, even at high levels of initial soil moisture. This indicates that

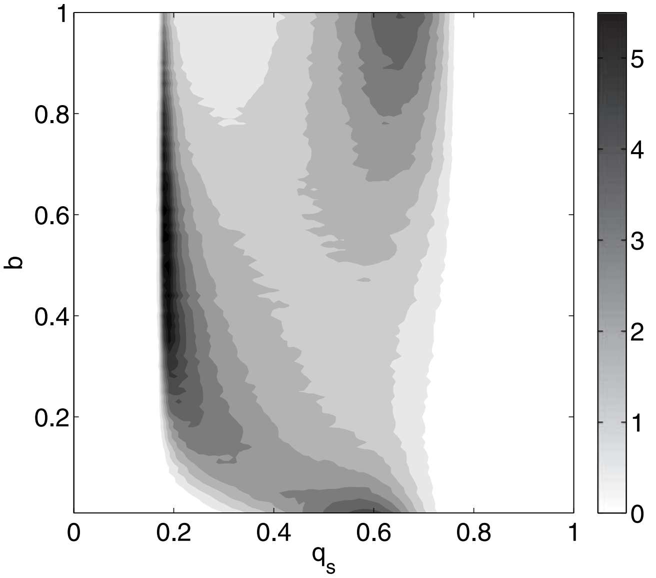

Figure 3. Histograms of soil moisture values (abscissae) for different values of vegetation cover (ordinates), for the case of stochastic synoptic forcing and cultivated vegetation. The soil moisture distribution is bimodal for $b \gtrapprox 0.5$. The probability density of being in a dry state is higher than in the case of natural vegetation, shown in Figure 2. The probability $p(q_s, b)$ is normalized so that $\int p(q_s, b)\mathrm{d}q_s\mathrm{d}b = 1$. Darker colors represent larger probabilities. These results correspond to an initial soil moisture value $q_s = 0.5$; no significant change appears when using different initial conditions on soil moisture.

changes in vegetation cover in arid and semi-arid regions can be of relevance for the occurrence of different weather regimes.

[35] Two different mechanisms are responsible for this effect, namely, the feedbacks associated respectively with evapotranspiration and with albedo. To study the separate effects of each of these two feedbacks, we have performed sensitivity integrations of the model by switching off one of these feedbacks at a time. The evapotranspiration feedback is always positive, and it is due to the fact that transpiration from plants causes an increase of the moisture flux to the atmosphere over vegetated areas. In turn, the larger atmospheric humidity enhances convective instability, causing more precipitation (through more precipitation-efficient convective systems) and favoring vegetation in return, through the increase in soil moisture. Other works showed that the vegetation feedback can be negative, mainly because of its cooling effect over the atmosphere [*Findell and Eltahir*, 2003; *Cook et al.*, 2006]. In our model we did not observe such effect.

[36] In this model, the albedo feedback is always positive at low values of atmospheric moisture inflow. The lower surface albedo due to the presence of vegetation causes an increase of the surface and PBL temperatures, enhancing convection and precipitation and eventually favoring vegetation growth through the increase of soil moisture. Clearly, the effectiveness of this feedback depends on several factors, including the color of the soil and the type of vegetation.

[37] However, the albedo feedback becomes negative when the evapotranspiration feedback is turned off (i.e., the evapotranspiration flux does not depend on vegetation cover) and there is large inflow of atmospheric moisture. The increase of atmospheric temperature leads to a larger evaporative demand. As a consequence, soil moisture is depleted and vegetation cover is reduced. When the atmospheric humidity is large, because of the large lateral moisture flux convergence, convective instability is very high and entirely controlled by moisture convergence. In this situation, an increase in air temperature and in evapotranspiration does not further increase precipitation efficiency. The resulting effect is thus soil moisture depletion, hence the negative feedback. Note, however, that even at large values of moisture influx the negative effect of albedo is completely shadowed when the evapotranspiration feedback is included. In general, we observe that the feedback between vegetation and evapotranspiration is stronger than the vegetation-albedo feedback.

[38] The inclusion of synoptic variability through a stochastic variation of the lateral moisture flux, F_q, leads the PBL-soil system to jump from one stable state to the other, generating a bimodal probability density as observed in soil moisture data. The synoptic forcing actually drives the system through the two states, a typical effect of additive noise acting on a system with multiple stable states. This is observed in model experiments, where the synoptic situation starts the drought phenomena but soil moisture and vegetation conditions enhance and maintain it [*Schär et al.*, 1999; *Hong and Kalnay*, 2000; *Wang and Eltahir*, 2000a; *Zeng et al.*, 1999].

[39] An important point concerns the time scales of natural vegetation reaction. Although the asymptotic vegetation state corresponding to dry/hot climate is $b = 0$, plant dynamics are slow and in our model vegetation cannot die completely in just one dry season. The stable bare state is therefore mainly an indication of what the ecosystem tendency is, more than an observable state. Dry/hot seasons are instead observed because of the shorter time scales of atmospheric dynamics, although interannual memory might also arise from soil moisture feedbacks, as underlined by *D'Odorico and Rodriguez-Iturbe* [2000]. From Figure 2 we also observe, as in BA, the advantage that rainfall intermittency, due to the stochastic variability of F_q, gives to vegetation in comparison to the equilibrium case. Indeed, vegetation never goes extinct, and even when the atmosphere-soil system is in a dry/hot state it remains larger than zero, although displaying larger oscillations than in the wet/cool state.

[40] A comparison of natural vegetation with croplands indicates that, in our simplified model, natural vegetation increases the probability of wet/cool seasons, as compared to cultivated plants. Larger vegetation cover leads to enhanced evapotranspiration, which on one hand increases convection and precipitation but on the other hand depletes the soil of water. Natural vegetation can balance the two effects adapting to lower soil moisture and reducing its surface coverage. The area coverage of cultivated vegetation is instead fixed and it can thus more easily dry out the soil to the point of inducing a drought.

[41] This underlines the importance of land use changes: A transition from natural to cultivated vegetation may cause the system to be more easily prone to the occurrence of dry conditions and seasonal droughts. Conversion of natural forested areas to pasture and croplands has been extensively investigated considering the effects of differences in root depth, roughness length and albedo changes [see, e.g., *Bonan*, 1997; *Hahmann and Dickinson*, 1997; *Eastman et al.*, 2001]. Our study instead considers the differences in the adaptation of areal coverage between natural and cultivated vegetation. The importance of agricultural lands as an active agent of the climate system and not only as a passive component has been acknowledged [*Pielke et al.*, 2007] but the specific mechanisms are not yet well understood and more effort is needed [*Osborne et al.*, 2007]. The effect of land use changes on summer convective rainfall was discussed by *Juang et al.* [2007a].

[42] The model introduced here is highly simplified and it does provide just a sketch of the complex biological and physical processes active in the soil-vegetation-atmosphere system, as already discussed by DA and by BA. Evaporation from bare soil is overestimated, as in BA, since it takes place from the whole root depth instead of drying up only a thin surface layer. Some important mechanisms are not included, such as the facilitation of water infiltration in vegetated soils as compared to bare areas, which was shown to be important for the strength of vegetation feedbacks in dry lands [*Dekker et al.*, 2007; *Janssen et al.*, 2008]. This mechanism cannot be easily taken into account since in our model soil moisture has the same average value in both vegetated and bare sites. However, it has been shown that the error introduced by not including the microscale feedbacks is partly compensated by considering temporal variability in rainfall [*Baudena and Provenzale*, 2008]. We lump the soil and vegetation in an only layer, thus considering evapotranspiration as well as an average temperature in the vegetated "soil" that includes also the vegetation,

differently from other works where the two layers were separated [e.g., *Margulis and Entekhabi*, 2001]. The nonlinear form of precipitation efficiency, which makes the DA approach different from other models previously introduced [*Brubaker and Entekhabi*, 1995; *Entekhabi and Brubaker*, 1995], is still an important but poorly known ingredient of the system (see, e.g., the discussion by DA or by *Market et al.* [2003] and references therein). In croplands, irrigation effects are not considered.

[43] Despite all these limits, we believe that the simple model discussed here provides useful insight in the mechanisms at work in the soil-vegetation-atmosphere system, and it leads to realistic values of temperature and humidity in the two stable states. The results obtained with this conceptual model agree with previous studies based on observations and numerical simulations with full-fledged climate models. For the study of the evapotranspiration and albedo feedbacks, see, e.g., *Pielke et al.* [1998], *Moorcroft* [2003], and *Scheffer et al.* [2005]. For the existence of alternative stable states in arid and semi-arid regions, associated with abrupt shifts between desert and vegetated conditions, see, e.g., *Claussen et al.* [1999], *Kleidon et al.* [2000], *Scheffer et al.* [2001], and *Wang and Eltahir* [2000b, 2000a]. The greater importance of evapotranspiration with respect to albedo was also observed in GCM studies [e.g., *Xue*, 1996; *Crucifix et al.*, 2005]. One advantage of the conceptual model discussed here is its simplicity that allows for exploring in detail the different processes and feedbacks at work in the system.

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