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Indirect vegetation–soil moisture feedback with application to Holocene North Africa climate[1]

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Abstract

Using a fully coupled climate–terrestrial ecosystem model, we demonstrate explicitly that an initial perturbation on vegetation induces not only a direct positive vegetation feedback, but also a significant indirect vegetation–soil moisture feedback. The indirect feedback is generated through either fractional cover change or soil moisture depletion. Both indirect feedback mechanisms are triggered by a vegetation perturbation, but involve subsequent effects of soil moisture and evaporation, indirectly. An increase in vegetation tends to reduce bare-ground evaporation through either the area reduction in bare ground or the depletion of soil moisture; the reduced evaporation may then counter the initial plant transpiration, favoring a negative net vegetation feedback. Furthermore, grasses are more effective in inducing the indirect vegetation–soil feedbacks, because of their limited plant evapotranspiration and shallower roots that tend to change surface soil moisture, and, in turn, evaporation, effectively. In comparison, trees favor a direct positive vegetation feedback due to their strong plant transpiration on subsurface soil moisture as well as a lower albedo.

Keywords: indirect feedback, soil moisture feedback, vegetation feedback

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Introduction

The complex interaction between the terrestrial ecosystem and climate remains poorly understood. A notable example is the arid climate in the subtropical region such as the North Africa. It has long been proposed that a change in vegetation can change rainfall through the changes in surface albedo and plant evapotranspiration (T) via a positive vegetation feedback (Charney, 1975; Charney *et al.*, 1977; Kutzbach *et al.*, 1996; Claussen 1997; Claussen *et al.*, 2004), which will now be called the direct vegetation feedback. However, the final impact of the vegetation change depends not only on this direct vegetation feedback, but also on the subsequent impact associated with soil moisture change. A change in vegetation is accompanied naturally by a change in soil moisture, which can then change surface albedo and bare-ground evaporation, and, eventually, rainfall through the soil moisture feedback (Charney *et al.*, 1977; Shukla & Mintz, 1982; Delworth & Manabe, 1989; Koster *et al.*, 2004). This subsequent soil moisture feedback that is triggered by an initial vegetation perturbation will be called the indirect vegetation–soil moisture feedback or simply the indirect vegetation feedback here. The final vegetation feedback is the net effect of the direct and indirect feedbacks and involves the three-way interaction among vegetation, soil and climate.

Most studies so far, however, have studied the direct vegetation feedback and soil moisture feedback separately. Here, we will examine the direct vegetation feedback, the indirect vegetation feedback and the associated vegetation–soil moisture interaction explicitly. Our study is motivated by a recent attempt to understand the role of vegetation feedback on North Africa climate in the Holocene in a fully coupled climate–ecosystem model and in the observation (Liu *et al.*, 2007): contrary to most previous hypotheses, the vegetation in our model appears to exert a negative feedback on annual rainfall in the mid-Holocene (Notaro *et al.*, 2008; N08 hereafter). Here, we performed specifically designed sensitivity experiments to understand the vegetation feedback and the associated vegetation–soil moisture interaction. It is found that the relative contribution of the direct and indirect vegetation feedbacks to the net vegetation feedback depends on the vegetation type. An expansion of trees tends to

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increase rainfall, because it induces primarily a direct positive vegetation feedback through a strong plant transpiration of subsurface soil moisture as well as a lower albedo. In contrast, an expansion of grass could reduce rainfall and favors a negative net vegetation feedback, because it triggers a reduction in surface evaporation, and in turn a subsequent soil moisture feedback, or the indirect vegetation feedback.

Experimental design

We will use a fully coupled climate–global dynamic vegetation model, Fast Ocean Atmosphere Model/ Lund-Potsdam-Jena (FOAM–LPJ) (Gallimore *et al*., 2005). The background climate is 'snapshot' simulation of a 400-year control (CTRL) that is forced by the orbital forcing of 6000 years ago. The simulated climate in the mid-Holocene exhibits an enhanced summer monsoon and a northward expansion of vegetation over northern Africa relative to the present (Liu *et al*., 2007). Vegetation cover extends northward up to 25°N (Fig. 1). Sensitivity experiments are performed with initial perturbation on the vegetation over the arid grassland region of northern Africa (18°–23°N, 11°–36°E) (the northern region marked in Fig. 1). The model climatology in this region is characterized by a summer monsoon from midsummer into the fall, with a peak rainfall of about 160 mm month^{-1} in August (Fig. 2c). The fractional cover of vegetation is about 0.6, comprised of ~ 0.5 cover in grasses and ~ 0.1 in temperate evergreen

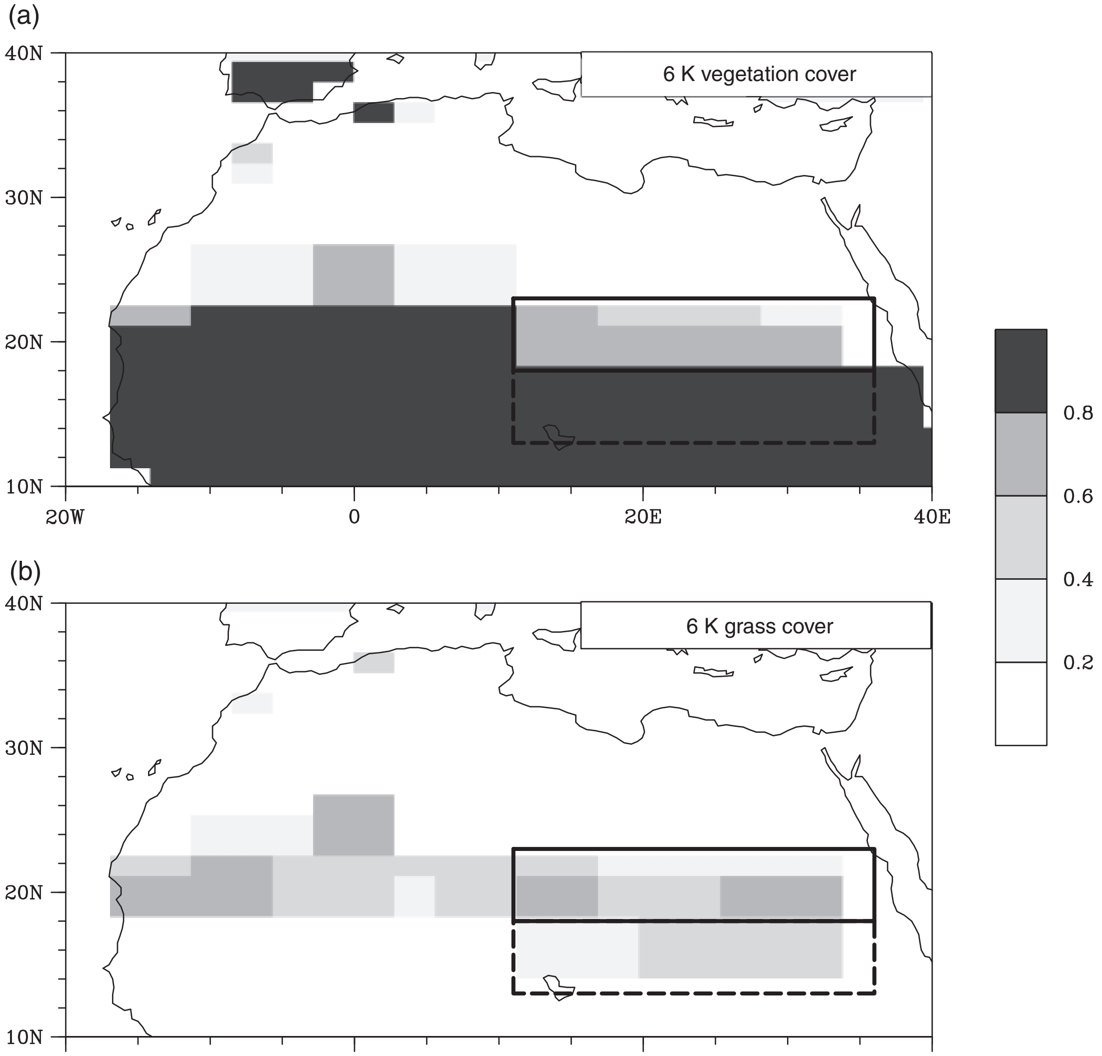

Fig. 1 Fractional coverage of (left) total vegetation and (right) grass simulated in FOAM–LPJ in the mid-Holocene. The two boxes marked show the northern and southern regions discussed in the text. Vegetation perturbation is applied to the northern region in the sensitivity experiments (adopted from Liu *et al*., 2007).

trees, both needleleaf and broadleaf. The leaf area index (LAI) increases in response to the monsoon rainfall in the rainy season, peaking in September (Fig. 2a). The grass evapotranspiration (T) (in this version of LPJ, the plant evapotranspiration T is not separated into the individual components of plant transpiration and canopy evaporation) and bare-ground evaporation (E) largely follow the monsoon rainfall, peaking in later summer (Fig. 2b). Soil water is calculated with a two-layer model: a top layer of 0.5 m depth with a field capacity of 75 mm, and a deep layer of 1 m depth with a field capacity of 150 mm (Sitch *et al.*, 2003). Because of its strong hydrological exchange with the lower atmosphere, surface soil moisture exhibits a clear seasonal cycle with the maximum lagging rainfall by 1 month and with the water content (in percentage of water-holding capacity) in the top layer increasing from <0.1 in the dry season to a peak of 0.6 in September (Fig. 2d). In comparison, the subsurface soil moisture exhibits a much subdued seasonal cycle that lags rainfall by $\sim$ 1–2 months. Finally, trees in our model have 70% roots in the top layer and 30% in the lower layer, while grasses have 90% roots in the top layer and 10% in the lower layer. The deeper roots of trees enable them to transpire water more effectively from the deep soil than grass. The different rooting depths, as will be seen later, lead to significantly different vegetation–soil moisture feedbacks.

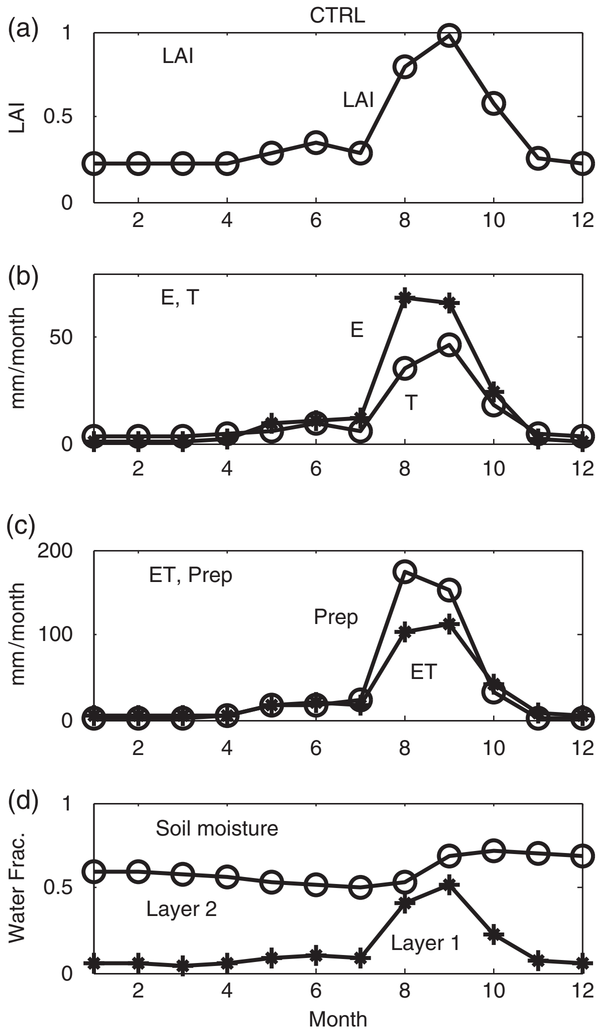

Fig. 2 Seasonal climatology of the CTRL in the North Africa (northern region, 18°–23°N, 11°–36°E). (a) Leaf area index, (b) bare-ground evaporation E and plant transpiration T (in mm month^{-1}), (c) total evapotranspiration (ET) and precipitation (Prep) (in mm month^{-1}) and (d) soil wetness (in percentage of water-holding capacity) in the top layer and deep layer. All values are taken from a 100-year section of the mid-Holocene CTRL simulation in FOAM–LPJ.

To examine the roles of direct and indirect vegetation feedback more clearly, we first perform two sets of experiments with perturbations in the fractional coverage of vegetation type, one for tree (TREE) and the other for grass (NOGRASS). As will be seen, trees and grasses have different feedbacks from soil moisture because of their different biophysical characteristics such as the rooting depth, seasonal phenology, evapotranspiration and albedo. Each set of experiments consists of 100 ensemble members, with each member starting from January 1 of a different year of the CTRL. In the TREE experiment, tree cover is increased at the expense of grass over the northern region. For each ensemble member, the grass cover ($\sim$ 0.5) is replaced by temperate evergreen trees, the same type as in the control, at the beginning of the year and the coupled model is integrated for the entire year. In LPJ (Sitch *et al.*, 2003), vegetation cover is updated at the end of each year, while leaf phenology changes daily in response to climate forcing throughout the year. Therefore, leaf phenology evolves during the year, even though the fractional coverage remains unchanged. The NOGRASS experiment, as described in (N08), is similar to TREE, except that the grass cover ($\sim$ 0.5) is replaced by bare ground. In the following, the response to the tree expansion (over grass) will be derived as the ensemble mean difference TREE$-$CTRL, while the response to a grass expansion (over bare ground) will be described as CTRL$-$NOGRASS. It should be pointed out that the two experiments are not exactly comparable. A more comparable tree experiment with NOGRASS experiment could have trees replace (or replaced by), bare ground. However, this is very unrealistic because trees are usually compensated by grasses in response to rainfall variability. Therefore, the comparison of the two experiments here should be taken qualitatively, rather than quantitatively.

Table 1 Annual mean response of LAI, bare-ground evaporation (E, mm month^{-1}), plant evapotranspiration (T, mm month^{-1}), total evapotranspiration (ET = E + T, mm month^{-1}), precipitation (P, mm month^{-1}), and soil water fraction in the top layer (Layer-1) and deep layer (Layer-2) in the CTRL and the differences between three sensitivity experiments and CTRL

Exp\variable	LAI	E	T	ET	Prep	Layer-1	Layer-2
CTRL	0.37	16	12	28	35	0.15	0.60
	(1.7)	(7.2)	(66)	(73)	(139)	(0.39)	(0.68)
TREE–CTRL	**1.0**	−0.25	**5.0**	**4.7**	**1.7**	−0.02	**−0.22**
	(0.2)	(0.47)	**(2.3)**	**(2.8)**	**(6.3)**	(0.01)	(0.01)
CTRL–NOGRASS	**0.14**	**−9.2**	**5.4**	**−3.8**	**−4.7**	−0.00	**−0.02**
LEAF–CTRL	**0.03**	**−1.6**	−0.2	**−1.8**	**−3.9**	−0.01	**−0.02**

The values are averaged over North Africa in the northern region (18–23°N, 11–33°E). For CTRL and TREE–CTRL, the nonlocal response in the neighboring southern region (13–18°N, 11–33°E) is also shown in bracket (nonlocal responses are weak for NOGRASS and LEAF experiments and are therefore not shown). (Bold numbers are beyond 1 ensemble standard deviation).
LAI, leaf area index.

Tree–soil moisture interaction

The mean annual response in the TREE experiment (Table 1) shows an increased LAI and T, as well as a modest increase in local rainfall (Fig. 3a), reflecting a weak positive feedback. The seasonal evolution shows that an initial replacement of grass, which is inactive in the dry season, by evergreen tree leads to a jump in LAI (Fig. 4a) and, in turn, T (Fig. 4b) at the beginning of the year. Bare-ground evaporation changes little (Fig. 4b) such that the sum of the anomalous T and bare-ground evaporation (E), or the total evapotranspiration (ET), follows that of T (Fig. 4c). The initial increase of T reduces soil moisture significantly in the lower layer (Fig. 4d), because trees have substantial deeper roots (70% top layer and 30% lower layer), and, thus, transpire water effectively from the deep soil moisture pool; in addition, evergreen trees transpire year round. In contrast to T, local precipitation is not increased in the dry season (Fig. 4c), because the background descending flow is unfavorable for convective precipitation. After the spring transition season, in May (Fig. 2c), the increased moisture supply from the enhanced ET destabilizes the atmosphere column, leading to an enhanced summer rainfall. The rainfall increase is also likely driven by a lower albedo (~ 0.055) caused by replacing grasses (~ 0.22) with evergreen trees (~ 0.14), making more surface energy available for driving the latent and sensible heat fluxes. The anomalous rainfall recharges the soil such that the soil moisture stops decreasing (Fig. 2d). Later in the summer (August and September), T, and in turn, precipitation, declines relative to CTRL, partly because of the enhanced grass T in the rainy season in the CTRL. The grass LAI increases dramatically in the rainy season in CTRL when grass leaves bloom as soil water exceeds drought limits (Fig. 2a), while the LAI of the evergreen trees remains unchanged throughout the year. Therefore, the change between TREE and CTRL is smaller in LAI, and in turn, T, ET and precipitation in the rainy season (Fig. 4a–c) – a point to be returned later in 'Grass–Soil Moisture Feedback.'

The TREE experiment also exhibits a significant increase of nonlocal rainfall in the neighboring southern region (13°–18°N, 11°–36°E, marked in Fig. 1) (Table 1) (Fig. 3a) (the annual rainfall increase relative to the CTRL climatology is, however, ~ 5% in the southern region, comparable with that locally over the region of vegetation change (Table 1), because the southern region has a much wetter climatology (Fig. 5b and c; Table 1), where is covered primarily by trees with a much larger LAI and wetter climate than in the northern region (Fig. 5a–d). The rainfall increases significantly from May to August, with a peak in June of over 30 mm month^{-1} (Fig. 5g). The increased rainfall also leads to a modest increase in soil moisture in the rainy seasons, and in turn an increase in T (Fig. 5f) and ET (Fig. 5g) that almost parallels the rainfall increase. The increased rainfall also leads to a small increase in soil moisture and vegetation (Fig. 5e and h, note the different scales from Fig. 4a and d). Because no vegetation change was imposed initially in the southern region, the rainfall enhancement there is associated with nonlocal atmospheric dynamics and moisture transport. The latter can be inferred from a larger rainfall anomaly than local moisture supply, with the rainfall anomaly about twice that in T (Fig. 5g). Relative to the northern region where the vegetation change is imposed, the southern region is located closer to the ITCZ, with a stronger ascent and greater moisture in the background climatology (Fig. 5c and d); the southern region also receives substantial moisture transport from regions further south. These conditions likely lead to a stronger sensitivity of precipitation to moisture increase there.

Overall, tree expansion induces a significant overall positive feedback on rainfall. The net vegetation feed-

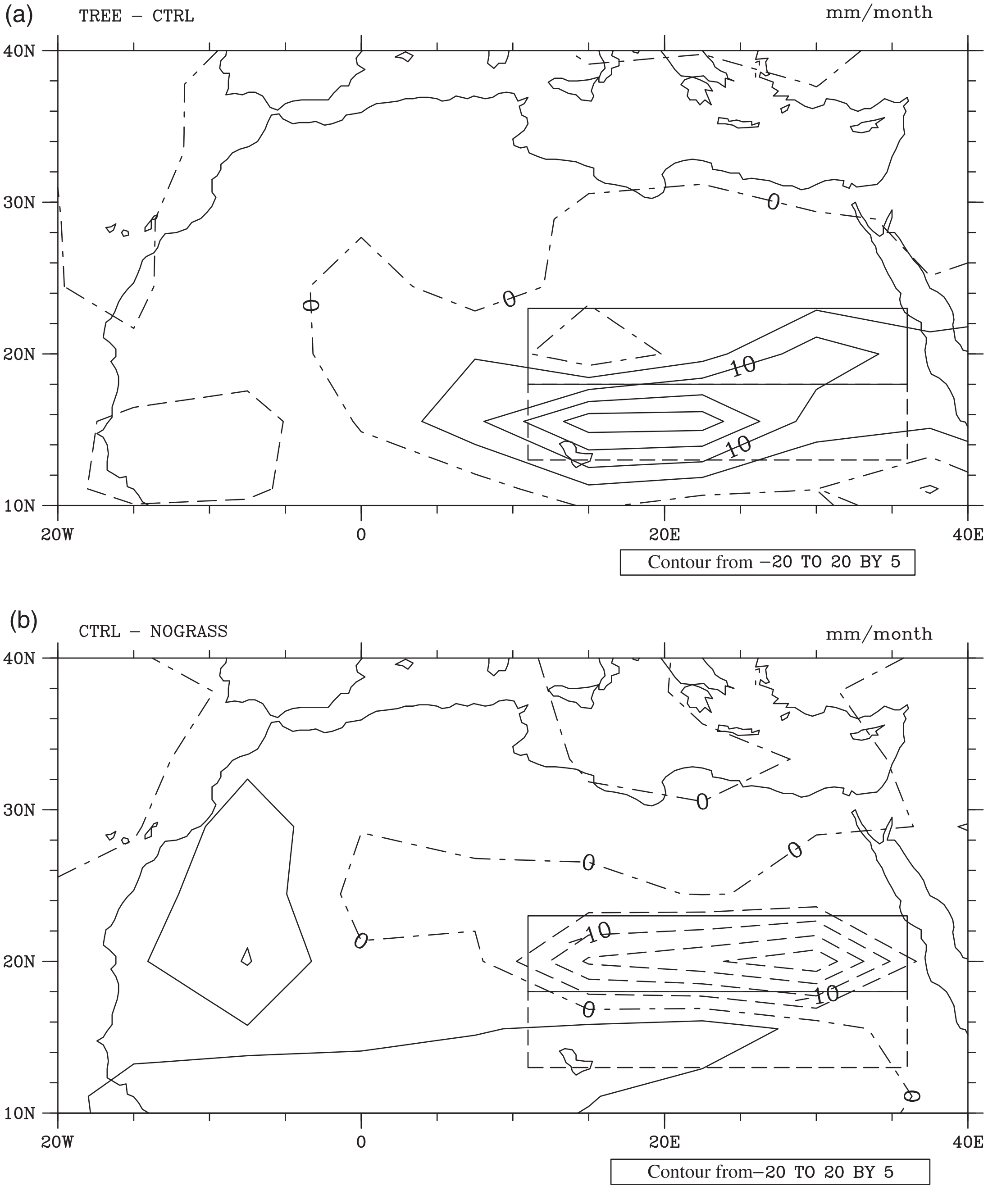

Fig. 3 Map of simulated summer (MJJAS) rainfall responses (mm month^{-1}) in (a) TREE–CTRL and (b) CTRL–NOGRASSF. The northern and southern regions are also marked as in Fig. 1. In TREE–CTRL, precipitation is increased modestly over the northern region locally, and significantly in the remote southern region. In CTRL–NOGRASS, precipitation is reduced significantly in the northern region locally, without significant nonlocal impact (positive solid, negative dash and contour interval 5 mm month^{-1}).

back is caused by the direct vegetation feedback through the plant transpiration of subsurface soil moisture, as well as a lower plant albedo. The indirect vegetation–soil moisture feedback induced by the tree perturbation is of secondary importance, because trees transpires significant amounts of water from the deep soil, and therefore, does not trigger significant reduction in surface moisture (Fig. 4d) and, in turn, evaporation (Fig. 4b).

Grass–soil moisture feedback

In contrast to the TREE experiment, the grass expansion inferred by CTRL–NOGRASS leads to a net negative

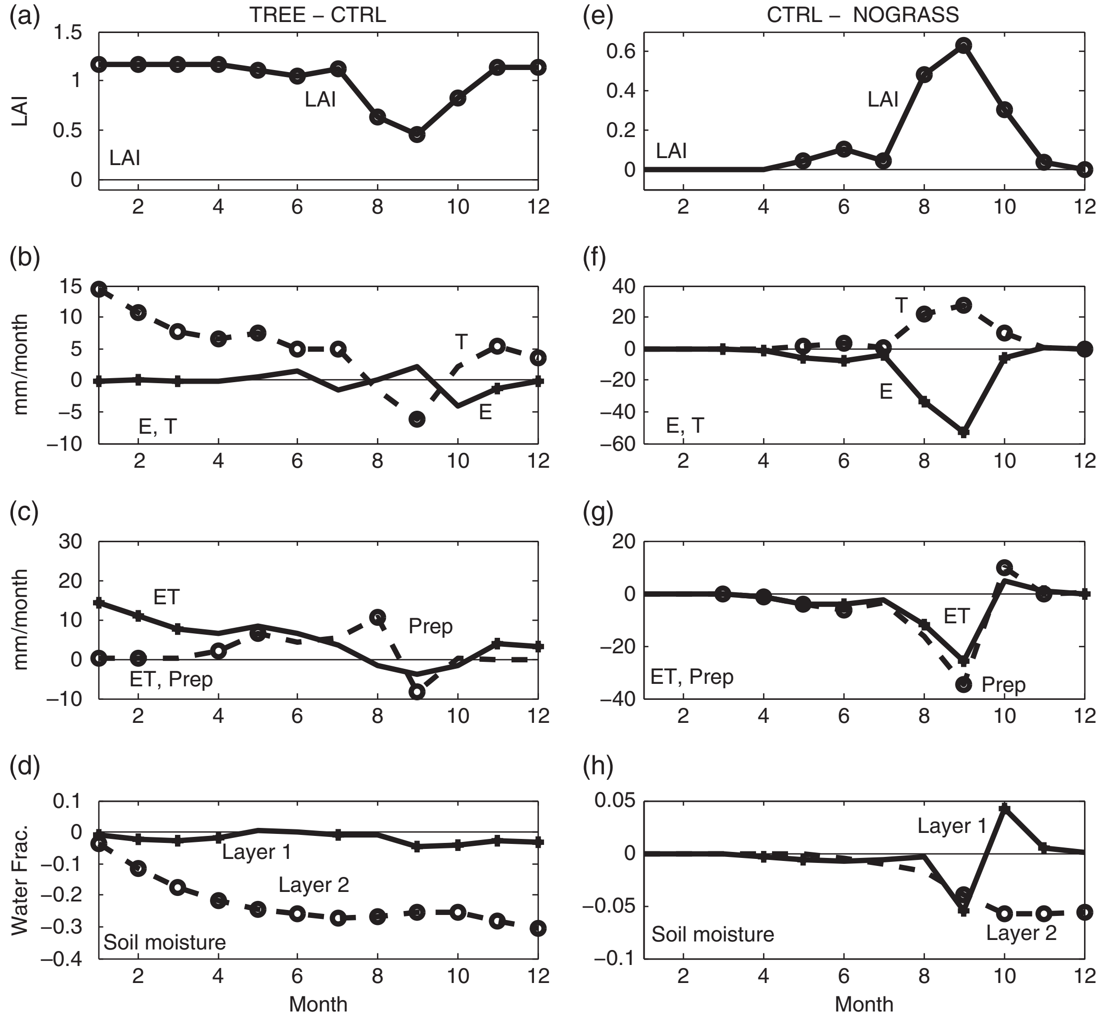

Fig. 4 Seasonal evolution of the climate–vegetation response in TREE and NOGRASS experiments. (a)–(d) Same as in Fig. 1a–d, but for the difference of ensemble means of TREE–CTRL: (a) leaf area index (LAI), (b) evaporation E and T (in mm month^{-1}), (c) ET and precipitation (Prep) (in mm month^{-1}) and (d) soil wetness in the top layer (layer 1) and deep layer (layer 2). (e)–(h) Same as (a)–(d) but for the difference of CTRL–NOGRASS. All the values are averaged in the northern region where the vegetation is changed initially. Values beyond 1 ensemble standard deviation are marked with circles/asterisks.

vegetation feedback, with a reduction in annual rainfall locally (Table 1) (Fig. 3b). This negative feedback appears to be associated with a large reduction in bare-ground evaporation that overwhelms the increase in T, in response to grass expansion (Table 1). As shown in the seasonal evolution, the grass expansion increases the LAI (Fig. 4e) and T (Fig. 4f) in the growing season significantly, similar to the case of tree perturbation. However, the increase in T is accompanied by an even stronger reduction in bare-ground evaporation (Fig. 4f) such that the ET is reduced (Fig. 4g). The reduced ET then decreases local monsoon rainfall from May to October through primarily a positive soil moisture feedback (Fig. 4g). This complex chain of interactions of vegetation and soil moisture eventually leads to a negative net vegetation feedback on rainfall. This negative grass feedback in North Africa during the mid-Holocene has been studied extensively in FOAM–LPJ (N08). This negative feedback is not caused by the surface albedo effect, because model grasses have an average albedo comparable with the bare-ground soil such that the albedo change is <0.015 (in the model, bare-ground average albedo increases linearly with top soil water content, from 0.135 for wet soil to 0.27 for dry soil. In the rainy season, top soil has a wetness (in water-holding capacity) of ~ 0.5 (Fig. 1d), and therefore an albedo of ~ 0.19, very close to the average albedo of ~ 0.22 for model grasses).

N08 proposed that the negative grass feedback is caused by the change in the area extent of bare ground.

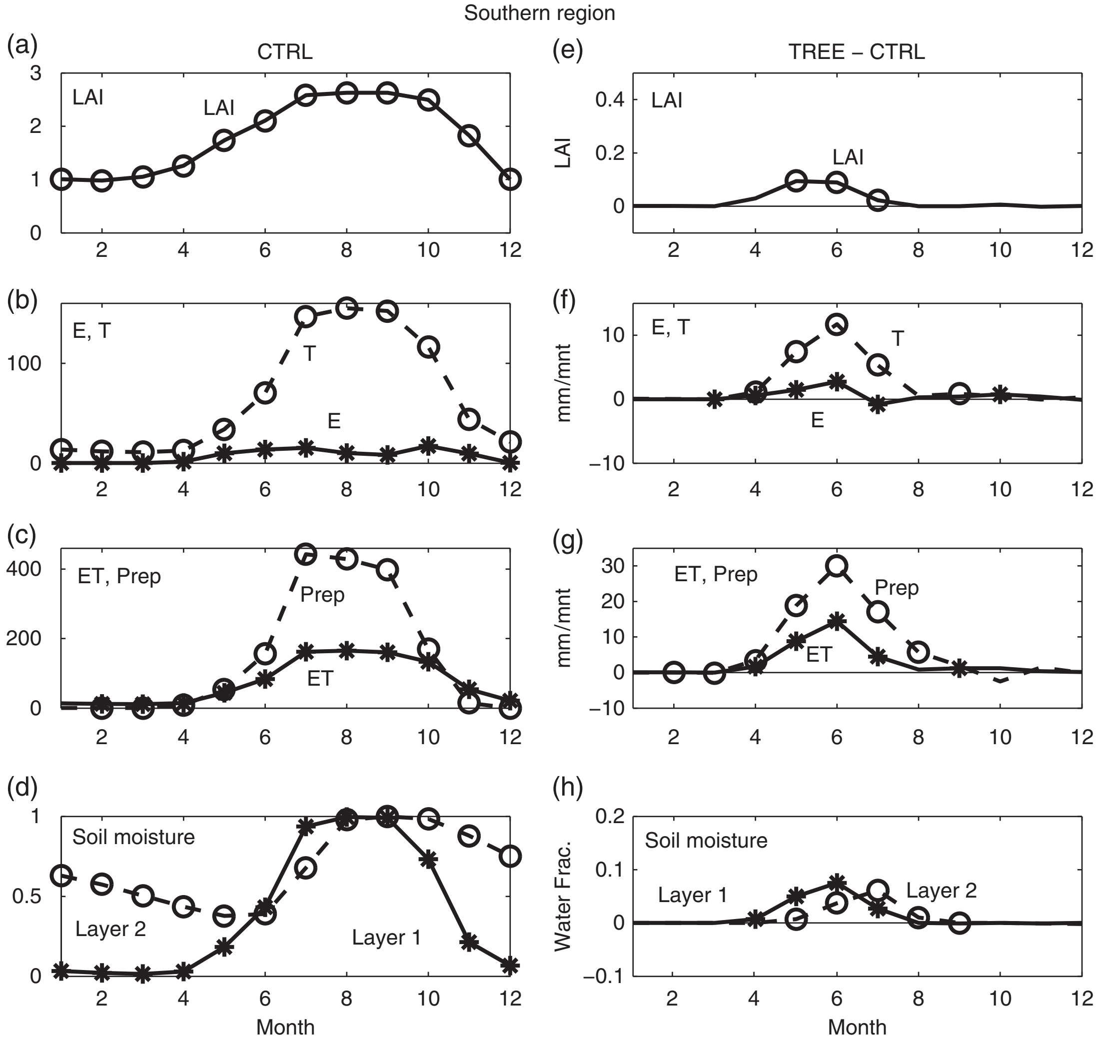

Fig. 5 Seasonal evolution in the southern neighboring region (13°–18°N, 11°–36°E) for (left) climatology of the CTRL and (right) anomalous response derived as TREE–CTRL. (a) Leaf area index (LAI), (b) E, T, (c) ET and precipitation, (d) soil moistures for the climatology of CTRL the same as Fig. 4a–d except for the southern neighboring region (13°–18°N, 11°–36°E). (e) LAI, (b) E, T, (c) ET and precipitation, (d) soil moisture derived as TREE–CTRL as (e)–(h), but for the southern neighboring region.

The expansion of grass, while increasing T, reduces the fractional coverage of bare ground and, in turn, bare-ground evaporation. Especially, in the rainy season when the climatological surface soil moisture is high (Fig. 1d), an anomalous decrease of soil moisture can induce a large evaporation decrease that overwhelms the anomalous increase of grass T. The large change in evaporation over T may be inferred from their climatology, at maximum in August/September, the evaporation is close to 70 mm month^{-1}, whereas the transpiration is $<$50 mm month^{-1} (Fig. 2b). This type of condition with a larger change in bare-ground evaporation than in grass transpiration has been reported in field experiments on pasture land, for example, in Australia (Murphy & Lodge, 2001), as well as in some modeling studies over North Africa (Penning *et al.*, 1975; Lawrence *et al.*, 2007). The negative grass feedback on Africa monsoon rainfall in the mid-Holocene is also present in CCSM2-CLM (N08) and another version of FOAM–LPJ in which the bare-soil evaporation is reduced with a parameterization of the skin soil crust effect (not shown). This grass–soil interaction associated with the relative change of coverage will be called the vegetation–soil feedback of fractional coverage. This feedback appears to contribute significantly to the reduced rainfall in the experiment with grass perturbation.

Here, we further propose that the reduced rainfall in response to grass expansion is also caused by another type of vegetation–soil interaction associated with soil moisture depletion. Even through the fractional vegetation coverage remains unchanged, an enhanced T from growing grass can deplete the surface soil moisture, through the dominant shallow roots (90% top layer, 10% lower layer), which in turn reduces the bare-soil evaporation, countering the effect of increased T on precipitation. The reduced soil moisture also increases surface albedo, which reduces the surface energy available for driving heat fluxes, also unfavorable for rainfall (Levis *et al*., 2004). This suggests potentially a negative indirect vegetation feedback induced indirectly by the soil moisture depletion.

To isolate the role of soil moisture depletion from that of fractional coverage, we performed another set of ensemble experiments (LEAF), in which the grass leaf phenology is increased artificially and held fixed throughout July (from ~ 0.1 to ~ 0.6) (we increase the LAI to a maximum of 0.6 in each member. However, since the LAI is different in each member, the net increase of LAI is different in each member. As a result, the ensemble mean LAI increase in Fig. 6a is slightly <0.6) while the grass fractional coverage remains unchanged. The response to the increase of grass leaf phenology is described as the difference between the ensemble mean difference LEAF−CTRL. In spite of the increase in LAI, annual rainfall decreases (Table 1), suggesting a negative net vegetation feedback. Initially in July, the enhanced leaf phenology increases the LAI by ~ 0.6 (Fig. 6a), which leads to a small increase of T by $\sim 6\,\mathrm{mm\,month^{-1}}$ (Fig. 6b). However, the increased T reduces the soil moisture in the top layer (Fig. 6d), and, in turn, the bare-ground evaporation by $\sim 3\,\mathrm{mm\,month^{-1}}$ (Fig. 6b). As a result, ET increases by $\sim 6–3 = 3\,\mathrm{mm\,month^{-1}}$ (Fig. 6c). This initial increase of T is small, however, and does not seem to induce any significant atmospheric dynamic responses, including rainfall, in July (Fig. 6c). After July, this artificial increase in LAI dies off rapidly such that the LAI anomaly virtually vanishes in August. This LAI die-off occurred because the local rainfall climatology is too low to sustain grasses of such a large LAI. One might then expect the disappearance of any significant response in the coupled system after July. Surprisingly, in the subsequent months, however, local rainfall is reduced significantly by up to $30\,\mathrm{mm\,month^{-1}}$ (Fig. 6c), accompanied by a large reduction in bare-ground evaporation (Fig. 6b) and ET (Fig. 6c). Thus, an increase in LAI eventually leads to a significant reduction in annual monsoon rainfall (Table 1), forming a negative net feedback that is reminiscent of that inferred from CTRL–NOGRASS.

The negative vegetation feedback in LEAF, we propose, is caused by the soil moisture depletion that is triggered indirectly by the initial increase of plant transpiration. Indeed, the initial increase of T reduces soil moisture in both layers in July (Fig. 6d). The reduced surface soil moisture is further sustained and enhanced in the following months, partly by a greater infiltration into the drier deep soil. The reduced surface soil moisture decreases bare-ground evaporation (Fig. 6b) initially in July, countering the increased T. The depletion of surface soil moisture into the following months further leads to a reduction of evaporation, T (Fig. 6b), ET, and, in turn, local rainfall (Fig. 6c). The reduced rainfall further reduces evaporation and soil moisture, forming a negative indirect grass feedback due to soil moisture depletion.

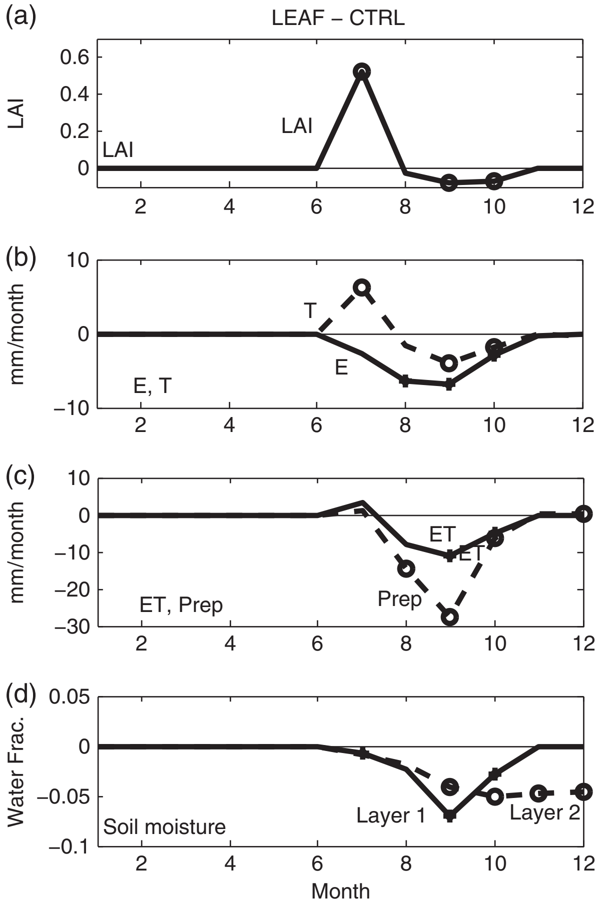

Fig. 6 Seasonal evolution of the climate–vegetation response due to leaf phenology perturbation (LEAF–CTRL) in the northern region (18°–23°N, 11°–36°E). (a)–(d) Same as Fig. 4e–h, respectively, but for LEAF–CTRL.

Further experiments show a preferred seasonality for this negative grass feedback associated with soil moisture depletion. In all sensitivity experiments with enhanced LAI in July or August, rainfall is reduced similarly as in LEAF, with the maximum reduction in September. In contrast, for experiments with LAI enhanced after September, the initial perturbation in LAI and the induced anomaly in the climate–ecosystem, die off completely in the subsequent months. Therefore, this indirect feedback is induced by vegetation perturbation in the growing season (July and August), when vegetation growth can deplete soil moisture most effectively.

Here, we discussed two negative indirect vegetation feedbacks associated with soil moisture, one due to fractional cover and the other due to moisture depletion. These indirect feedbacks can help explaining the opposite responses between TREE–CTRL and CTRL–NOGRASS. Indeed, neither of the negative indirect feedbacks is effective in the TREE experiment. The feedback due to fractional cover is absent in TREE because grass is replaced by tree and the fractional coverage of both total vegetation and bare ground remains unchanged. Furthermore, and more importantly, evergreen trees are characterized by deeper roots than grasses, depleting soil moisture much more effectively in the deep soil year round (Fig. 2d). The soil moisture depletion in the deep layer alone is ineffective in forming a negative indirect feedback because it cannot reduce the surface soil moisture, and, in turn, evaporation, sufficiently fast to counter the increase in T (Fig. 2b). In contrast, grasses are dominated by shallow roots and therefore grass transpiration can reduce surface soil moisture, and, in turn, evaporation, effectively to counter the increase of T, and, in turn, rainfall. Finally, both types of indirect vegetation feedbacks are less effective for trees than for grasses because trees have a large leaf area and in turn a large potential T, which tends to overwhelm the soil moisture effect. In other words, trees tend to be dominated by the direct positive vegetation feedback through T, while grasses can be favorable for the indirect negative vegetation feedbacks through E.

Finally, unlike the TREE experiment (Fig. 3a), nonlocal response in NOGRASS (Fig. 3b) and LEAF experiments are weak. This could be related to different vegetation characteristics such as surface albedo, surface roughness and T, as well as different atmospheric climatology, such as the mean ascent motion, mean moisture transport and mean lower atmospheric stability. A more clear explanation, however, remains to be developed.

Summary

Using a fully coupled climate–vegetation model, we demonstrate that a vegetation perturbation induces not only a direct positive vegetation feedback on climate, but also a significant indirect negative vegetation–soil moisture feedback that is associated with both fractional vegetation cover change and soil moisture depletion (Fig. 7). The indirect vegetation feedbacks are triggered by a vegetation perturbation, but involve subsequent effects of soil moisture and evaporation, indirectly. An increase in vegetation tends to reduce bare-ground evaporation through either the reduction of bare-ground area or the depletion of soil moisture; the reduced evaporation counters the initial transpiration effect of the direct vegetation feedback, acting as a negative indirect vegetation feedback (see Fig. 7 for a schematic figure of the direct and indirect feedbacks). Furthermore, grasses are more effective in inducing the

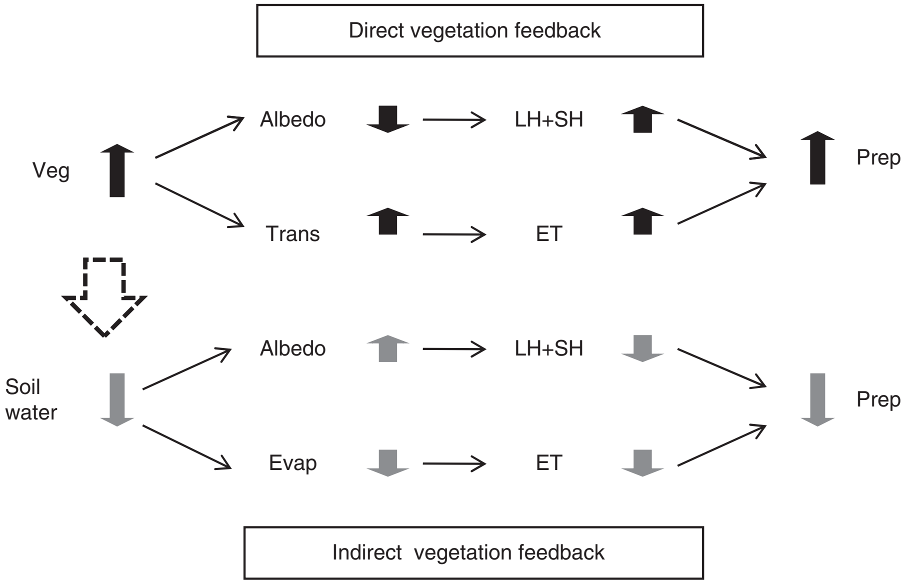

Fig. 7 Schematic diagram for the direct (upper) and indirect (lower) vegetation feedbacks in response to a vegetation perturbation.

indirect vegetation feedbacks, because of their limited T, as well as the dominant shallow roots, which tend to change surface soil moisture, and, therefore, evaporation, effectively. In comparison, trees tend to favor the direct positive vegetation feedback due to their strong plant transpiration on subsurface soil moisture as well as low albedo. As such, trees tend to produce a net positive vegetation feedback because of the dominant direct vegetation feedback, while grasses tend to favor a net negative vegetation feedback, because of their effective excitation of the indirect vegetation feedback.

So far, the two indirect vegetation feedbacks are studied explicitly only in FOAM–LPJ. It remains to be further studied how these feedbacks apply to the real world. We speculate that our results should remain qualitatively valid in other models and, perhaps, in the real world, although, quantitatively, the details may differ depending on models. That is, relatively, trees tend to favor the direct feedback, and, in turn, a positive net vegetation feedback, while grasses tend to favor the indirect feedback, and, in turn, less positive net vegetation feedback. Partly, this is because FOAM–LPJ is comparable with other current generation of complex terrestrial-climate models. More importantly, our speculation is based on the seemingly robust physical mechanisms involved. For example, trees have more deep roots than grasses, and therefore tend to increase transpiration through the depletion of soil moisture more in the deeper soil, favoring the direct vegetation feedback. Even if some plant characteristics differ significantly from those in our model, one may still expect a vegetation–soil feedback that can be understood from our study here. For example, if the grass T overwhelms bare-ground evaporation, the grass–soil feedback due to fractional cover change may become positive. Nevertheless, it is unlikely this feedback can exceed the positive feedback with trees, because trees tend to have a larger potential T than grasses. This seems to be the case in our model in the late Holocene, according to a statistical analysis of vegetation feedback (N08). In the late Holocene, the mean monsoon rainfall is reduced in the North Africa, and therefore the climatological E is not much stronger than T.

Overall, our study implies that vegetation feedback on climate is complex, because it involves interactions with other processes, notably soil moisture. As such, the final effect of vegetation feedback depends not only on the vegetation itself, but also on other land surface properties and climate dynamics. In addition to the potentially complex interaction between vegetation and soil (e.g. Zeng *et al*., 2005), atmospheric dynamics is potentially important for nonlocal feedback. Thus, in spite of our general understanding of some individual vegetation feedbacks on climate, much work is needed to better understand the complete vegetation feedback. Indeed, the complexity of vegetation feedback may help explaining some uncertainties of vegetation feedbacks in the observations and models. For example, in North Africa, an assessment of vegetation feedback on precipitation from recent remote sensing observations shows subtle signals without a dominant sign (Liu *et al*., 2006). State-of-the-art coupled climate–vegetation models also show a large disparity, some exhibiting positive feedback, while others negative feedbacks (Claussen *et al*., 2004; Braconnot *et al*., 2007). These discrepancies may be associated with different relative contributions of direct and indirect vegetation feedbacks. Clearly, more systematical studies are needed in other models to understand the vegetation feedback.

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